# EFFECTS OF HYDROTHERMAL UNREST ON STRESS AND DEFORMATION: INSIGHTS FROM NUMERICAL MODELING AND APPLICATION TO VULCANO ISLAND (ITALY)

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## Abstract

A numerical approach is proposed to evaluate stress and deformation fields induced by hydrothermal fluid circulation and its influence on volcano-flank stability. The numerical computations have been focused on a conceptual model of Vulcano Island, where geophysical, geochemical and seismic signals have experienced several episodes of remarkable changes likely linked to the hydrothermal activity. We design a range of numerical models of hydrothermal unrest and computed the associated deformation and stress field arising from rock-fluid interaction processes related to the thermo-poroelastic response of the medium. The effects of model parameters on deformation and flank stability are explored considering different multilayered crustal structures constrained by seismic tomography and stratigraphy investigations. Our findings highlight the significant role of model parameters on the response of the hydrothermal system and, consequently, on the amplitudes and the timescale of stress and strain fields. Even if no claim is made that the model strictly applies to the crisis episodes at Vulcano, the numerical results are in general agreement with the pattern of monitoring observations, characterized by an enhancing of gas emission and seismic activity without significant ground deformation. The conceptual model points to a pressurization and heating of the shallow hydrothermal system (1 – 0.25 km bsl) fed by fluid of magmatic origin. However, for the assumed values of model material and source parameters (rate of injection, fluid composition and temperature) the pressure and temperature changes do not affect significantly the flank stability, which is mainly controlled by the gravitational force.

***Key words***: Hydrothermal circulation; Numerical modeling; Thermo-poro-elasticity; Volcano flank instability; Vulcano Island

## 1. Introduction

Active volcanoes grow and build up so rapidly that their edifices are inherently unstable and their flanks are usually displaced under the actions of different agents. While volcano instability has been recognized and documented at many volcanoes worldwide, the cause-and-effect relationships among the involved processes have so far been difficult to capture. The volcano dynamics plays an active role in the factors controlling volcano deformation and structural stability, and numerous processes such as magma intrusions (McGuire, 1996; Iverson, 1995; Elsworth and Voigth, 1996; Elsworth and Day, 1999; Voight and Elsworth, 1997), replenishment of fresh magma in reservoirs and seismic activity (Voight et al., 1981) can trigger volcanic flank failure. Among the others, also hydrothermal fluid circulations due to thermal expansion and pore pressure acting on rocks may significantly alter the stress state of the volcanic edifice (Rinaldi et al., 2010; Bonafede, 1990, 1991; De Natale et al., 1991, 2001; Hurwitz et al., 2007; Hutnak et al., 2009; Orsi et al., 1999; Hayba and Ingebritsen, 1994) and hydrothermal alteration often plays a major part in increasing susceptibility to failure. However, up to date few studies have addressed the quantification of expected deformation and stress variations caused by hydrothermal fluids during a generic unrest period (Reid, 2004). To afford this topic, we implemented a hydro-mechanical model to evaluate stress, strain and deformation fields caused by hydrothermal fluid circulation.

The hydro-mechanical model is implemented by coupling a thermo-poroelastic numerical code, developed under COMSOL software (Comsol, 2012), with TOUGH2, a commercial software simulating multi-phase and multi-component fluid flow and heat transfer. Based on thermo-poroelasticity theory and the definition of a failure criterion, stress and strain fields are evaluated to define the regions of the volcano edifice more likely to displace and fail. Numerical results show the contribution of hydrothermal fluid flow circulation associated with induced thermoelastic and pore-pressure changes, providing a quantitative estimate for deformation and failure of a volcano edifice.

The model is applied to the case-study of Vulcano Island, an active volcanic complex which is potentially affected by significant geohazards related to the activity of the magmatic and hydrothermal systems. In 1988, indeed, fracturing and increase of the hydrothermal activity resulted in an enhanced slope instability and caused part of the northeastern sector, a volume of about $2\times10^5$ $m^3$, to slide into the sea generating a small tsunami (Achilli et al., 1998; Tinti et al., 1999). A correlation between the gravitational instability of the slope and the increased volcanic activity was suggested as the direct cause of the slide (Tinti et al., 1999). However, other processes such as water-rock interaction and repetition of inflation–deflation cycles, which could lower the rock shear strength of the volcanic edifice, were not ruled out (Rasà and Villari, 1991). In recent times, no magmatic eruptions have taken place at the island, but recurrent thermal and seismic crises, attributed to magma–water interaction (Federico et al., 2010; Alparone et al., 2010), each lasting no more than a few months, have occurred. These crises are accompanied by sudden and intense changes in the set of geophysical and geochemical monitored parameters. These evidences are a signature of the significant interplay between rock and fluid circulation within the hydrothermal system. Therefore, the estimate of the rock-fluid interaction processes and their influences on volcano-flank stability are of primary importance for the Vulcano Island hazard assessment. Here, we investigate the role played by fluid injection, composition, and medium rheology in controlling the internal stress state of the volcano, whose amplitudes and distributions outline the volcano edifice regions that are more likely to fail. As most parts of the system are inaccessible to direct observations, the exploration of different scenarios by means of numerical simulations will help in understanding and characterizing the hydrothermal system activity and in interpreting the associated geophysical observations.

## 2. Geological setting

Vulcano is the southernmost island of a NW-SE elongated submarine volcanic ridge which rises more than 1 km from the continental slope (Romagnoli et al., 2013). The ridge develops along two main systems of NW–SE trending right-lateral strike-slip faults, parallel to the NW-SE Tindari-Letojanni system, which extends up to NE Sicily (De Astis et al., 2013; De Ritis et al., 2005). The Vulcano structural pattern is generally dominated by a NNW–SSE trend, representing a surficial expression of the Tindari–Letojanni system, and by minor N-S to NE-SW trending normal faults and cracks associated to the main NW–SE shear zone (Ventura et al., 1999). The subaerial morphology of the island is characterized by: (i) two main overlapping structural depressions more than 2.5 km wide, named the Piano caldera and La Fossa caldera; (ii) La Fossa cone, a 390 m high composite edifice, located within the central sector of La Fossa caldera, and (iii) the pyroclastic edifice of Vulcanello, which with its lava platform form a roughly circular peninsula on the northern tip of the island (Chiarabba et al., 2004; Revil et al., 2008).

Since the 70s, several geophysical studies were executed to detect and define the shallow structures of the volcano complex (Iacobucci, 1977; Barberi et al., 1994; Blanco-Montenegro et al., 2007; Napoli and Currenti, 2016). Moreover, valuable information on the subsurface structure of the La Fossa caldera were obtained, by two deep geothermal drillings (Fig.1) in 1983–1987 (Giocanda and Sbrana, 1991), located at the foot of the south-western and northern flanks of La Fossa cone. The former encountered a shallow monzodioritic intrusion at about 1350 m bsl, and reached a vertical depth of 2050 m where a temperature of 419 C° was founded. The second well found a temperature of 243°C at 1338 m and between 350 and 400°C at 1578 m (Faraone et al. 1986).

Since 1890, when the last eruption of La Fossa cone occurred, the volcano activity has been characterized by recurrent thermal and seismic crises due to magma-water interaction (Federico et al., 2010; Alparone et al., 2010). Gravity, seismological and geochemical studies (Berrino 2000; Chiodini et al., 1992; Alparone et al., 2010) detected an active hydrothermal system beneath La Fossa caldera, between 500 and 1500 m bsl, whose activity is represented by the intense fumarolic emissions in the summit area. In particular, a few high temperature (400 °C) zones are active (De Astis et al., 2003) within the crater, while on the southern and northern flanks of the edifice, temperatures generally do not exceed 98°C (Barde-Cabusson et al., 2009). Fumaroles temperature and the superficial manifestations strongly increase when input of fluids of magmatic origin occurs even without evidence of magmatic rising, as happened in 2004-2006. In these cases the anomalous degassing episodes could derive from changes in rock permeability (Todesco, 1997) or reflect a pulsating degassing process from a deep pressurized stationary magma body (Granieri et al., 2006).

## 3. Hydro-mechanical model

The hydro-mechanical model is based on the governing equations of the thermo-poro-elasticity theory, which describes the response of a porous medium to the propagation of hot fluid. A one-way coupling model is here considered in which the pore pressure and temperature changes influence the elastic stresses, but not vice-versa. Although it is a limitation of the model, this assumption is not so restrictive since uncoupled and coupled pore pressure solutions are quite close for many ranges of medium properties (Roeloffs, 1988).

*Fluid flow model*

The hot fluid circulation in the hydrothermal system is simulated using the EOS2 module of TOUGH2 software (Pruess et al., 1999), which incorporates $CO_2$-$H_2O$ equations of state in the temperature and pressure range 0–350 °C and 0–100 MPa, respectively. It solves the mass and energy balance equations for a multiphase ground-water flow (Pruess et al., 1999). The mass balance equations can be resumed as follows:

$$\frac{\partial Q^m}{\partial t} q \ \nabla \cdot \boldsymbol{F}^m - q^m = 0 \qquad (1)$$

where . is the accumulation term, $\boldsymbol{F}$ the flux and $q$ the source (or sink) term and $m$ the mass component (water or $CO_2$). A full list of the symbols with their unit of measurements is provided in Table 1. The accumulation term for mass balance equation is described by $.^m = 0 \sum_2 1_2 3_2 \chi_2^m$, where the subscript $\gamma$ refers to the liquid $\boldsymbol{F}lm$ or gas $\boldsymbol{F}gm$ phase, respectively, $0$ is the porosity, $1_2$ the density, $3_2$ the saturation and $\chi_2^m$ the mass fraction of component $m$ present in phase $\gamma$. The fluid flux $\boldsymbol{F}^m = \sum_2 \chi_2^m \boldsymbol{F}_2$ is described by the Darcy's law extended to two-phase conditions with separate equations for the gas and liquid phases:

$$\boldsymbol{F}_2 = \mathbf{v}_2 1_2 = \frac{k\, k_{r\gamma}\, \rho_\gamma}{\mu_\gamma} \left( \nabla P_2 - 1_2\, \widehat{\boldsymbol{g}} \right) \qquad (2)$$

where $\mathbf{v}_2$ is the Darcy's velocity, $k$ and $k_{r2}$ are the absolute and relative permeability to phase $\gamma$, respectively, $\mu_2$ the viscosity, $P_2$ the fluid pressure, and $\widehat{\boldsymbol{g}}$ the gravitational acceleration vector.

The energy balance equation is represented by (1) as well (with the subscript $m = E$ standing for energy), where the accumulation and the flux terms are, respectively, $.^E = 0 \sum_2 1_2 e_2 3_2 \ q \ (1-0) 1_R C_R T$ and $\boldsymbol{F}^E = -\lambda_r \nabla T \ q \sum_2 h_2 \boldsymbol{F}_2$, where $e_2$ is the specific internal energy of the phase $\gamma$, $1_R$ and $C_R$ are the density and the specific heat of the rock, respectively, $T$ is the temperature, $\lambda_r$ the thermal conductivity of the rock, and $h_2$ the specific enthalpy of the phase $\gamma$.

Each phase may be at a different pressure $P_2$ due to interfacial curvature and capillary forces. The difference between the gas and liquid pressures is referred as capillary pressure $P_c$. In order to close the equation system, relationships for the capillary pressure and relative permeabilities are needed. These relationships are usually posed as a function of the liquid saturation $3_l$, on the basis of experimental data. Therefore, the fluid flow process is also controlled by capillary pressure - saturation – relative permeability relationships. TOUGH2 allows for investigating several hydrological models. One of the most common formulations used in hydrological models to describe these relationships is the Brooks-Corey function (Brooks and Corey, 1964) based on experimental observations and defined as:

$$P_c = \frac{P_b}{S_e^{1/\delta}}, \quad \text{with} \quad 3_e = \frac{S_l - S_{lr}}{1 - S_{lr}} \tag{3}$$

where $3_e$ denotes the effective liquid saturation, $3_{lr}$ the residual liquid saturation, $P_b$ the bubbling pressure and $\delta$ the pore size distribution index. The bubbling pressure, which is also called the displacement pressure, is the extrapolated capillary pressure at full liquid saturation. Brooks and Corey exploited the Burdine theory to derive the relative permeability-saturation relationships for gas and liquid phases:

$$k_{rl} = 3_e^{\frac{2+3\delta}{\delta}}$$

$$k_{rg} = (1 - 3_e)^2 \boldsymbol{F}1 - 3_e^{\frac{2+\delta}{\delta}} m \tag{4}$$

The parameters for the Brooks-Corey two-phase characteristic curves are fixed to average values of $3_{lr} = 0.3$, $\delta = 2$ and $P_b = 5000$ Pa.

*Elasto-mechanical model*

Assuming that the timescale of deformation is slow enough to allow for pressure equilibration, the rock is in quasi-static equilibrium and the displacement can be found by solving the stress equilibrium equations coupled with thermo-poroelastic extension of the Hooke's law (Jaeger and Cook, 2007; Fung, 1965), giving the following set of equations:

$$
\begin{aligned}
&\nabla \cdot \boldsymbol{\sigma} = \mathbf{H} \\
&\boldsymbol{\sigma} = \lambda tr(\boldsymbol{\varepsilon})\mathbf{I} + 2G\boldsymbol{\varepsilon} + \alpha_T K \Delta T \mathbf{I} + \beta \Delta P \mathbf{I} \\
&\boldsymbol{\varepsilon} = \frac{1}{2}\left(\nabla \mathbf{u} + (\nabla \mathbf{u})^T\right)
\end{aligned}
\tag{5}
$$

where **σ** and **ε** are the stress and strain tensors, respectively, *H* is the body force, **u** is the deformation vector and λ and *G* are the Lame's elastic medium parameters, related to the Poisson ratio and Young's modulus. To the elastic stress tensor of the general Hooke's law, two terms are added: (i) the ΔP pore-pressure contribution from poroelasticity theory through the $\beta = (1\text{-}K/K_s)$ Biot-Willis coefficient and (ii) the ΔT temperature contribution from thermo-elasticity theory through the volumetric thermal expansion coefficient $\alpha_T$.

Within this framework the stress field inside the volcanic edifice originates from two main contributions (Iverson and Reid, 1992; Reid, 2004; Marti and Geyer, 2009; Zang and Stephansson, 2010; Muller et al., 2001): (1) the background stress composed of the gravitational loading and (2) the stress field generated by the pore pressure and thermo-elastic effect. Gravitational loading is included in the model by imposing on each element an internal body force per unit volume $\boldsymbol{H} = -1_R\, \hat{\boldsymbol{g}}\, z$, where $1_R$ is the density of the host rock, $\hat{\boldsymbol{g}}$ is the gravitational acceleration vector and $z$ the elevation. The gravitational body force is included in the simulations to obtain an overburden stress of rock at any given depth in the medium. The volcanic edifice itself, acting as a load on the upper crust, generates a stress regime (Liu and Zoback, 1992; Pan et al., 1995; Pinel and Jaupart, 2004; Currenti and Williams, 2014) that affects the flank stability (Reid, 2004). The mathematical problem is closed by imposing zero displacements at infinity and stress-free boundary condition $\boldsymbol{\sigma} \cdot \mathbf{n_s} = 0$ on the ground surface, where $\mathbf{n_s}$ is the normal vector to the ground surface. The problem is solved by finite element method using COMSOL Multiphysics (COMSOL, 2012). The pore-pressure and temperature contributions are fed from the outputs of the TOUGH2 fluid flow model solutions by implementing a MATLAB script to automate the COMSOL computations at each time steps.

*Failure criteria*

Assuming that the subsurface comprises poroelastic media, the mechanical response is governed by its effective stress distribution, which is the total stress modified by fluid pressure as follows:

$$\boldsymbol{\sigma}^{eff} = \boldsymbol{\sigma} - \beta P\mathbf{I} \qquad (6)$$

where $\boldsymbol{\sigma}$ is the total applied stress on the rock-fluid mixture and $P$ is the pore fluid pressure (Ranalli, 1995). The pore pressure acts against the total stress, effectively reducing the resistance to failure. Several yield criteria have been investigated to model the mechanical behavior of rocks undergoing failure (Fung, 1965; Ranalli, 1995; Jaeger et al., 2007). Such failure depends on both the deviatoric stress and the overburden stress through the friction coefficient. The Drucker-Prager failure law can be properly used for modeling brittle behavior in the upper crust (Cattin et al., 2005; Cianetti et al., 2012; Apuani et al., 2013; Got et al., 2013). The yield function for Drucker-Prager failure in the case of no hardening (perfect plasticity) may be written as:

$$Y_f = -\eta_1\, q\ \eta_2 I_1\, q\ \sqrt{\mathrm{t}_2} \qquad (7)$$

where $I_1$ is the first invariant of the effective stress, $J_2$ the second invariant of the deviatoric stress tensor, $\eta_1$ and $\eta_2$ medium coefficients. Generally, when the effective stress state satisfies the yield criterion $Y_f \geq 0$, the material will undergo failure. The Drucker-Prager criterion represents a smoothed version of the Mohr-Coulomb frictional failure criterion for a three-dimensional case. Indeed, the coefficients $\eta_1$ and $\eta_2$ may be related to the Mohr-Coulomb frictional failure properties cohesion ($c$) and friction angle ($\theta$), derived from laboratory experiments (Jaeger et al., 2007; Chen, 1982):

$$\eta_1 = \frac{\text{wc} \cos\theta}{\sqrt{3}F3 - \sin\theta m} \qquad \eta_2 = \frac{2\sin\theta}{\sqrt{3}F3 - \sin\theta m} \qquad (8)$$

The Drucker-Prager criterion, such as the Mohr-Coulomb one, accounts for the experimental observations that yield stress of most rocks increases with increasing mean normal stress (Jaeger et al., 2007; Mazzini et al., 2009; Liu et al., 2004). The volume, where the condition in Eq. 7 is satisfied, delineates shear-failure potential at any point within the volcano edifice and thus assists in pinpointing locations that have an increasing exposure to flank failure.

## 4. Numerical Simulations

In order to investigate the effect of temperature and pore pressure changes on flank instability at Vulcano Island, it is necessary to outline a reasonable thermal and mechanical regime existing in the shallowest crust of the volcanic edifice, in agreement with geochemical and geophysical evidences. Firstly, TOUGH2 is run to evaluate pressure and temperature variations with respect to their initial distributions, which, then, are fed into the thermo-poroelastic solver to compute the deformation and stress fields. Without losing of generality, the model is designed in an axi-symmetric formulation in a computational domain of 10 x 1.5 $km^2$ which describes the shallowest portion of the hydrothermal system (Fig. 2). We included the real topography of Vulcano using a digital elevation model from the 90 m Shuttle Radar Topography Mission (SRTM) data and a bathymetry model from the GEBCO database (http://www.gebco.net/). The profile runs from the center of La Fossa cone toward North-East in order to describe the average slope of the Vulcano Island in the area affected by the 1988 landslide (Fig. 1). The domain was discretized in the radial direction by a set of logarithmically spaced nodes with a horizontal resolution starting from 20 m along the axis of symmetry and decreasing to 470 m at the external boundary. Vertically, the domain was divided into 90 equally spaced layers, which corresponds to a resolution of 20 m/layer. This discretization leads to about 10000 grid cells. In the thermo-poroelastic model the computational domain is bounded by infinite mapped elements and meshed into 152859 isoparametric and arbitrarily distorted triangular elements connected by 26090 nodes. The infinite mapped elements use appropriate transformation functions to map the finite domain into an infinite one and, hence, to make the displacements and stress fields vanish toward infinity (Zienkiewicz et al., 1983). Due to the different grids adopted in the fluid-flow and thermo-poroelastic solvers, pore pressure and temperature solutions are interpolated from the TOUGH2 grid to the thermo-poroelastic solver nodes (Fig. 2).

*Fluid flow parameter*

Fluid flow simulations require the definition of boundary conditions, hydrological medium properties and initial conditions.

As for boundary conditions, primary variables of TOUGH2/EOS2 (pressure, temperature and $CO_2$ partial pressure; Pruess et al., 1999) are fixed at the top surface for the entire simulation. In particular, atmospheric pressure is fixed along the subaerial boundary and hydrostatic pressure is assigned to the submarine boundary. Atmospheric values of temperature and $CO_2$ partial pressure are also assigned (Todesco et al., 2003). In the faraway side boundaries hydrostatic pressure and geothermal gradient (130 °C/km; AGIP, 1987) are assigned, whereas the bottom boundary has a constant basal heat flux of 60 mW/m$^2$ to represent heat flow from depth (Okubo and Kanda, 2010) and is impermeable except at locations where fluids are being injected.

The hydrological properties of the medium strongly control the response of the hydrothermal system to the applied sources of fluids and boundary conditions and, hence, affect the amplitude and the temporal evolution of the physical variables (pore pressure, temperature, gas saturation, stress-strain field and deformation). This is a relevant aspect, as hydrological properties may change significantly with rock type and chemical-physical condition and, consequently, a careful definition of in situ rock properties would be necessary. Unluckily, at Vulcano Island, site-specific values are poorly known. In order to define the hydrological model properties, we exploited the stratigraphy inferred from logs of the deep VP1 borehole (Fig. 1), which allows to identify the succession of different layers from top to bottom (Gioncada and Sbrana, 1991; Tommasi et al., 2016). We grouped the lithology units into two main classes distinguished, on average, by characteristic high (hyaloclastites, pyroclastites, scories) and low (latitic lava flows, trachytic intrusion) permeabilities. The complex structure of the volcanic edifice is, therefore, simplified by a stratified model (A) consisting of two alternating layers with low (Layer 1) and high (Layer 2) porosity and permeability values (Fig. 3). In order to account for alteration along and across the main flow paths of hydrothermal fluids, a second model (B) is also investigated. It includes a 250 m wide inner zone, simulating the central pathway, with high porosity and permeability, and transition zones between the central pathway and the remaining domain, which extend for 250 m and have intermediate hydrological parameters (Table 2; Fig. 3). Porosity values, determined from laboratory test on rock samples from a 100 m deep borehole BL1 located near the VP1 borehole (Fig. 1), are scattered in the range between 0.1 and 0.4 (Tommasi et al., 2016). In the lack of in-situ measurements, the values of permeability are set up on the basis of estimates in other similar volcanic environments (Todesco et al. 2010, Okubo and Kanda, 2011; Rinaldi et al., 2011; Troiano et al., 2011; Coco et al., 2016).

The initial conditions of the numerical simulations are designed on the basis of a conceptual model derived from geochemical investigations, which hypothesizes the presence of a hydrothermal system at a depth of 0.5 – 1.5 km bsl with equilibrium temperature up to 400°C (Federico et al., 2010). It is fed by fluids of magmatic origin, and unrest events are ascribed to periods of increased fluid injections. Fluxes of chemical components have been estimated from compositional analysis of the discharged fumaroles. Carbon dioxide represents the main constituent of anhydrous gases discharged from the summit areas through plumes and crater fumaroles (Inguaggiato et al., 2012). Over the past 15 years of observations, the $CO_2/H_2O$ weight ratio has varied from 0.05 to 0.5, with an average value of approximately 0.23. Such large variations were interpreted as due to variable mixing of the magmatic fluids, rich in $CO_2$, with a shallower hydrothermal component (Chiodini et al. 1992; 1996). A water flux of about 1300 t/day and a total $CO_2$ output of 482 t/day have been estimated for the whole area of Vulcano Island considering discharged fluids from crater fumaroles, soil degassing over the island, and bubbling and dissolved gas (Inguaggiato et al., 2012). These estimates refer to periods of moderate hydrothermal activity. On the basis of these evidences, the model is run for thousands of years to reproduce a fluid state (pressure, temperature and gas saturation), which resembles the conceptual model of the hydrothermal system. Quasi-steady state conditions are reached by simulating a continuous injection of 1300 t/day of pure water and 450 t/day of $CO_2$ at a temperature of 350 °C for both model A and B. The fluid is injected in a 250 m wide inlet located at the bottom of the domain near the symmetry axis. The pressure, temperature and gas saturation distributions are shown in Figs. 4a-c. In the model A, after reaching a quasi-steady state solution, a high-temperature deep zone develops above the fluid injection area extending horizontally up to 0.5 km of distance from the symmetry axis (Fig. 4a). In such a volume, the rock temperature reaches more than 300 °C, and the highest value of gas saturation is reached (Fig. 4c). A gas-saturated zone is observed at the inlet, and saturation decreases in the shallow layers. Temperatures are distributed in a larger area up to the ground surface where values greater than 100 °C still appear. Pressure mainly follows the hydrostatic condition with perturbation concentrated within 500 m (Fig. 4b). Also in the model B (Figs. 4d-f), a high-temperature deep zone develops above the fluid injection area but it affects a smaller volume extending horizontally up to 0.25 km of distance from the symmetry axis. In such a volume, the temperature reaches more than 300°C, and even if it gradually decreases in the shallow layers, values greater than 100 °C still appear just below the ground surface (Fig. 4d). Remarkable differences are observed in the temperature and gas saturation distributions with respect to model A. In model B the higher permeability in the inner and transition zones favours the flow within these regions, which, consequently, are more heated. A gas-saturated area develops just below the ground surface and spreads up horizontally in the shallowest permeable layer (Fig. 4f).

*Mechanical parameters*

A heterogeneous medium for the subsurface structure of Vulcano Island is considered using a piecewise linear depth dependent distribution of the elastic material properties derived from seismic tomography (Chiarabba et al., 2004; Ventura et al., 1999). Within the computational domain the P-wave seismic velocity $V_p$ varies from 2 km/s to 4 km/s and the rock density $\rho$ ranges between 2100 kg/m$^3$ to 2400 kg/m$^3$ (Table 3). Low values of $V_p$ and $\rho$, related to the pyroclastics, hyaloclastites and hydrothermally altered rocks, were assigned to the shallow part of the volcanic edifice, up to 1 km of depth. Because of the axis-symmetric formulation, horizontal heterogeneities and local effects of high velocity bodies ($V_p$ greater than 3 km/s) were disregarded. With increasing depth, higher values of seismic velocity and rock density related to intrusive or sub-intrusive bodies, as well as to crystallized conduits system, were assigned to the medium (Chiarabba et al., 2004). These $V_p$ and $\rho$ values were used to define the elastic Young modulus by the following equation (Kearey and Brooks, 1991):

$$\mathrm{E} = \mathrm{V}_\mathrm{p}^2 \rho_R \frac{(1-2\nu)(1+\nu)}{1-\nu} \tag{9}$$

An average value of 0.3 for the Poisson's ratio $\nu$ was used. Under these assumptions, within the computational domain the Young modulus increases from 9.0 GPa in the shallow layer to 23.3 GPa at the bottom of the domain.

On the basis of literature data (Coco et al., 2016; Rinaldi et al., 2010; Todesco et al.,2010; Troiano et al., 2011), we chose average values for the volumetric thermal expansion parameter $\alpha_T$, fixed to $10^{-5}$ °$C^{-1}$, and for the drained bulk modulus, set to 5 GPa.

Estimates of the mechanical characteristics of rock masses are also required to evaluate the Drucker-Prager failure criterion. Generally, in a geothermal volcanic area such as Vulcano Island, built up by a stack of pyroclastics, tuffs, hyaloclastites and hydrothermally altered clasts (De Astis et al., 2013), low friction angle and cohesion are expected. Mechanical parameters of Vulcano rocks have rarely been measured. Recently, a geotechnical characterization of cores collected from the BL1 borehole and of rock samples from Punte Nere deposits was conducted to estimate their mechanical properties (Tommasi et al., 2016). The laboratory tests have evidenced that hydrothermal alteration around fumaroles has changed mineralogy, texture and mechanical behavior of the material. In particular, a reduction of friction angle down to 26° or 21° was reported. Based on the results of laboratory test, cohesion values from 0.5 to 1 MPa were investigated for the parameters of the ideal elastoplastic rheological model.

## 5. Results

Since the last eruption occurred during 1888–1890, the active volcanic center of La Fossa cone displays fumarole activity, characterized by periodic phases of increased output flux and temperatures of emitted fluids (Alparone et al., 2010; Federico et al., 2010). Enhanced fluid discharge from the crater fumaroles is supposed to be sustained by an enhanced deep fluid injection, as experimentally supported by higher $CO_2/H_2O$ ratios measured in high-temperature fumaroles during unrest (Granieri et al., 2006; Paonita et al., 2013). In order to evaluate the effect of enhanced fluid injection in the volcano-hydrothermal system on the deformation and stress fields, we simulated an unrest phase of 1 year by increasing (after reaching the steady-state solution) the flux rate to 2000 t/day for the water and to 1000 t/day for the carbon dioxide content, in agreement with geochemical data collected at Vulcano Island during unrest phase (Chiodini et al., 1996; Granieri et al., 2006). The distributions of saturation and pressure and temperature changes at the end of the unrest, with respect to the steady-state initial conditions, are displayed in Fig. 5. In the model A, temperature and pore pressure changes do not propagate farther than 1 km from the source region (Fig. 5a, b). The pressure changes reach 2 MPa in correspondence of the fluid inlet, while the maximum temperature changes of about 10°C are concentrated at its edge. Only slight temperature changes within 0.2 °C are observed near the ground surface. A gas saturated area is concentrated at the fluid inlet and the saturation value suddenly drops below 0.5 at a depth of about 1.25 km bsl extending up to the ground surface (Fig. 5c) with a similar pattern to that produced in the steady-state conditions (Fig. 4c). In the model B, the fluid easily propagates upward and the flow is mainly confined in the inner and transition zones where the rock permeability is higher. The heated region extends up to the sea level showing the maximum values at the depth of about 1 km bsl in correspondence of the edge of the inlet area (Fig. 5d). The main pressure variations of about 0.5 MPa are concentrated within the inner region between 1 km and 0.25 km bsl (Fig. 5e). The gas saturation distribution preserves the pattern produced in the steady-state conditions, with a gas-saturated area just below the ground surface, and another area with high values of saturation grows in the deeper part near the symmetry axis between 1.5 km and 1 km bsl (Fig. 5f).

*Ground displacements*

Using the pressure and temperature solutions achieved from the fluid flow simulations, the evolution of ground deformation is then evaluated using the thermo-poroelastic solver. In the model A, after 1 year of continuous injection the radial distribution of horizontal deformation reaches a maximum amplitude of about 0.4 cm at 1.25 km away from the symmetry axis, where the volcano edifice submerges. Concurrently, the vertical uplift attains about 0.5 cm from the origin (r = 0) to 1.25 km and diminishes to 0.1 cm within 2.5 km (Fig. 6). Remarkable differences are observed in the model B, where the radial distribution of horizontal deformation shows the maximum amplitude value of about 0.8 cm from 0.5 km to 1.5 km and then decreases within 4.5 km. Simultaneously, the vertical uplift attains about 2.2 cm at the origin and vanishes to 0.1 cm within 2.5 km (Fig.6). The discrepancies between model A and B are attributable to the different distributions of pressure and temperature changes (Fig. 5). In model A these changes affect a wider area (Fig.5 a,b) and, in turn, reflect in a more extensive deformation pattern. Conversely, in model B the temperature and pressure changes (Fig. 5d,e) are more confined in the inner zone because of the higher permeability of this region and, hence, they engender a narrower deformation pattern. The evolution of the vertical ground displacement arising from the simulated unrest was calculated at the ground surface at the origin point (Fig. 7). The vertical deformation in the summit area evolves almost linearly in time for the model A. Uplift begins as soon as the injection rate is increased and reaches the maximum amplitude of about 0.5 cm at the end of the unrest. In the model B, the evolution of the vertical uplift follows a non-linear trend. After an initial constant linear increase, the rate of deformation decays from 0.3 to 0.6 years, when it starts to rise again. The rise of deformation rate after 0.6 years in model B is due to the concurrent onset of the upward migration of the pore pressure changes, which primarily contribute to the deformation with respect to thermo-elastic effect. Dissimilar to model A, where the pressure changes remain confined at depth in proximity of the inlet (Fig. 5b), in model B, the pressure change front proceeds toward the surface (Fig. 5e) engendering to an increase in the ground deformation rate. The maximum uplift and the higher deformation rate are, hence, attained for the model B, which reflects the ability of the hydrothermal system to respond faster to the injection of fluids due to higher permeability in the inner and transition zones.

For the model B, we have also investigated the effects on deformation generated by an increase in the rock permeability in the inner zone. Indeed, an increase in the gas output may be not only the consequence of an enhanced gas input but it may also reflect a temporal permeability increase in the gas pathway by rock fracturing. As the permeability value in the inner zone is doubled, the hydrothermal system is decompressed and subsidence in ground deformation is observed (Fig. 6). A further simulation is performed in which both the gas flux rate at the inlet and the permeability in the inner zone are increased. This simulation allows to investigate the effect of a potential positive feedback among enhanced gas input, pressurization of the hydrothermal system, rock fracturing and increase of permeability. In such a case, an uplift is still observed even if the amplitude is lower. Moreover, the deformation field is narrower and rapidly decays with distance. Indeed, the cause-effect relationship governing this feedback process may be more complex. Gas input may result in pressure increase, that leads to hydro-fracturing and permeability increase, that on its own results in fluid release, pressure decrease and closure of fracture. The simulation of this coupled mechanism could be more suitably described assuming a pressure-dependent permeability model (Rutqvist et al., 2002).

*Failure Surfaces*

To investigate failure conditions induced by the hydrothermal activity of Vulcano Island during the simulated unrest, a stress-strain analysis is performed. The volcanic edifice itself acts as a load on the upper crust and generates shear stress components that greatly affect the volcano flank stability (Currenti and Williams, 2014). Therefore, in addition to the introduction of the pore pressure and thermo-elastic stress, the model is subjected also to the gravity body force to become fully compressed under its own weight (Reid, 2004). The stress state induced by the topography loading is computed through the activation of gravitational body forces (Bonaccorso et al., 2010; Cianetti et al., 2012; Currenti and Williams, 2014). This procedure allows to reach an equilibrium state in the presence of gravity loading and provides the stress distribution induced by the topography. A gradient in the stress component $\sigma_r$ is achieved at the change of the topography curvature (Fig. 8a), whereas $\sigma_z$ almost follows the topography shape (Fig. 8b). A shallow local concentration in $\sigma_{rz}$ is observed where an abrupt change of slope occurs at about 700 m from the origin (Fig. 8c). The stress components generated by the pore pressure and thermo-elastic contributions are used to compute the failure surfaces expected at Vulcano Island at the end of the simulated unrest phase. On the basis of the wide variabilities of mechanical parameters of Vulcano rock samples (Tommasi et al., 2016), the yield failure function is computed for two different values of friction angles: 20° and 30° (Figs. 9-10). The contour lines refer to the failure surfaces, which are defined using Eq. 7 and computed for different values of the $\eta_1$ parameter obtained for two values of the cohesion coefficient c = 0.5 MPa (white line) and c = 1 MPa (black line). Failure regions are located in the La Fossa cone and in the steepest slope of the edifice. Both aerial and submarine failure surfaces develop along the flank of the volcanic edifice (Figs. 9,10). The failure condition is strongly dependent on the friction angle, whose reduction promotes the rocks to undergo failure. The extension of the failure region is controlled by the cohesion coefficient through the $\eta_1$ parameter. The lower the cohesion, the wider the area affected by failure. It is worth noting that for lower cohesion coefficient the failure zone also extends in the submarine portion of the volcanic complex. Similar results are obtained both for model A and B. Slight enhancement in the extension of the failure zone is observed for the model B due to the upward migration of the pore pressure changes. Indeed, since pressure and temperature changes are confined around the inlet region for model A and mainly in the inner zone for model B, the stability of the volcano flanks, under the assumed model parameters, is not significantly influenced by the hydrothermal fluid circulation. The failure surfaces slightly differ from that computed under the only effect of gravitational loading (Fig.11).

## 6. Discussion and Conclusions

Stress-strain numerical analysis in volcanic areas is an increasingly interesting research topic, which may help in driving inferences on the internal state of a volcano edifice (Coco et al., 2014; Currenti,2014; Currenti and Williams, 2014; Schopa et al., 2011; Cianetti et al., 2012). Particularly, in this paper we proposed a geomechanical approach to evaluate the deformation and flank instability in volcanic hydrothermal systems with application to Vulcano Island. Numerical results support that thermal heating and pore pressure due to an increment in the inflow of volcanogenic fluids may generate temporarily pressurization of pore fluids, which induce ground deformation increasing in time as the injection of fluid is sustained. The comparison between the results obtained from model A and B indicates that the amplitudes, distribution and temporal evolution of the physical variables are strongly dependent on the model material assumptions. By investigating the numerical results, the model B seems to represent the more likely scenario since it is able to reproduce geophysical phenomena observed at Vulcano Island. In particular, in model B, the gas saturation distribution (Fig. 4f) covers a wide zone corresponding to the areas where diffuse emissions from the soil have been recorded all around the base of the volcanic La Fossa cone (Diliberto et al., 2002). With the present choice of initial conditions, injection rate and rock properties, based on a sound hydrothermal conceptual model, both horizontal and vertical deformation during unrest phase are within few centimeters. These findings are in agreement with geodetic observations from tilt (Cannata et al., 2011) and DInSAR measurements (Azzaro et al., 2013), which show no significant ground deformation during the most recent recorded anomalies in temperature, chemical composition and seismicity recorded during 2004-2006 (Alparone et al., 2010; Granieri et al., 2006). Continuous monitoring reveals that strong increases in fumaroles temperature and in superficial manifestations, with a remarkable enlargement of the exhalative area and a progressive increase in the $CO_2$ emission rate, are generally observed when input of fluids of magmatic origin occurs (Granieri et al., 2006; Inguaggiato et al., 2012). The recurrence of these degassing events could be related to the progressive accumulation of volatile at the top of an accumulation zone, followed by a volatile release affecting the hydrothermal fluid budget and the pressurization in the surrounding media. Such anomalous degassing periods are accompanied by increases in the number and amplitude of volcano-seismic events at shallow depth (<1-1.5 km) under La Fossa cone (Alparone et al., 2010). Model B results highlight a pore pressure increase in a 0.5 km wide zone above the hydrothermal system at depths ranging between 1 and 0.25 km bsl, which may induce fracturing in the same area where micro-seismicity has been generally recorded. Supported by the lack of volcano-tectonic seismicity and significant deformation, our results agree with the hypothesis that micro-seismicity is likely related to pore pressure increase induced by the release of fluids from a deep magma zone rather to magma migration into the shallow hydrothermal system (Cannata et al., 2011).

On the basis of the solved elastic effective stress field, the distributions of failure surface estimated using the Drucker-Prager criterion pinpoint locations that have an increasing exposure to flank failure. The numerical results evidence the strong dependence of the failure surface by the friction angle and cohesion. In the history no evidences for a large failure slope such as the one obtained for a friction angle of 20° and a cohesion of 0.5 MPa (Figs. 9-10) have been reported. The failure surfaces achieved for higher cohesion (1 MPa) values are in general agreement with a landslide height evaluated by Achilli et al. (1998) for the 1988 event. The high sensitivity of the results on mechanical coefficients rises the need to conduct a more detailed characterization of rock in situ properties to better define the failure criterion parameters. Moreover, because of simplistic assumption on the mechanical rock properties due to the lack of measurements of in situ rock properties, the actual geometry of the failure surface could be likely more complex than the failure surfaces modeled here. However, our analysis provides for the first time an initial quantitative flank stability assessment induced by hydrothermal processes. For the investigated scenarios, the pore pressure and temperature change do not seem to affect significantly the edifice stability, which is mainly controlled by gravitational loading (Fig. 11) in agreement also with the results reported recently in Tommasi et al. (2016). Indeed, under the model assumption, hydrothermal activity is confined within the higher permeability areas near the inner zone and does not cause significant stress perturbations along the volcano flank.

The findings of this study provide baseline information about the stability of the volcano edifice, and lead to a reliable method for assessing the hazard associated with crisis, or intensified activity, of the hydrothermal system fed by a source of hot fluids. A more detailed numerical modeling of the proposed process in terms of definition of fluid injection rates, model materials and parameters will benefit from a multidisciplinary approach that enables to clarify cause-and-effect relationships and to identify the critical controlling factors. Different distributions of rock properties may greatly affect the results. Particularly, hydrothermal alteration may have locally impacted permeability, porosity, thermal parameters, rock strength and mechanical rock properties. Hydrothermally altered rocks seem to cover large zones in the northern and southern flanks of La Fossa cone edifice, as evidenced by the distribution of reduced magnetization areas inferred from 3D magnetic model (Napoli and Currenti, 2016). The wide extension of this low magnetization regions is a fingerprint of a more diffuse historical hydrothermal activity than in present days, which may have drastically altered over time the hydrological and mechanical properties of the rocks. Local changes in rock permeability may significantly alter the migration pathway of hydrothermal fluids, which tend to easily flow upwards along highest permeability zones. Moreover, the presence of local fractures may also affect the shallow hydrothermal circulation influencing pressure and temperature changes and, consequently, deformation and stress fields. Additionally, reduction in the friction angle and in the cohesion depending on the grade of hydrothermal alteration may locally enhance the extent of the region undergoing failure.

As a first attempt to integrate these elements in a unified framework, it raises several issues that require further in depth study by experts in different fields to refine the model and validate it with observations during crises period at Vulcano Island. We are confident that the methodology presented here can contribute to improve the joint interpretation of the geophysical, geochemical and seismological data recorded at Vulcano and in similar volcanic environments. Consequently, this model-based approach integrated with the monitoring observations may provide new hints concerning fluid-rock interaction processes to allow for the specific characterization and assessment of the volcano hydrothermal system.

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**Table Captions**

**Table 1 –** List of symbols in SI units.

**Table 2 –** Hydrological properties assigned to the model material.

**Table 3 –** Mechanical elastic parameter

**Figure Captions**

**Figure 1 –** Simplified geological map of the Vulcano Island. Legend: 1) alluvium and beach deposits; 2) Vulcanello pyroclastics: 3) Vulcanello lava flows; 4) Fossa cone pyroclastics; 5) Fossa cone lava flows; 6) Lentia domes and lava flows; 7) hyaloclastites and pillow lava; 8) lava flows and minor pyroclastics; 9) South Vulcano lavas and scorias; 10) drilling location. The profile crossing the 1988 landslide (black line) to define the topography of the axi-symmetric model is also reported.

**Figure 2** - Representation of the fluid flow (top) and mechanical (bottom) computational domains. The model is axis symmetric. The mechanical domain is extended to use infinite elements (green area). The actual grid and mesh are much finer than resolution shown in the figure.

**Figure 3 –** A simplified scheme of the complex geological structure of the volcanic edifice for the Models A (top) and B (bottom) based on the stratigraphy of the VP1 deep borehole reported on the left (after Blanco-Montenegro et al., 2007). The model domain is composed of five different regions: Layer 1 (L1); Layer 2 (L2); Transition zone for L1 (T1); Transition zone for L2 (T2); Inner zone (IZ).

**Figure 4 –** Temperature, pressure and saturation distributions used as initial conditions. A quasi-steady state solution is achieved by simulating a thousand years of continuous injection of 1300 t/day of $H_2O$ and 450 t/day of $CO_2$ at a temperature of 350 °C for model A (top) and model B (bottom).

**Figure 5 –** Changes in temperature and pressure with respect to initial conditions after 1 year of unrest simulated increasing the flux rate to 2000 t/day for the water and to 1000 t/day for the carbon dioxide content for model A (top) and model B (bottom). The gas saturation distributions are also shown.

**Figure 6 –** Radial distributions of horizontal (top) and vertical (bottom) displacements for model A (red lines) and model B (blue lines) after 1 year of unrest simulated increasing the flux rate to 2000 t/day for the water and to 1000 t/day for the carbon dioxide content. For model B the displacements obtained for an increase of the permeability in the inner zone (black line) and increases both in flux rate and permeability (green line) are also shown.

**Figure 7 –** Temporal evolution of vertical ground displacement for model A (red lines) and model B (blue lines) at the ground surface in the origin point (r=0) during a 1-year of unrest.

**Figure 8 –** Stress components under gravitational topographic loading.

**Figure 9 –** Volcanic edifice failure estimated after a 1-year long unrest by simulating an increase in the injection rate of a mixture of water and carbon dioxide for model A. The failure criterion is computed for a cohesion of 0.5 MPa and friction angles of 20° (top) and 30° (bottom). The contour lines define the failure surfaces for values of cohesion coefficient of 0.5 (black line) and 1 MPa (white line).

**Figure 10 –** Volcanic edifice failure estimated after a 1-year long unrest by simulating an increase in the injection rate of a mixture of water and carbon dioxide for model B. The failure criterion is computed for a cohesion of 0.5 MPa and friction angles of 20° (top) and 30° (bottom). The contour lines define the failure surfaces for values of cohesion coefficient of 0.5 MPa (black line) and 1 MPa (white line).

**Figure 11** - Volcanic edifice failure controlled by only gravitational loading. The failure criterion is computed for a cohesion of 0.5 MPa and friction angles of 20°. The contour lines define the failure surfaces for values of cohesion coefficient of 0.5 MPa (black line) and 1 MPa (white line).