Seismic reflections reveal a massive melt layer feeding Campi Flegrei caldera

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Seismic reflections reveal a massive melt layer feeding Campi Flegrei caldera

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[1] Campi Flegrei is an active, resurgent caldera that is located a few kilometres west of the city of Naples, a densely populated urban settlement in southern Italy. Identifying, locating at depth and better defining the geometry of the magma feeding system of the caldera is highly relevant for assessing and monitoring its volcanic hazard. Based on a high resolution seismic reflection data set, we investigated the deep structure of the volcano. Here we show that seismic wave amplitude variations with distance from the radiating source provide clear evidence for large amplitude seismic reflections from the top of an extended supercritical fluid-bearing rock formation at about 3,000 m and of an about 7,500 m deep, 1,000 m thick, low velocity layer, which is associated with a mid-crust, partial melting zone beneath the caldera. The modeling of magma properties based on measured seismic velocities indicates a relatively high melt percentage (in the range 80–90%). These new data suggest that a large magmatic sill is present well within the basement formations, which is possibly linked to the surface through a system of deep fractures bordering the caldera. The lateral extension and similar depth of the melt zone observed beneath the nearby Mt.Vesuvius support the hypothesis of a single continuous magma reservoir feeding both of these volcanoes. **Citation:** 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, *Geophys. Res. Lett.*, *35*, L12306, doi:10.1029/2008GL034242.

1. Introduction

[2] Campi Flegrei, Mount Vesuvius and Ischia are active volcanoes that are threatening the densely inhabited city of Naples and its surrounding neighbourhoods. These Neapolitan volcanoes are located inside the graben-like structure of the Campanian Plain, at the eastern margin of the Tyrrhenian Sea. Campi Flegrei is a partly submerged caldera, wherein a large part of the city of Naples is located. Its shape is believed to be a consequence of two huge areal collapses generated by large explosive eruptions that occurred 39 ky and about 15 ky ago [*Deino et al.*, 2004; *De Vivo et al.*, 2001]. Several intra-caldera eruptions have occurred over the last 10 ky, the most recent of which gave rise to a 130-m-high spatter cone (Mount Nuovo), in 1538. Since that time, the caldera floor has been continuously sinking, at an average speed of 1.3 cm per year [*Dvorak and Mastrolorenzo*, 1991; *Berrino et al.*, 1984]. Two uplift episodes occurred in 1970–1972 and 1982–1984, with about 3.5 m of cumulative uplift, which amply illustrates the unrest conditions of the caldera. Over the last two decades, about 30% of this uplift has been recovered by renewed sinking.

[3] The inflation episodes of 1970–1972 have been accompanied by a moderate, low magnitude seismicity, while earthquake activity was very intense during the 1982–1984 unrest, with most of events concentrated at depths shallower than 3 km [*Aster and Meyer*, 1988], in the on-shore area where the maximum uplift gradient has been measured [*Berrino et al.*, 1984].

[4] The structure of the caldera has been investigated previously via several 1-to-3-km-deep boreholes, using local earthquake seismic tomography, gravity and magnetic surveys and some teleseismic and wide-angle seismic observations [*Rosi and Sbrana*, 1987; *Aster and Meyer*, 1988; *Ferrucci et al.*, 1992]. The boreholes have shown that the caldera is characterized by high temperatures at shallow depths (390°C measured at 3 km depth) [*De Lorenzo et al.*, 2001]. The caldera appears to be filled by a layer of volcanic deposits, a few kilometres thick, which forms an inner basin that is characterized by a low P-velocity (Vp), high Vp/Vs, and high P-wave attenuation and low density. The possible occurrence of a magmatic reservoir at about 4–5 km depth has been hypothesized, mainly based on the extrapolation at depth of temperature data and teleseismic observations. However it should be noted that the seismicity distribution excludes the hypothesis of very shallow magma bodies occurring above 4 km depth.

[5] In September 2001, an extended marine, active seismic survey (SERAPIS) was carried out in the submerged part of the caldera (in the bays of Naples and Pozzuoli), with the aim of investigating its shallow structure and detecting the deeper magmatic system using seismic tomography and migration techniques [*Zollo et al.*, 2003; *Judenherc and Zollo*, 2004] (Figure 1). The inferred, high resolution, 3-D tomographic images of the Campi Flegrei caldera revealed the presence of a ring-like, high P-velocity and high density body at 800–2,000 m in depth, with a diameter of about 8,000–12,000 m and a thickness of 1,000–2,000 m, which has been interpreted as the buried rim of the caldera [*Zollo et al.*, 2003]. The eastern side of the caldera rim appears to be bordered by a regional SW-NE normal fault that affects the carbonate basement underlying the volcano structure down to an approximate depth of 4,000–5,000 m, with a scarp height of 1,000–2,000 m

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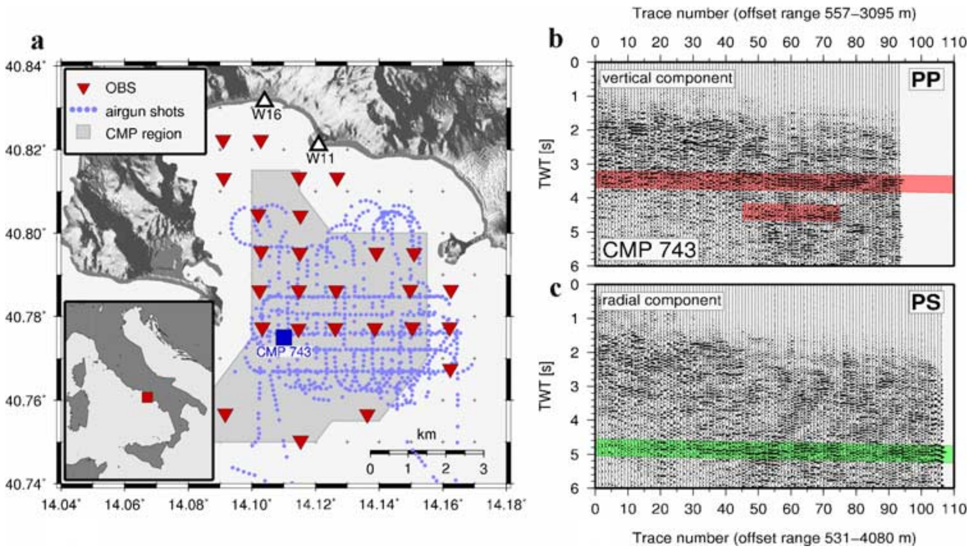

Figure 1. Acquisition geometry map and examples of recorded seismic sections. (a) The SERAPIS source geometry followed a N-S and E-W grid pattern for a total explored surface of 5 $\times$ 5 km^2. The distance between shots along each profile was about 125 m, and the distance between two adjacent profiles was 250 m. The gray-shaded area marks the region of CMPs analyzed. Example of (b) vertical- and (c) horizontal-component seismograms at CMP 743, sorted by increasing source-receiver offset, bandpass-filtered (5–15 Hz), and trace-normalized. The shaded regions mark PP and PS reflections from the reflector at 7.5 km depth, and the secondary PP arrival around 4.5 s (Figure 1b) indicates a reflection from the bottom of the low-velocity layer.

[*Judenherc and Zollo*, 2004]. This suggest that the eruptive activity of the Campi Flegrei caldera is fed by regional fractures that are directly linked to a deep magma reservoir.

2. Data Acquisition and Analysis

[6] The results shown in the present study are based on the active seismic reflection data from this SERAPIS experiment. To apply reflection processing techniques to vertical and horizontal component seismograms, the traces from 75,000, three-component records acquired by 30 sea-bottom receivers were initially arranged in three-dimensional (3-D) common mid-point (CMP) gathers (500 $\times$ 500 m cells), following the approach used by *Henrys et al.* [2004]. In the central part of Pozzuoli Bay, each gather analyzed was formed by more than 120 recordings. The processing included 5–15 Hz band-pass filtering for the amplitude analyses and additional trace normalization for phase identification.

[7] A P-to-P and P-to-S move-out analysis was applied using an average 1-D velocity model for the Campi Flegrei bay area that was obtained from the 3-D tomographic model [*Zollo et al.*, 2003]. The identification of reflection events was primarily based on lateral waveform coherency, move-out alignments and stack amplitudes of both the vertical and horizontal component sections. In addition, waveform polarization was used for P-to-S converted phases.

[8] Three major reflection events were identified at P-to-P travel times (TWTs) of about 0.6 s, 2.0 s and 3.6 s, corresponding to reflectors at 600 m, 2,700 m and 7,500 m in depth, respectively (Figures 2a and S1).[1] Kinematic ray tracing provided a preliminary 1-D layered velocity model that fitted the observed travel-times and served as a reference model for the amplitude analysis. We generalized the standard amplitude versus offset (AVO) technique to estimate the elastic contrasts from the PS-to-PP amplitude ratios seen in a wide range of source-receiver offsets. A constant P-velocity (*Vp*), a P-to-S velocity ratio (*Vp*/*Vs*), and a density ρ was assigned to each layer, with the contrasts of the three parameters characterizing a given reflector.

3. Modelling of Seismic Wave Amplitudes Versus Offset

[9] The amplitudes of the reflected/converted waves at the same interface were extracted from the seismograms at several source-to-receiver offsets, covering a distance range that was large enough for a significant *PS-to-PP* amplitude ratio variation with offset. The velocity structure above the reflecting interface was constrained by the travel times, and the theoretical *PS-to-PP* amplitude ratios were computed using dynamic ray modeling, including the correction for geometrical divergence. Finally, the unconstrained model parameters were varied to minimize the RMS misfit between the observed and the theoretical ratios. The best-fit model was found by a grid search through the parameter space, combined with a local downhill simplex optimization (see Figure S2).

[10] Figures 2b and 2c show the mean ($\pm$standard deviation) P-to-S amplitude ratios measured at the seismic interfaces located at about 2,700 m and 7,500 m in depth, respectively. The shallower interface shows amplitude ratios that reach a maximum at 2,500 m offset, and towards greater offsets there is a decrease followed by a rapid increase. Based on comparisons with theoretical curves, this behavior suggests a well-constrained positive *Vp* contrast (from 3,500 to 4,700 m/s), and the minimum at about 3,300 m offset is related to the critical incidence of the

[1]Auxiliary materials are available in the HTML. doi:10.1029/2008GL034242.

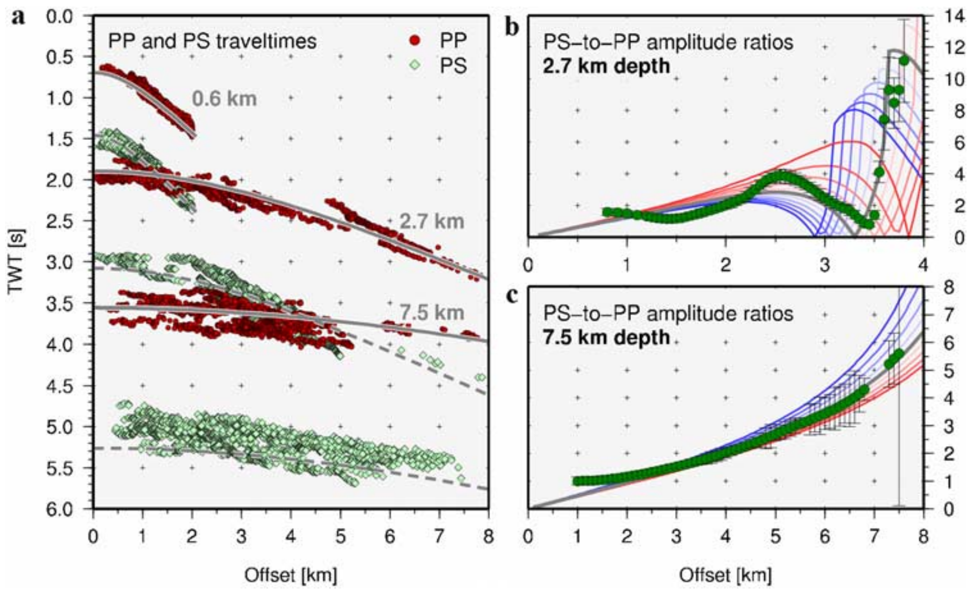

Figure 2. Travel time curves. (a) P-to-P and P-to-S travel time picks (dots) for three major reflectors overlaid with the theoretical times (gray lines) from a 1-D average model. Average of the amplitude ratio variations seen with the offset for reflectors at (b) 2,700 m and (c) 7,500 m. The figure shows the observed amplitude ratio (green dots), as compared with a set of theoretical curves computed for the Vp and Vp/Vs velocity contrast models. Given the best-fit model parameters at each interface, the blue and red curves indicate 10% positive and negative model variation with respect to Vp. Error bars for the estimated amplitude ratios are also shown. Density contrasts at both interfaces have not been included in the analysis, as this parameter has a minor effect on amplitude ratio variation with offset.

P-to-S rays at the interface. The best fit between measured (dots) and modeled (line) *P-to-S* ratios was obtained for a *Vp* increase and a *Vp*/*Vs* decrease from 1.70 to 1.58.

[11] The sharp increases in both *Vp* and *Vs* seen at the seismic discontinuity at 2,700 m in depth beneath Campi Flegrei indicate the presence of a marked lithology transition that is characterized by P- and S-velocity increments of about 35% and 45%, respectively, as compared to the rock velocities in the layer above (Figures 3b and 3c). This abrupt change can be related to the occurrence of thermo-metamorphic rock formations, such as those in the Mofete and San Vito wells at about 2.5 km in depth along the coastal Pozzuoli Bay area [*Rosi and Sbrana*, 1987]. The decrease in the *Vp*/*Vs* ratio marks the presence of a densely fractured, gas-and/or brine-bearing rock layer, due to crack opening induced by increasing pore pressure [*Dvorkin et al.*, 1999].

[12] The presence of a gas-bearing formation underneath the Campi Flegrei caldera has been previously hypothesized from tomographic velocity analyses, at an approximate depth in the range of 2,000–4,000 m [*Vanorio et al.*, 2005]. A fluid substitution analysis from initial gas to brine saturation was applied to determine the elastic parameters of the rock from cores extracted from the Mofete and San Vito wells, but this provides variations in *Vp*/*Vs* produced by fluid changes that are smaller than the uncertainty of the data. Fluid properties are calculated for temperatures of 35°C and 400°C at a pressure of 35 MPa, with the mineral bulk modulus set to $K0$ = 50 GPa. We used the observed Vp = 4.7 km/s, Vp/Vs = 1.58, and assumed a brine with salinity S = 0.05 and gas gravity G = 1.5 ($CO2$).

[13] However the temperature of about 400°C measured at 3000 m by the deepest borehole is close to the critical point of such a brine, thus suggesting the probable rock saturation with a fluid at near- or super-critical conditions.

[14] For the deeper seismic discontinuity at about 7,500 m in depth, the mean P-to-S amplitude ratios increase monotonically with offset, as expected for a reflection at a moderate to strong negative P-velocity contrast across the interface. The theoretical modeling of amplitude ratios of the P-reflection amplitudes from the two interfaces constrain Vp in the layer below the 7,500-m discontinuity to values around 2,800 m/s (Figure 2c), i.e., a strong negative Vp contrast with an estimated Vp/Vs increase from 1.65 to 2.35, or to higher values (Figures 3a and 3b). These values indicate that the 7,500-m interface is the top of a low velocity layer, whose seismic velocities are consistent with values expected for a magma body set in a densely fractured volume of rock [*Murase and McBirney*, 1973; *Rivers and Carmichael*, 1987; *Kress et al.*, 1988]. Similar values of P- and S-velocities for mid-crustal melts were estimated seismically at the Katla and Grímsvötn volcanoes in Iceland [*Gudmundsson et al.*, 1994; *Alfaro et al.*, 2007] and at the East Pacific Rise [*Taylor and Singh*, 2002; *Singh et al.*, 1998].

[15] The melt fraction in the layer below the 7,500-m reflector at the Campi Flegrei caldera can be estimated from the measured seismic velocities by computing the Hashin-Shtrikman bounds generalized for a two-phase medium [*Mavko et al.*, 1998], which gives the narrowest possible range of the elastic moduli without specifying the geometries of the constituents (see Figure S3). We used Vp = 5,500 m/s and Vs = 3,175 m/s for the solid phase, and Vp = 2,400 m/s and Vs = 0 m/s for the pure melt, while the density was held constant at ρ = 2,600 kg/m^3. The results

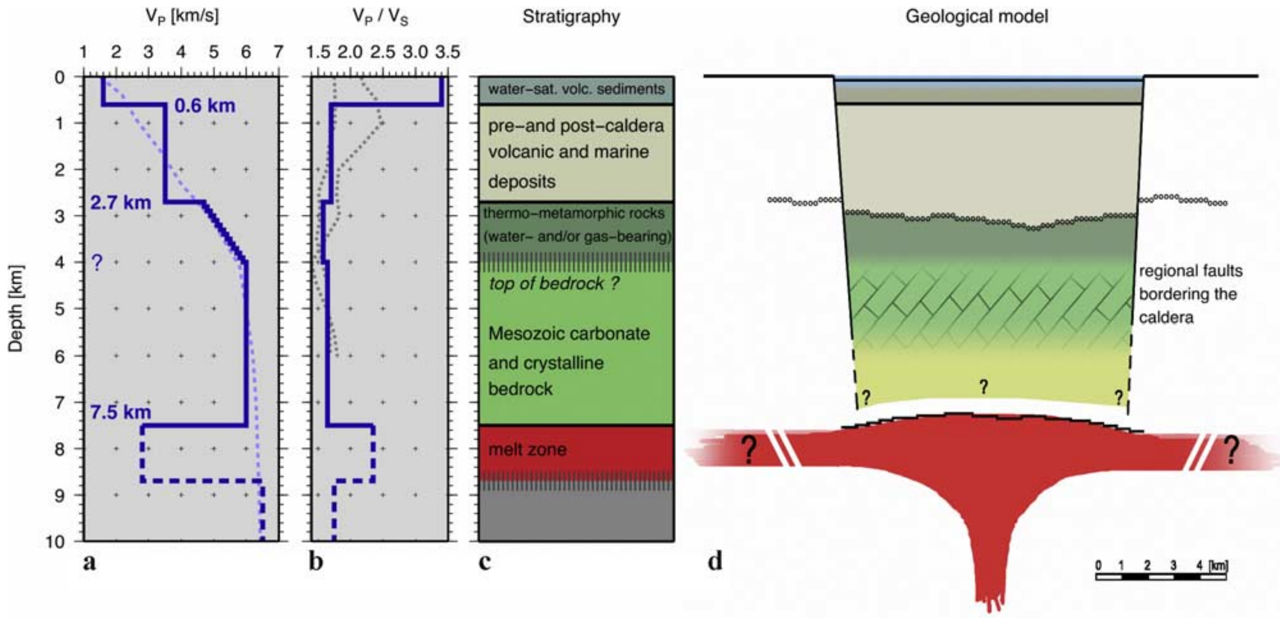

Figure 3. Geophysical and structural model of the deep structure of the Campi Flegrei caldera. (a) Average 1-D P-velocity model for the Campi Flegrei caldera, based on *PP* and *PS* travel times, and on *PS-to-PP* amplitude ratios. The dashed line is the average of the 3-D P-velocity model in the study area. (b) *Vp*/*Vs* ratio as a function of depth, as estimated in the present study. The dotted lines are two *Vp*/*Vs* depth profiles that were estimated from the local earthquake tomography [*Vanorio et al.*, 2005]. C: Stratigraphic model. D: Geological sketch model of the Campi Flegrei caldera.

indicate the presence of 65%–75% melt directly beneath the reflector. However it has been proposed that magmas at depth could glassify in cooler regions generating significant Vs and changing Vp to an unrelaxed value [*Dingwell and Webb*, 1989, 1990]. The Hashin-Strikman bounds have been recalculated with nonzero Vs for the melt, which generally leads to a higher percentage of partial melt in the Campi Flegrei melt zone. For instance, using Vp = 2,400–2,600 m/s and Vs = 800–1,000 m/s for the melt, our model suggests the presence of around 80%–90% partial melt (up to 100% possible).

[16] After normal move-out correction, the majority of the analyzed CMP gathers show evidence for an energetic arrival 0.7 s to 1.2 s after the P-to-P reflected wave generated at the 7,500-m discontinuity. According to theoretical travel-time and waveform computations, this secondary arrival can be interpreted as a reflected P-wave at the bottom of the low-velocity layer, the thickness of which can be estimated as being in the range of 1,200–1,500 m.

4. Discussion and Conclusions

[17] Recent geochemical investigations of phenocryst-hosted melt inclusions from Campi Flegrei shoshonites that erupted about 10 ky ago provide two ranges of high and low pressure trapping depths, at 5,000–9,000 m and 1,500–2,000 m, respectively [*Mangiacapra et al.*, 2007]. Moreover, an approximate depth of 9,000 m was inferred for the crystallization of the shoshonite phenocryst assemblage.

[18] The latter is consistent with the depth of the magma layer detected by seismic reflection amplitudes (Figure 3). The modeling of magma properties based on measured seismic velocities indicates the occurrence of a partial melting zone, with a relatively high melt percentage (in the range 65–90%), although this value critically depends on the assumed value of shear modulus for the melt. *James et al.* [2004] measured the shear moduli of lavas from different volcanoes and in particular, for Mt. Vesuvius they found shear moduli up to about 2.5–3.0 GPa for temperatures larger than 1,100°C, which corresponds to Vs of about 1,000 m/s. Assuming a similar elastic behaviour for magmas beneath Campi Flegrei and Vesuvius would therefore constrain the melt percentage to 80–90%.

[19] The occurrence of magma batches at shallow depths beneath the Campi Flegrei caldera is not supported by seismic tomography and reflection amplitude analysis, although the spatial resolution of seismic waves might not be adequate for the detection of kilometer-scale, or smaller size, magma bodies.

[20] The horizontal extension of the melt layer beneath Campi Flegrei can be estimated to be not less than 30 km^2 and having an approximate thickness of 1 km.

[21] A similar flat, low velocity layer was inferred underneath the nearby Mount Vesuvius volcano [*Auger et al.*, 2001]. The analogy of geometry, depth and thickness indicate the possibility that a single continuous magma reservoir feeds both of these volcanoes. This is in apparent contrast with the significant compositional differences between the magma that have erupted at Campi Flegrei and Mount Vesuvius, which appear to be at least partly produced by an intra-crustal differentiation processes. However, different and independent magma batches can occur in a magma sill if lateral temperature variations exist. Temperature decreases in the magma of 200–300°C will produce viscosity barriers around the hotter zones that are not revealed by seismic methods.

[22] It has been shown that if the magma sill is fed from the mantle and the velocity of the magma ascent in the sill-feeding conduits is more than twice the velocity of the horizontal displacement in the sill, colder zones will develop, separating different hot magma bodies [*De Lorenzo et al.*, 2006]. If Campi Flegrei and Mount Vesuvius are located above two different sill-feeding conduits, hotter and independent magma batches can settle and differentiate separately within the low velocity layer as long as the deep-feeding process continues.

[23] Several evidences exist for wide mid-crustal magma sills beneath different volcanic regions worldwide [*Auger et al.*, 2001, and references therein], supporting the hypothesis for a similar feature extending beneath Mt.Vesuvius and Campi Flegrei volcanoes. *Balch et al.* [1997] reports the areal extent of the Socorro magma body as about 3400 km^2 as inferred from body wave reflections on microearthquake records. *Chang et al.* [2007] found an inflating magma sill of $\sim$1,200 km^2 at 6–14 km depth underneath Yellowstone caldera modeling GPS and InSar data. Using local and teleseismic receiver functions *Zandt et al.* [2003] identified the Altiplano-Puna magma sill, about 60,000 km^2 wide and 1 km thick, at about 20 km depth, characterized by a low S-velocity layer (Vs $\leq$ 1 km/s). The widespread occurrence of nearly flat magma sills at mid-crustal depths can be originated by neutral buoyancy conditions for magmas uprising from the mantle [*Christensen*, 1982]. The formation of a wide and thin sill can be therefore favoured by the slow accumulation of magma rising from the mantle to the mid-crustal reservoir, under neutral buoyancy conditions and in the presence of an extensional tectonic regime similar to that active in the Neapolitan area for the past 2 million years at least.

[24] The existence at Campi Flegrei caldera of a shallow, extended and fractured rock layer saturated with supercritical fluids, above a mid-crustal melt zone (as sketched in Figure 3c), is also consistent with similar observations at other unrest calderas, such as Yellowstone and Long Valley [*Husen et al.*, 2004; *Chang et al.*, 2007; *Sorey et al.*, 1998]. In these cases an extensive gas/brine layer overlying magma bodies has also been widely interpreted. We remark that it can have a significant role in the eruption mechanism style, increasing its explosiveness; this could be the cause of the prominent inflation/deflation pattern that is characteristic of the Campi Flegrei volcano.

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References

Alfaro, R., B. Brandsdóttir, D. P. Rowlands, M. T. White, and R. S. Gudmundsson (2007), Structure of the Grimsvötn central volcano under the Vatnajökull Icecap Iceland, *Geophys. J. Int.*, *168*, 863–876.

Aster, R. C., and R. P. Meyer (1988), Three-dimensional velocity structure and hypocenter distribution in the Campi Flegrei caldera, Italy, *Tectonophysics*, *168*, 195–218.

Auger, E., P. Gasparini, J. Virieux, and A. Zollo (2001), Seismic evidence of an extended magmatic sill under Mt. Vesuvius, *Science*, *294*, 1510–1511.

Balch, R. S., H. E Hartse, A. R. Sanford, and L. Kuo-Wan (1997), A new map of the geographic extent of the Socorro mid-crustal magma body, *Bull. Seismol. Soc. Am.*, *87*, 174–182.

Berrino, G., G. Corrado, G. Luongo, and B. Toro (1984), Ground deformation and gravity changes accompanying the 1982 Pozzuoli uplift, *Bull. Volcanol.*, *47*, 187–200.

Chang, W.-L., R. B. Smith, C. Wicks, J. M. Farrell, and C. M. Puskas (2007), Accelerated uplift and magmatic intrusion of the Yellowstone caldera, 2004 to 2006, *Science*, *318*, 952–956.

Christensen, N. (1982), *Handbook of Physical Properties of Rocks*, vol. 11, edited by R. S. Carmichael, pp. 1–22, CRC Press, Boca Raton, Fla.

Deino, A. L., G. Orsi, S. de Vita, and M. Piochi (2004), The age of the Neapolitan Yellow Tuff caldera-forming eruption (Campi Flegrei caldera - Italy) assessed by $^{40}Ar/^{39}Ar$ dating method, *J. Volcanol. Geotherm. Res.*, *133*, 157–170.

De Lorenzo, S., A. Zollo, and F. Mongelli (2001), Source parameters and three-dimensional attenuation structure from the inversion of microearthquake pulse width data: Qp imaging and inferences on the thermal state of the Campi Flegrei caldera (southern Italy), *J. Geophys.Res.*, *106*, 16,265–16,286.

De Lorenzo, S., V. Di Renzo, L. Civetta, M. D'Antonio, and P. Gasparini (2006), Thermal model of the Vesuvius magma chambre, *Geophys. Res. Lett.*, *33*, L17302, doi:10.1029/2006GL026587.

De Vivo, B., G. Rolandi, P. B. Gans, A. Calvert, W. A. Bohrson, F. J. Spera, and H. E. Belkin (2001), New constraints on the pyroclastic eruptive history of the Campanian volcanic plain (Italy), *Mineral. Petrol.*, *73*, 47–65.

Dingwell, D. B., and S. L. Webb (1989), Structural relaxation in silicate melts and non-newtonian melt rheology in geologic processes, *Phys. Chem. Minerals*, *16*, 508–516.

Dingwell, D. B., and S. L. Webb (1990), Relaxation in silicate melts, *Eur. J. Mineral.*, *2*, 427–449.

Dvorak, J. J., and G. Mastrolorenzo (1991), The mechanism of recent vertical crustal movements in Campi Flegrei caldera, southern Italy, *Spec. Pap. Geol. Soc. Am.*, *263*, 47 pp.

Dvorkin, J., M. Prasad, A. Sakai, and D. Lavoie (1999), Elasticity of marine sediments: Rock physics modeling, *Geophys. Res. Lett.*, *26*, 1781–1784.

Ferrucci, F., A. Hirn, G. De Natale, J. Virieux, and L. Mirabile (1992), P-SV conversions at a shallow boundary beneath Campi Flegrei caldera (Italy): Evidence for the magma chamber, *J. Geophys. Res.*, *97*, 15,351–15,359.

Gudmundsson, O., B. Brandsdóttir, W. Menke, and G. E. Sigvaldason (1994), The crustal magma chamber of the Katla volcano in south Iceland revealed by 2-D seismic undershooting, *Geophys. J. Int.*, *119*, 277–296.

Henrys, S. A., D. J. Woodward, D. Okaya, and J. Yu (2004), Mapping the Moho beneath the southern Alps continent-continent collision, New Zealand, using wide-angle reflections, *Geophys. Res. Lett.*, *31*, L17602, doi:10.1029/2004GL020561.

Husen, S., R. B. Smith, and G. P. Waite (2004), Evidence for gas and magmatic sources beneath the Yellowstone volcanic field from seismic tomographic imaging, *J. Volcanol. Geotherm. Res.*, *131*, 397–410.

James, M. R., N. Bagdassarov, K. Müller, and H. Pinkerton (2004), Viscoelastic behaviour of basaltic lavas, *J. Volcanol. Geotherm. Res.*, *132*, 99–113.

Judenherc, S., and A. Zollo (2004), The Bay of Naples (southern Italy): Constraints on the volcanic structures inferred from a dense seismic survey, *J. Geophys. Res.*, *109*, B10312, doi:10.1029/2003JB002876.

Kress, V. C., Q. Williams, and I. S. E. Carmichael (1988), Ultrasonic investigation of melt in the system Na_2O-Al_2O_3-SiO_2, *Geochim. Cosmochim. Acta*, *52*, 283–293.

Mangiacapra, A., M. Rutherford, and L. Civetta (2007), Pre-eruption conditions of Minopoli2 shoshonitic magma from melt inclusions and experimental studies, *Geophys. Res. Abstr.*, *9*, Abstract 05747. European Geosciences Union.

Mavko, G., T. Mukerji, and J. Dvorkin (1998), *The Rock Physics Handbook: Tools for Seismic Analysis in Porous Media*, Cambridge Univ. Press, Cambridge, U.K.

Murase, T., and A. R. McBirney (1973), Properties of some common igneous rocks and their melts at high temperature, *Geol. Soc. Am. Bull.*, *84, 3563–3592.*

Rivers, M. L., and I. S. E. Carmichael (1987), Ultrasonic studies of silicate melts, *J. Geophys. Res.*, *92*, 9247–9270.

Rosi, M., and A. Sbrana (1987), *Phlegraean Fields*, *Quad. de La Ric. Sci.*, CNR Ed.. Rome.

Singh, S. C., G. M. Kent, J. S. Collier, A. J. Harding, and J. A. Orcutt (1998), Melt to mush variations in crustal magma properties along the ridge crest at the southern East Pacific Rise, *Nature*, *394*, 874–878.

Sorey, M. L., W. C. Evans, B. M. Kennedy, C. D. Farrar, L. J. Hainsworth, and B. Hausback (1998), Carbon dioxide and helium emissions from a reservoir of magmatic gas beneath Mammoth Mountain, California, *J. Geophys. Res.*, *103*, 15,303–15,323.

Taylor, M. A. J., and S. C. Singh (2002), Composition and microstructure of magma bodies from effective medium theory, *Geophys. J. Int.*, *149*, 15–21.

Vanorio, T., J. Virieux, P. Capuano, and G. Russo (2005), Three-dimensional seismic tomography from P wave and S wave microearthquake travel times and rock physics characterization of the Campi Flegrei caldera, *J. Geophys. Res.*, *110*, B03201, doi:10.1029/2004JB003102.

Zandt, G., M. Leidig, J. Chmielowski, D. Baumont, and X. Yuan (2003), Seismic detection and characterization of the Altiplano-Puna magma body, central Andes, *Pure Appl. Geophys.*, *160*, 789–807.

Zollo, A., et al. (2003), Evidence for the buried rim of Campi Flegrei caldera from 3-D active seismic imaging, *Geophys. Res. Lett.*, *30*(19), 2002, doi:10.1029/2003GL018173.

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