Tectonic stress and renewed uplift at Campi Flegrei caldera, southern Italy: new insights from caldera drilling

Stefano Carlino [1], Christopher R. J. Kilburn[2], Anna Tramelli[1], Claudia Troise[1], Renato Somma[1] and Giuseppe De Natale[1]

[1] Istituto Nazionale di Geofisica e Vulcanologia, Sezione di Napoli, Osservatorio Vesuviano, Via Diocleziano 328, Naples, 80124, Italy, [2] UCL Hazard Centre, UCL, Gower St, London WC1E 6BT, UK

Corresponding author: S. Carlino

stefano.carlino@ingv.it

Key Words: Caldera unrest, minimum principal stress, drilling, Campi Flegrei, leak-off test.

Abstract

We present the first *in-situ* measurement of the stress field in the crust of Campi Flegrei, an active volcanic caldera in Southern Italy. Measurements were performed to depths of 500 m during a pilot study for the Campi Flegrei Deep Drilling Project. The results indicate an extensional stress field, with a minimum horizontal stress about 75-80% of the maximum horizontal stress, which approximately equals the vertical stress. The deviation from lithostatic conditions is consistent with the progressive increase in an applied horizontal stress during episodes of unrest since at least 1969. Because the stress field is evolving with time, the outcome of renewed unrest cannot be assessed by analogy with previous episodes. Interpretations of future unrest must therefore accommodate the possibility that Campi Flegrei is approaching conditions more favourable to a volcanic eruption than has previously been the case. Long-term accumulations of stress are not expected to be unique to Campi Flegrei and so may provide a basis for improved forecasts of eruptions at large calderas elsewhere.

1. Introduction

Campi Flegrei is an active volcanic caldera, 13 km across, that borders the western suburbs of Naples in southern Italy. It is one of the highest-risk volcanic areas in the world, because nearly 2.5 million people are vulnerable to an eruption, most especially the 350,000 who live within the caldera itself. The caldera is underlain by a primary zone of magma storage 1.2-1.5 km thick and with a top about 7.5 km below the surface (Zollo et al., 2008) and, possibly, by a smaller source at a depth pf 3-4 km (Carlino and Somma, 2010). The caldera last erupted in 1538, after a century of net uplift of at least 15 m, centred around the port of Pozzuoli (Figure 1; Bellucci et al., 2006). Following more than 400 years of quiescence, intermittent uplift began in the late 1950s (Del Gaudio et al., 2010). Most of the uplift occurred during two episodes, in 1969-72 and 1982-84, which resulted in a maximum net uplift at Pozzuoli of more than 3 m (Bellucci et al., 2006). The second episode was accompanied by elevated rates of local seismicity, which produced more than 16,000 earthquakes with recorded magnitudes between 0 and 4 (De Natale et al., 1995, 2006). The persistent seismicity damaged older buildings in Pozzuoli and eventually led to the evacuation of some 40,000 people (Barberi et al., 1984).

The most recent episode of uplift began in 2005 and, by the time of writing in 2014, had raised Pozzuoli by an additional ~0.23 m (Figure 2; INGV, 2014). Although 30 times slower than the two previous crises, it is the first episode of sustained uplift since 1984. The potential for major crustal failure during renewed uplift increases with the amount of stress that has already been accumulated in the crust. Such failure would favour the onset of seismic swarms and, if magma were available at shallow depth, also the onset of a volcanic eruption. A continuing concern, therefore, is whether the stresses driving the uplifts have been accumulating in the crust or have been partially relaxed between rapid unrest episodes. To address this concern, we have measured the current stress field as part of the Campi Flegrei Deep Drilling Project (CFDDP), which is a contribution to the

International Continental Scientific Drilling Program (ICDP). The results provide the first direct stress measurements in Campi Flegrei and are consistent with an accumulation of stress since at least the uplift of 1969-72, so that the potential for eruption appears to have increased with successive episodes of unrest. They also extend the database of deep borehole studies in volcanic districts, including Iceland (Haimson and Rummel, 1982), Long Valley caldera in California (Moos and Zoback, 1993), Hawaii (Morin and Wilkens, 2005) and Unzen in Japan (Nakada et al., 2005), and reinforce the importance of borehole data for investigating the structure and evolution of magmatic systems, as well as the interplay between tectonic, magmatic and geothermal processes (Eichelberger and Uto, 2007; Harms et al., 2007).

2. The stress field in Campi Flegrei

Although the stress field in Campi Flegrei had not been measured directly before the CFDDP, recent changes in stress can be constrained from measurements of ground deformation and associated local seismicity. Thus, geodetic movements recorded by levelling surveys during the 1969-72 and 1982-84 uplifts are consistent with two components: radial patterns of deformation about locations near Pozzuoli, produced by buried sources of overpressure, combined with superimposed increases of horizontal extension, of about 1 MPa and 1.4 ± 0.6 MPa respectively (Bianchi et al., 1987; Woo and Kilburn, 2010). Each major episode of unrest was accompanied by volcano-tectonic (VT) earthquakes with hypocentres at depths between 0.5 and 5 km and, at least since 1980, almost 77% have involved normal and strike-slip displacements (Zuppetta and Sava, 1991; De Natale et al., 1995; d'Auria et al., 2015). In terms of principal stresses, normal and strike-slip displacements are characterised respectively by $S_v > S_H \geq S_h$ and $S_H > S_v \geq S_h$, where S_v, S_H and S_h correspond to the local vertical, maximum horizontal and minimum horizontal stresses. Mixed normal and strike slip faulting is thus consistent with values for S_H that vary locally from just greater than to just less than S_v, that is with $S_H/S_v = 1 \pm \alpha$, where α is less than 1.

3. Borehole measurements of stress field

During December 2012, a pilot CFDDP borehole was sunk to a depth of 501 m in the abandoned steel-works at Bagnoli, 4.5 km ESE along the coast from Pozzuoli (Figure 1 and 3). Well logging to 422 m, including a caliper log, showed no evidence of significant rock fracture and breakout around the borehole. After the logging, a standard Leak-Off Test (LOT) was conducted for which a drilling fluid (mud or water) is fed into the well to balance the lithostatic pressure being exerted by the surrounding rock. Operational and logistic problems prevented more extended tests, such as XLOT (Extended LOT). The primary purpose of a LOT is to evaluate the maximum weight of drilling fluid that can be used without inducing hydraulic fracture or damaging the well (Engelder, 1993). LOT data, however, also provide reliable estimates of the minimum *in-situ* principal stress acting in the host rock at the depth of measurement, with an error generally less than 5% of the same value calculated using the more complex XLOT (Zoback et al., 2003; White et al., 2002).

The LOT was performed by pumping water at a constant flow rate of about 6.3 x 10^{-4} m^3 s^{-1} (38 litres min^{-1}), after the well had been cleaned and the blowout preventer locked. The pressure was measured at the wellhead pump at five-samples second. The lowermost 80 m of the well (from depths of 422 to 501 m) was equipped with a slotted casing that permitted contact with the host rock formation (Figure 3), which consisted predominantly of welded tuffs.

Under a constant pumping rate, the rheological behaviour of the host rock can be inferred from the variation of wellhead pressure with time. At Campi Flegrei, the wellhead pressure-time graph can be divided into four main stages (Figure 4): For the first 50 s, the wellhead pressure increased from zero to 1.4 MPa (14 bar) at an approximately constant rate of 0.027 MPa s^{-1}; between 50 and 100 s the change became strongly non-linear, increasing to a turning point (local maximum) at 1.5 MPa, before decreasing to 1.46 MPa. Between 100 and 517 s, the pressure increased to ~1.75 MPa, at a rate that decayed slowly to a mean rate of 8 x 10^{-4} MPa s^{-1}. Between 517 and 634 s, it remained steady at ~1.75 MPa, until the pressure being supplied was reduced and allowed to decrease until the end of the test.

As described by Hickman and Zoback (1983) and Zoback (2007), the initial linear response corresponds to an elastic deformation of the surrounding rock. The onset of non-linear behaviour marks the Leak-Off Pressure (LOP; 1.4 MPa in Figure 4), when a hydraulic fracture formed allowing the escape of the drilling fluid and, hence, a drop in the rate of pressure increase. Initial growth of the fracture occurs at the Formation Breakdown Pressure (FBP; 1.5 MPa in Figure 4), at which stage fluid begins to escape from the well more quickly than it is supplied and so induces a decrease in pressure. Commonly, incipient failure at the FBP leads almost immediately to the propagation of a tensile fracture away from the borehole, so that the FBP represents an upper limit to the wellhead pressure. Occasionally, however, incipient failure begins at a pressure significantly lower than the value for fracture propagation, which, for a circular borehole, is equal to the least principal stress (Jaeger, 1969). In this case, the wellbore pressure continues to increase above the FBP until it has reached the least principal stress (Hickman and Zoback, 1983). The second condition occurred during our test, for which the pressure slowly increased above the FBP to the value for fracture propagation (1.76 MPa), where it remained steady until the test was ended.

During the Campi Flegrei test, the key wellhead pressures for the LOP, FBP and least principal stress were 1.4, 1.5 and ≈1.8 MPa respectively (Figure 4). The total pressure exerted on the host rock is the sum of the wellhead pressure and the hydrostatic pressure of the column of water in the well. For a water density of 1000 kg m^{-3}, the hydrostatic pressures at the top and the base of the test section were 4.1 and 4.9 MPa. Failure may have occurred at any location along the 80 m section of the test. Hence, from the sum of the final and hydrostatic pressures, estimates of the LOP, FBP and least principal stress along the test 80 m of the borehole were 5.5, 5.6 and 5.9 MPa for a well breakout at 422 m, and 6.3, 6.4 and 6.7 MPa for failure at 501 m.

The dominance of normal and strike-slip faulting during unrest indicates that the minimum principal stress is horizontal and so S_h can be equated with the least principal stress from the borehole measurements. For a vertical borehole, the circumferential stress around the well is given by $3S_h$ - S_H - P_p, where S_h and S_H are the minimum and maximum horizontal *in-situ* stresses, and P_p is the pore-fluid pressure of the host rock (Jaeger, 1969; Zoback, 2007). Failure occurs when ~~when~~ the borehole pressure exceeds the stresses acting around the circumference of the well by an amount equal to the tensile strength of the enclosing rock. The borehole pressure at failure P_{bf} is thus:

$$P_{bf} = 3S_h - S_H - P_p + T \tag{1}$$

where T is the magnitude of the host rock's bulk tensile strength. Apart from P_p, the terms in Eq. (1) refer to the total stresses at depth before taking account of pore-fluid pressure.

To a good approximation, P_{bf} corresponds to the FBP (Zoback et al., 2003; Zoback, 2007). The widespread distribution of shallow aquifers across the caldera (Celico et al., 1992) and the presence of water-saturated rock from shallow and deep boreholes (Rosi and Sbrana, 1987), including the CFDDP pilot hole, suggest that pore-fluid pressures remain close to water-saturated conditions to depths of at least 3 km (Carlino et al., 2012; Piochi et al. 2013). Hence P_p can be equated with the hydrostatic load $\rho_{water}gz$, where the density of water $\rho_{water} \approx 1000$ kg m^{-3}), g is gravity and z is depth.

In common with other volcanic districts (Haimson and Rummel, 1982), the shallow crust at Campi Flegrei is expected to have a field-scale tensile strength on the order of MPa or less. Low tensile strengths are favoured also by the low stress drops associated with earthquakes recorded since 1982 and by the scattering analysis of the shallow crust (De Natale et al., 1987; Tramelli et al., 2006; Douglas et al., 2013), as well as by hydrothermal alteration in the shallow geothermal system (Rice and Cleary, 1976). We therefore assume that the crust around the borehole was characterised by a tensile strength close to the lower values on the order of 0.1 MPa found in volcanic environments (Haimson and Rummel, 1982) and so much smaller than the stresses of c. 7-9 MPa in the crust near the base of the borehole (see below).

Setting P_{bf} at the Formation Breakdown Pressure and assuming T is negligible, Eq. (1) yields for a water-saturated crust around the borehole $S_H \approx 7.7$ and 8.5 MPa for failure depths of 422 and 501 m (Figure 5). Data from across Campi Flegrei indicate a mean density of 1,750 ± 150 kg m^{-3} for the upper 500 m of crust, which consists predominantly of partially consolidated tuffs (Piochi et al., 2013). This value has been confirmed during laboratory measurements performed on two samples cored at depth of 438m and 500m respectively, in the CFDDP pilot hole. The vertical lithostatic stress S_v at depths of 422 and 501 m is thus estimated to be 7.2 ± 0.6 and 8.6 ± 0.7 MPa (Figure 5).The corresponding ratios of S_H/S_v are 1.07 ± 0.09 and 0.99 ± 0.08, both of which are consistent with the value of 1 ± α that was previously inferred from the observed mixture of normal and strike-slip faulting. If the tensile strength had been significant, Eq. (1) shows that the resulting ratios of S_H/S_v would increase in value by an amount T/S_v. Thus, for significant tensile strengths on the order of 1-10 MPa, T/S_v is ~1 and S_H/S_v could increase to about 2. Such a large value of S_H/S_v is inconsistent with the observed style of faulting and so supports the initial assumption that T can be neglected.

The values of stress can be checked for consistency against the requirement for the effective stress ratio to lie between a lower limit for lithostatic equilibrium and an upper limit for deformation controlled by frictional slip along faults. The effective stress is the difference between the far-field stress applied to the crust and pore-fluid pressure (S - P_p). Under lithostatic conditions, the ratio $(S_1 - P_p)/(S_3 - P_p)$ of maximum to minimum effective stress is 1. In the critical, non-lithostatic limit for which stresses are resisted by friction along optimally-oriented faults, the ratio is given by (Jaeger et al., 2007):

$$\frac{(S_1 - P_p)}{(S_3 - P_p)} = \left[(1 + \mu^2)^{\frac{1}{2}} + \mu \right]^2 \tag{2}$$

where μ is the coefficient of sliding friction along faults. The optimal faults are oriented at an angle β from the direction of maximum principal stress, where $\tan 2\beta = \pm 1/\mu$. For common values of μ

between 0.6 and 0.7 (Zoback, 2007), the upper limit for the effective stress ratio lies between 3.1 and 3.7 and the optimal faults orientation can be obtained as *tan* $2\beta = \pm 1/\mu$, where β is the angle between maximum principal stress and the fault plane. Setting $S_1 = S_V \sim S_H$ and $S_3 = S_h$, Eq. (2) yields $(S_1 - P_p)/(S_3 - P_p)$ ~2.0 for water-saturated crust at both depths of measurement, confirming that the stresses are within sub-critical conditions. However, critical conditions could be achieved if the minimum principal stress at the measured depths decreased by about 0.6-0.8 MPa. Such conditions would be associated with the onset of significant seismic activity as slip is activated along a population of optimally-oriented faults.

4. Implication for future behaviour of Campi Flegrei caldera

The total stress field is the sum of the effective lithostatic stress (lithostatic stress – P_p), the stress due to uplift, and external stresses (e.g., tectonic stress). By definition, the lithostatic component has the same magnitude in all directions. The axi-symmetric deformation field produced during the 1969-1972 and 1982-1984 uplifts is generally attributed to pressurized sources at depths of about 3 km, with modelled source geometries from circular sills to spheres (Battaglia et al., 2006; Gottsmann et al., 2006; Amoruso et al., 2008; Woo and Kilburn, 2010). The typical variation in topography across the caldera is c. 0.25 km (from c. 50 bsl to 200 m asl), which is less than 2% of the caldera's diameter and less than 10% of the source depth. In common with previous geodetic studies, therefore, the top of the caldera can be approximated to a flat surface (Berrino et al., 1984; Battaglia et al., 2006; Gottsmann et al., 2006; Amoruso et al., 2008; Woo and Kilburn, 2010). For a flat crust that behaves as a mechanically homogeneous medium, the induced changes, ΔS, in each principal stress decay with distance, r, from the centre according to $\Delta S = \Delta S_0\ (r_0/r)^m$ (for r larger than the representative distance between the centre and margins of the pressure source), where ΔS_0 is the additional stress due to the source pressure at a distance r_0 and, to a good first approximation for the preferred source geometries, m = 3

(Fialko et al., 2001; Gudmundsson, 2011). For example, using vertical deformation and choosing surface conditions as reference values, then r_0 = 3 km and, to satisfy the condition of no net vertical stress (= ΔS_0 – Atmospheric Pressure) at the surface, ΔS_0 = Atmospheric Pressure (0.1 MPa). Thus, at a level equivalent to the base of the CFDDP borehole (set for simplicity at r = 2.5 km, corresponding to a borehole depth of 0.5 km), the change in stress induced in the crust by a pressurised magma body at a depth of 3 km will be less than 0.2 MPa. Although the calculation is approximate, it suggests that the stress from such a magma body will be an order of magnitude smaller than the effective lithostatic load of 3.7 MPa at a depth of 0.5 km in water-saturated crust. In the vicinity of the borehole, therefore, the effective lithostatic stresses dominate changes due to axi-symmetric uplift and so would generate a stress field with equal principal stresses, for which S_v - S_H = 0 and S_H - S_h = 0.

The borehole measurements yield S_v - $S_H \approx 0$, as expected, but also a value for S_H - S_h = 1.8 MPa. The difference in horizontal stress lies within the range of 2.4 ± 0.6 MPa for the total additional horizontal extension inferred for the 1969-72 and 1982-84 uplifts (Bianchi et al., 1987; Woo and Kilburn, 2010). This calculation follows from their geodetic modelling of the uplifts in 1969-72 and 1982-84. The similarity of the estimates is remarkable. Given that S_H has remained similar to S_v during the same time interval, the simplest interpretation is that, following more than 400 years of net subsidence across the caldera, the crust was in approximately lithostatic equilibrium before 1969 and that the additional horizontal stress was accumulated between 1969 and 1984 and has since been maintained with only a minor stress relaxation. An important open question is related to the possible source of additional horizontal stress. Two possibilities are that periods of unrest have been accompanied by change in regional tectonic strain rate superimposed on doming of caldera, or that the magma reservoir at 7.5 km spreads horizontally during unrest and drags outward the overlying crust (Figure 6).

An immediate implication for Campi Flegrei is that, by accumulating differential stress by decreasing S_h, successive episodes of uplift increase the potential for seismic swarms and an eruption. Seismic swarms would additionally be favoured when the stresses in the crust satisfy the critical conditions in Eq. (2) for deformation controlled by fault sliding. The borehole data suggest that critical conditions can be expected when S_h is further reduced by about 0.6-0.8 MPa. The 1969-72 and 1982-84 uplifts have each been associated with a reduction in S_h by ~1 MPa (Bianchi et al., 1987; Woo and Kilburn, 2010). A future uplift by the same magnitude (on the order of metres) may therefore be accompanied by rates of seismicity greater than have occurred since 1969. In addition, increases in pore pressure due to disturbances in the circulation of geothermal fluids would also favour an increase in seismicity, without requiring a major increase in the deviatoric stress in the crust (Sibson, 1974; Moos and Zoback, 1993). In this case, even small amounts of renewed uplift (of less than metres) could potentially trigger a high rate of seismicity.

5. Conclusions

Results from the first *in-situ* measurements of crustal stress at Campi Flegrei during the CFDDP pilot study suggest that the caldera has been accumulating a horizontal tensile stress since at least 1969. The stress is associated with episodes of unrest and appears not to be relaxed during intervals of quiescence. The stress field is therefore evolving with time, so that the outcome of renewed uplift cannot be assessed by analogy with previous uplifts. Plans for responding to uplift must therefore accommodate the scenario that the Campi Flegrei caldera is progressively approaching conditions more favourable to elevated seismicity or to a volcanic eruption. The long-term accumulation of horizontal stress before eruptions has also been proposed for Rabaul caldera in Papua New Guinea (Robertson and Kilburn, 2012), and so may represent a condition common among large calderas.

The results must be verified against additional *in-situ* measurements of stress, especially at depths greater than 500 m. Such measurements will be included in the next phase of the CFDDP, for which drilling will continue to depths of 3 km, while the pilot well is being converted into a permanent station for the remote sensing of temperature, pressure, water and gas composition. The results also highlight the important contribution of deep drilling to an improved knowledge of the dynamics of large calderas. In particular, a better definition of stress fields with depth and location across the caldera offer the possibility of understanding why and where magma intrudes in the shallow crust and, from this, of quantifying long- and short-term forecasts of volcanic eruptions.

Acknowledgments

This work and related data were supported by the ICDP Campi Flegrei Deep Drilling Project and by the European GEISER project FP7-ENERGY-2009-1. We also thank the Baker Hughes Company for providing technical assistance during the borehole tests. Agust Gudmundsson, David Hill, Tamsin Mather and an anonymous reviewer provided very helpful comments for improving the manuscript. The borehole data can be requested from Stefano Carlino (stefano.carlino@ingv.it).

References

Amoruso, A., L. Crescentini, and G. Berrino (2008), Simultaneous inversion ofdeformation and gravity changes in a horizontally layered half-space: Evidences for magma intrusion during the 1982-1984 unrest at Campi Flegrei caldrea (Italy), *Earth Planetary Science Letters*, v. 272, p. 181-188.

Barberi, F., G. Corrado, F. Innocenti, and G. Luongo (1984), Phlegraean Fields 1982-1984: Brief chronicle of a volcano emergency in a densely populated area, *Bulletin Volcanology*, v. 47, p. 175-185.

Battaglia, M., C. Troise, F. Obrizzo, F. Pingue, and G. De Natale (2006), Evidence for fluid migration as the source of deformation at Campi Flegrei caldera (Italy), *Geophysical Research Letters*, v. 33, doi:10.1029/ 2005GL024904.

Bellucci, F., J.Y.L. Woo, C.R.J. Kilburn, and G. Rolandi (2006), Ground deformation at Campi Flegrei, Italy: Implications for hazard assessment, in Mechanisms of activity and unrest at large calderas, *Spec Pub Geological Society London*, edited by C. Troise, G. De Natale, and C.R.J. Kilburn, vol. 269, p. 141–158.

Berrino, G., G. Corrado, G. Luongo, and B. Toro (1984), Ground deformation and gravity changes accompanying the 1982 Pozzuoli uplift, Bulletin of Volcanology, v. 47, p.187–200.

Bianchi, R., A. Coradini, C. Federico, G. Giberti, P. Lanciano, J.P. Pozzi, G. Sartoris, and R. Scandone (1987), Modeling of surface deformation in volcanic areas: The 1970–1972 and 1982-1984 crises of Campi Flegrei, Italy, *Journal Geophysical Research*, v. 92, p. 14139–14150.

Bonafede, M. (1991), Hot fluid migration: an efficient source of ground deformation; application to the 1982–1985 crisis at Campi Flegrei—Italy, *Journal Volcanology Geothermal Research*, v. 48, p.187–198.

Carlino, S., and R. Somma (2010), Eruptive versus non-eruptive behaviour of large calderas: the example of Campi Flegrei caldera (southern Italy), *Bulletin Volcanology*, v. 72, no 7, p. 871-886.

Carlino, S., R. Somma, C. Troise, and De G. Natale (2012), The geothermal exploration of Campanian volcanoes: Historical review and future development, *Renewable and Sustainable Energy Reviews*, v. 16, no 1, p. 1004-1030.

Campi Flegrei Deep Drilling Project (CFDDP), *https://sites.google.com/site/cfddpproject/*

Celico, P., M. Dall' Aglio, M.R. Ghiara, D. Stanzione, M. Brondi, and M.Prosperi (1992)Geochemical monitoring of the geothermal fluids in the Phlegraean Fields from 1970 to 1990. *Bolletino Societa Geologicà Italiana*, v. 111, p. 409-472.

D'Auria, L., B. Massa, E. Cristiano, C. Del Gaudio, F. Giudicepietro, G. P. Ricciardi, and C. Ricco (2015), Retrieving the stress field within the Campi Flegrei caldera (Southern Italy) through an integrated geodetical and seismological approach, *Pure and Applied Geophysics*, doi:10.1007/s00024-014-1004-7.

Del Gaudio, C., I. Aquino, G.P. Ricciardi, C. Ricco, and R. Scandone (2010) Unrest episodes at Campi Flegrei: a reconstruction of vertical ground movements during 1905–2009, *Journal Volcanology Geothermal Research*, v. 195, p.48–56.

De Natale G., G. Iannaccone, M. Martini, and A. Zollo (1987), Seismic sources and attenuation properties at the Campi Flegrei volcanic area, *Pure and Applied Geophysics*, 125, 6, p. 883-917, doi 10.1007/BF00879360

De Natale, G., A. Zollo, A. Ferraro, and J. Virieux (1995), Accurate fault mechanism determinations for a 1984 earthquake swarm at Campi Flegrei caldera (Italy) during an unrest episode: Implications for volcanological research, *Journal Geophysical Research*, v. 100 no B12, p. 24167–24185.

De Natale G., S.M. Petrazzuoli, and F. Pingue (1997), The effect of collapse structures on ground deformations in calderas, *Geophysical Research Letters*, 24,13, p. 1555-1558, doi 10.1029/97GL01600

De Natale, G., C. Troise, F. Pingue, G. Mstrolorenzo, L. Pappalardo, M. Battaglia, and E. Boschi (2006), The Campi Flegrei caldera: unrest mechanisms and hazards. In: Troise C, De Natale G, and Kilburn CRJ (eds) Mechanisms of activity and unrest at large calderas, *Geological Society London. Spec. Pub*. v. 269, p. 25–45.

Douglas, J., B. Edwards, V. Convertito, N. Sharma, A. Tramelli, D. Kraaijpoel, B. M. Cabrera, N. Maercklin, and C. Troise (2013), Predicting Ground Motion from Induced Earthquakes in Geothermal Areas, *Bulletin of the Seismological Society of America*, 103,. 3, doi: 10.1785/0120120197.

Eichelberger, J.C., and K. Uto (2007) Active Volcanic Systems, In: Harms, U., C. Koeberl, M.D. Zoback (eds), Continental scientific drilling, Springer, p.213-234

Engelder, T. (1993), Stress regimes in the lithosphere. *Princeton University Press*, p. 475.

Fialko, Y., Y. Kazhan, and M. Simons (2001), Deformation due to a pressurized horizontal circular crack in an elastic half-space, with applications to volcano geodesy, *Geophysical Journal International*, v. 146, p. 181– 190.

Gottsmann, J., H. Rymer, and G. Berrino (2006), Unrest at the Campi Flegrei caldera (Italy): A critical evaluation of source parameters from geodetic data inversion, *Journal Volcanology Geothermal Research*, v. 150, p. 132-145.

Gudmundsson, A. (2011), Rock fractures in geological processes, *Cambridge University Press*, 1-578.

Haimson, B.C., and F. Rummel (1982) Hydrofracturing stress measurements in the Iceland Research Drilling Project drill hole at Reydarfjordur, Iceland, *Journal Geophysical Research*, v.87 (B8) p. 6631-6649.

Harms, U., C. Koeberl, and M.D. Zoback (2007), Continental scientific drilling, Springer, pp. 366

Hickman, S. H., and M.D. Zoback (1983), The interpretation of hydraulic fracturing pressure-time data for in-situ stress determination, *Hydraulic Fracturing Stress Measurements* M. D. Zoback, B. C. Haimson (eds), National Academy Press, Washington, D. C. p. 44–54.

ICDP (International Continental Scientific Drilling Project) *www.icdp-online.org*

INGV (Istituto Nazionale di Geofisica e Vulcanologia), 2014. *Bollettino di Sorveglianza Vulcani Campani*, Novembre 2013, p. 16. Available online at www.ov.ingv.it.

Jaeger, J.C. (1969), Elasticity, fracture and flow, 3rd edition, *Chapman and Hall*, 1– 276.

Jaeger, J.C., N.G.W. Cook, and R.W. Zimmerman (2007), Fundamentals of rock mechanics, 4th edition, *Blackwell*, 1-475.

Morin, R.H., and R.H. Wilkens (2005) Structure and stress state of Hawaiian island basalts penetrated by the Hawaii Scientific Drilling Project deep core hole, *Journal Geophysical Research*, doi: 10.1029/2004JB003410

Moos, D., and M.D. Zoback (1993) State of stress in the Long Valley caldera, California, *Geology*, v.21, p. 837-840

Nairn, I. A., C. O. McKee, B. Talai, and C. P. Wood (1995), Geology and eruptive history of the Rabaul Caldera area, Papua New Guinea, *Journal Volcanology Geothermal Research*, 69, 255 – 284

Nakada, S., K. Uto, S. Sakuma, J.C. Eichelberger, and H. Shimizu (2005), Scientific results of conduit drilling in the Unzen Scientific Drilling Project (USDP), *Scientific Drilling*, doi:10.2204/iodp.sd.1.03.2005.

Piochi, M., C. R.J. Kilburn, M. Di Vito, A. Mormone, A. Tramelli, C. Troise, and G. De Natale (2013), The volcanic and geothermally active Campi Flegrei caldera: an integrated multidisciplinary image of its buried structure, *International Journal Earth Science* (Geol Rundsch), doi: 10.1007/s00531-013-0972-7

Rice, J.R. and M.P. Cleary (1976), Some basic stress diffusion solutions for fluid-saturated elastic porous media with compressible constituents, *Rev. Geophys*., 14, p. 227-241, doi 10.1029/RG014i002p00227.

Robertson, R.M. and C.R.J. Kilburn (2012), Precursors to eruptions at large calderas: Rabaul Caldera, Papua New Guinea, *Cities on Volcanoes 7, Colima, Mexico, 19-23 November 2012*.Rosi, M., and A. Sbrana (1987), Phlegrean fields. Rosi M, Sbrana A (eds) (1987) Phlegrean fields. Consig Naz Ric, *Quaderni de la Ricerra Scientifica* v. 114, p. 1–175

Sibson, R.H. (1974) Frictional constraints on thrust, wrench and normal faults, *Nature*, v.249, p. 542-544

Tramelli, A., E. Del Pezzo, F. Bianco, and E. Boschi (2006), 3D scattering image of the Campi Flegrei caldera (Southern Italy): New hints on the position of the old caldera rim, *Physics of the Earth and Planetary Interiors*, 155, 3–4, pag 269-280, ISSN 0031-9201.

Vitale, S., and R. Isaia (2013), Fractures and faults in volcanic rocks (Campi Flegrei, southern Italy): insight into volcano-tectonic processes. *International Journal Earth Science*, doi:10.1007/s00531-013-0979-0.

White, A. J., M.O. Traugott, and R.E. Swarbrick (2002), The use of leak-off test as means of predicting minimum in-situ stress, *Petroleum Geoscience* v. 8, p.189-193

Woo, J.Y.L., and C.R.J. Kilburn (2010), Intrusion and deformation at Campi Flegrei, southern Italy: Sills, dikes, and regional extension, *Journal Geophysical Research*, doi:10.1029/2009JB006913.

Zoback, M.D., 2007, Reservoir Geomechanics. Cambridge University Press, p. 449

Zoback, M.D., C.A. Barton, M. Brudy, D.A. Castillo, T. Finkbeiner, B.R. Grollimund, D.B. Moos, P. Peska, C.D. Ward, and D.J. Wiprut (2003), Determination of stress orientation and magnitude in deep wells, *International Journal Rock Mechanis and Mining Scienc*e, v. 40 (7–8) p. 1049-1076.

Zollo, A., N. Maercklin, M. Vassallo, D. Dello Iacono, J. Virieux, and P. Gasparini (2008), Seismic reflections reveal a massive melt layer feeding Campi Flegrei caldera, *Geophysical Research Letters*, doi:10.1029/2008GL034242.

Zuppetta, A., and A. Sava (1991), Stress pattern at Campi Flegrei from focal mechanisms of 1982-1984 earthquakes (Southern Italy), *Journal Volcanology Geothermal Research*, v. 48, p. 127-13.

Figures Captions

Figure 1. Digital map of the Campi Flegrei caldera showing: caldera rim (*green*), main faults (*blue*), principal post-caldera cones and craters (*brown*), relict structures (*yellow*). The red circle is the location of the CFDDP pilot hole at Bagnoli.

Figure 2. Ground uplift (*red line*) and cumulative seismic energy (*blue line*) at Campi Flegrei caldera from 2004 to present time. Seismic activity during this period generally occurred as swarms with maximum magnitude smaller than 2.0. The unrest, which started in 2005, is ongoing with latest period of uplift (July 2014) occurred at rate of about 1.5 cm/month. The insert on the top left is the recorded uplift from 1970 to 2004. (Data from the INGV seismic and ground deformation database, *www.ov.ingv.it*).

Figure 3. CFDDP pilot well section. The LOT was performed along the 80m of open liner from 422m to 501m of depth.

Figure 4. Wellhead pressure-time plot during the LOT. The LOP, FBP and steady conditions represent stages in the hydraulic failure and fracture propagation (see text for details). The steady value of ~1.75 MPa is the wellhead pressure when the borehole pressure equals the minimum horizontal stress. The bleed-off identifies the pumping stop.

Figure 5. Trend of different pressures in the CFDDP pilot hole and range of inferred SH, Sh and Sv values along the open section of the well (see text for details).

Figure 6. Interpretive cartoon of Campi Flegrei caldera structure with indication of possible magmatic sources, extension related to spreading of deep magma chamber (this process could produce the outwards dragging of the above brittle crust), and rheology of the crust. It is also indicated the location of CFDDP pilot well (not to scale).

Figure 1.

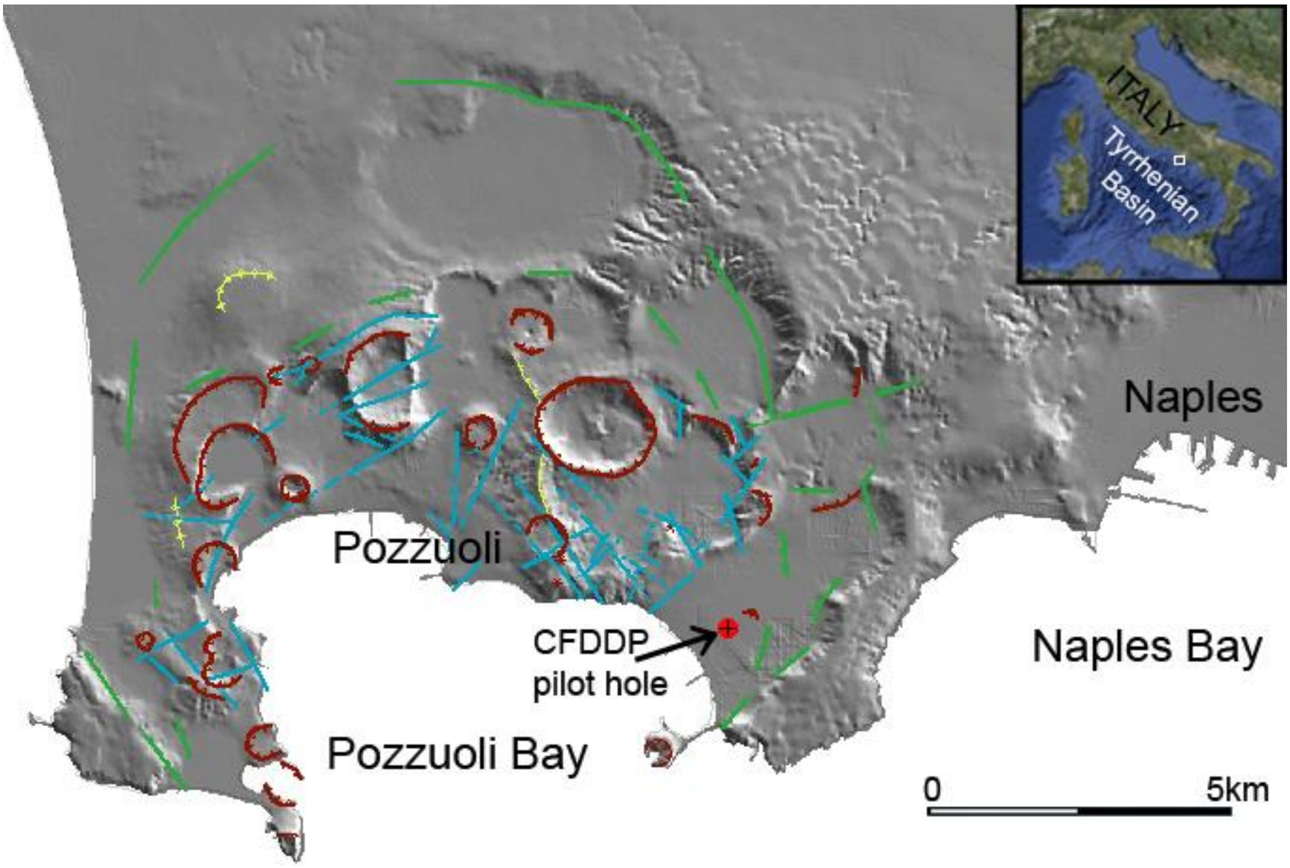
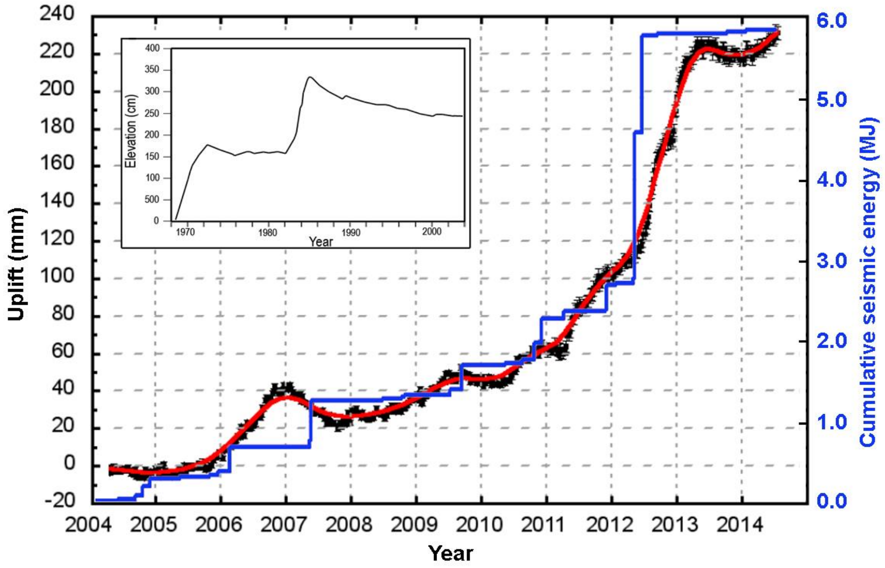
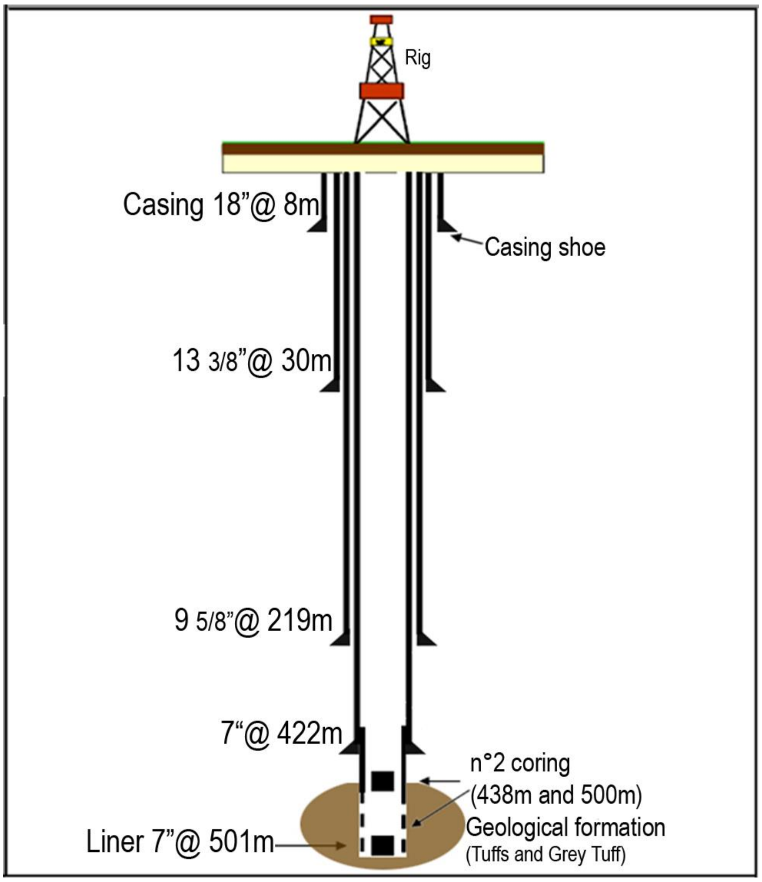

ITALY
Tyrrhenian
Basin
Naples
Pozzuoli
CFDDP
pilot hole
Naples Bay
Pozzuoli Bay
0
5km

Figure 2



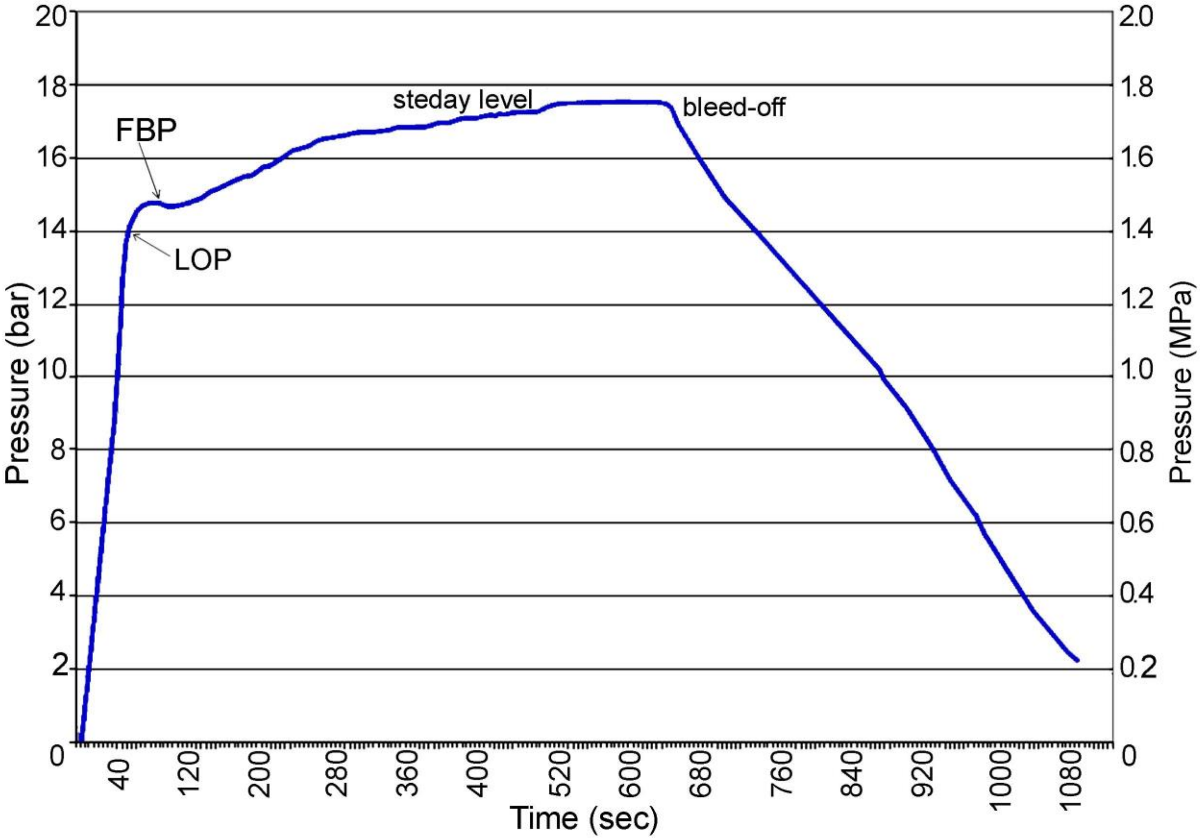
Uplift (mm)
Cumulative seismic energy (MJ)
Year
Elevation (cm)
Year

Figure 3



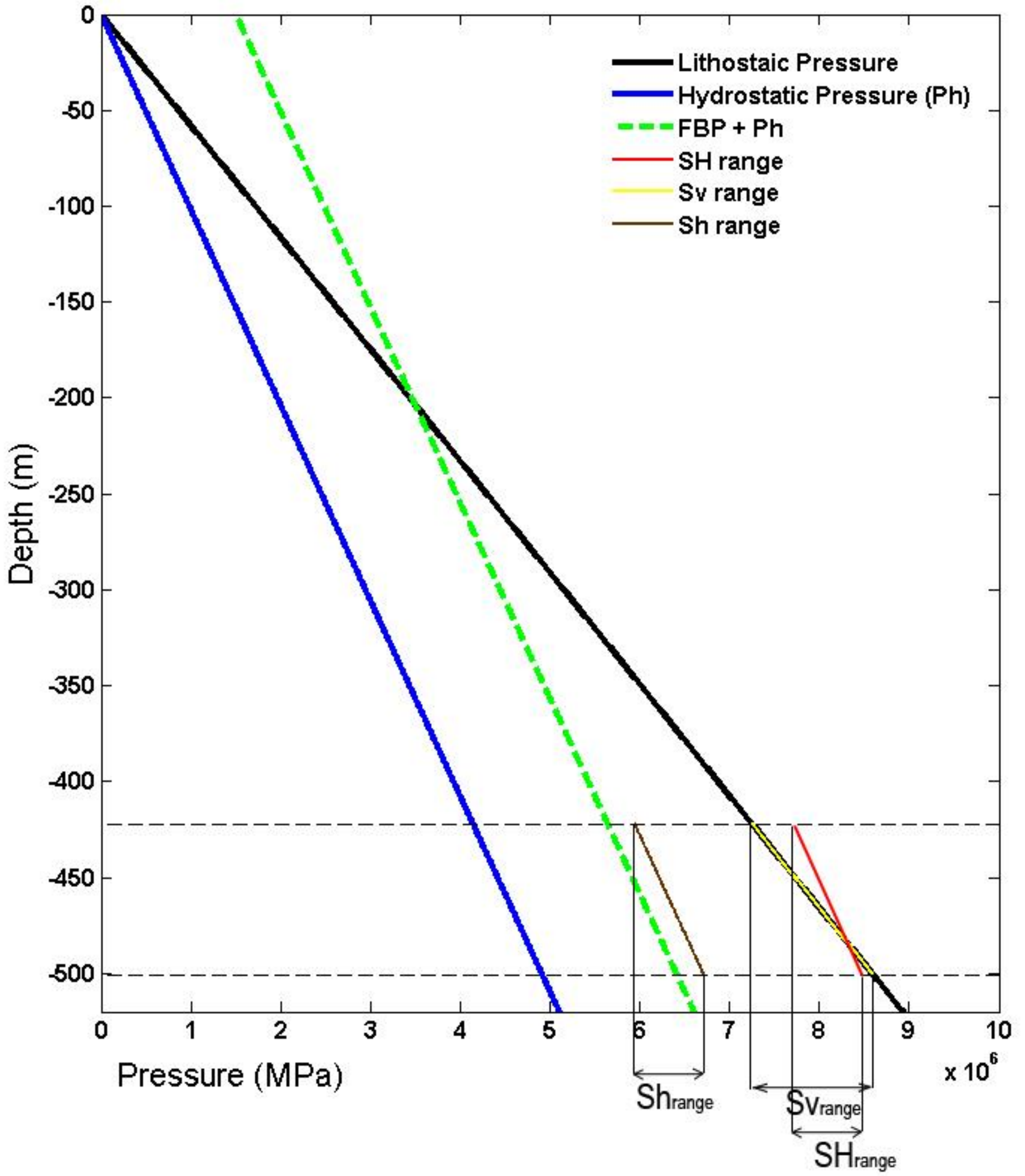
Rig
Casing 18”@ 8m
Casing shoe
13 3/8”@ 30m
9 5/8”@ 219m
7“@ 422m
n°2 coring
(438m and 500m)
Geological formation
(Tuffs and Grey Tuff)
Liner 7”@ 501m

Figure 4



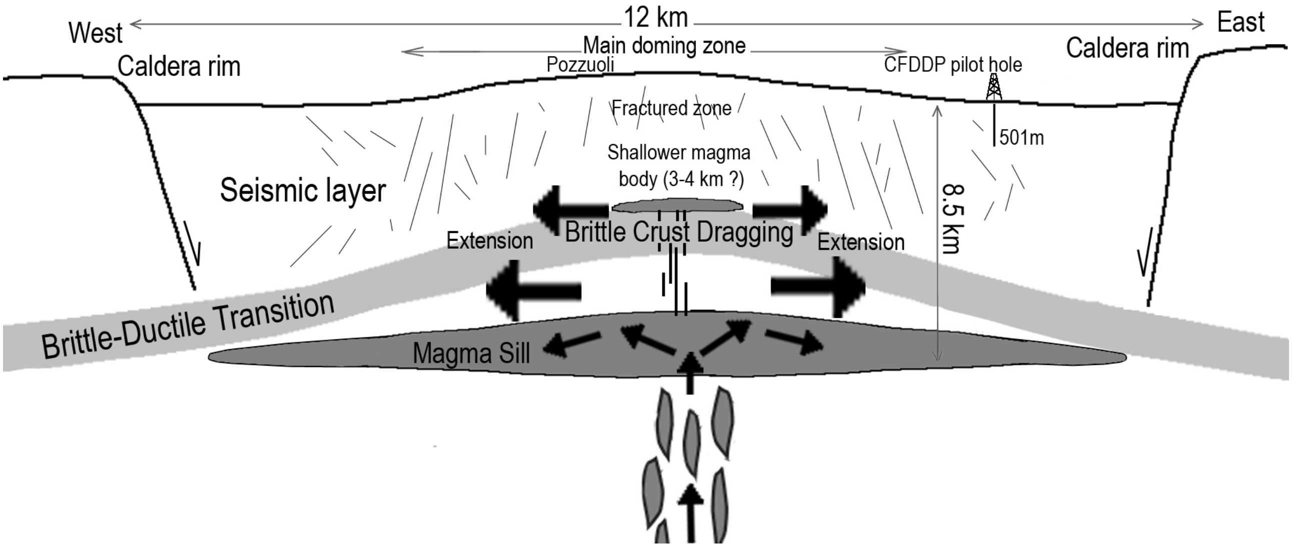
FBP
LOP
steday level
bleed-off
Pressure (bar)
Pressure (MPa)
Time (sec)

Figure 5



Lithostaic Pressure
Hydrostatic Pressure (Ph)
FBP + Ph
SH range
Sv range
Sh range
Depth (m)
0
-50
-100
-150
-200
-250
-300
-350
-400
-450
-500
0
1
2
3
4
5
6
7
8
9
10
Pressure (MPa)
x 10^6
Sh range
Sv range
SH range

Figure 6



12 km
West
East
Caldera rim
Main doming zone
Pozzuoli
CFDDP pilot hole
Caldera rim
Fractured zone
501m
Shallower magma body (3-4 km ?)
Seismic layer
8.5 km
Brittle Crust Dragging
Extension
Extension
Brittle-Ductile Transition
Magma Sill