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The Campi Flegrei caldera unrest: Discriminating magma intrusions from hydrothermal effects and implications for possible evolution

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A B S T R A C T

The Campi Flegrei caldera in Southern Italy is one of the most populated active volcanoes on Earth. It has an unprecedented record of historical unrest and eruption that dates back to 2.2 ka BP and provides key insights for understanding the dynamic evolution of large calderas. Since 1950, it has undergone four episodes of caldera-wide uplift and seismicity, which have raised the coastal town of Pozzuoli, near the centre of unrest, up to 4.5 m and triggered the repeated evacuation of some 40,000 people. After about 20 years of subsidence, following the uplift peak reached in 1984, the caldera started a new, low rate uplift episode, accompanied by low magnitude seismicity and marked geochemical changes in fumaroles. In this area it is crucial to discriminate episodes of shallow magma intrusion from hydrothermal perturbations, which are both able to generate unrest signals. In this paper, by a critical review of previous literature and some new results, we discriminate, in the unrest episodes, the relative contributions of hydrothermal effects and shallow magma intrusions. Our review is aimed also to show the different behavior of the largest unrest episodes, such as the 1982–1984, and the present, ungoing unrest characterized by smaller rate but longer lasting uplift. We show that for the former, larger uplift of the 80's, there is clear evidence for shallow magma intrusion, and we are able to compute the amount of intruded magma volume. For the present, on-going uplift, on the contrary, there is no evidence for magmatic activity at shallow depth. As a main result of our analysis, we demonstrate here the present unrest, characterized by much lower uplift rates and seismicity, is only interpretable as due to large gas fluxes coming from the deeper magma reservoir; without any appreciable contribution from shallow magma or recent magmatic intrusion. Our results shed new light on the interpretation of caldera unrest worldwide, and clearly indicate the most constraining data and the most rigorous procedures of data analysis for a correct interpretation of volcanic unrest.

1. Introduction

Volcanic eruptions are generally preceded by unrest episodes, but, mainly in calderas, many unrest episodes are not followed, in the short/medium term, by eruptions (see also Newhall and Dzurisin, 1988 and Acocella et al., 2015, for a thorough discussion). Large collapse calderas are the most explosive volcanic areas on the Earth, able to cause global catastrophes (de Silva et al., 2015; Oppenheimer, 2002). After caldera formation, periods of caldera resurgence occur, characterized by unrest episodes (de Silva et al., 2015) in which uplift and subsidence can alternate, with of without eruptions, driven by complex mechanisms involving magma, geothermal system, tectonic stresses. Several unrest episodes at calderas involve up and down ground displacements, not interbedded by eruptions as it normally occurs at other kinds of volcanoes (e.g. De Natale et al., 2006a,b; Acocella et al. 2015). De Natale et al. (2001, 2006a,b) argued the up and down ground displacements, often observed at calderas, call for the involvement of hydrothermal perturbations in the shallow aquifer, which add up to the effects of possible shallow magma intrusions. Discriminating shallow magma intrusions from hydrothermal effects, during unrest episodes at calderas, is then a key argument to improve eruption forecast and, more in general, our understanding of caldera unrest mechanisms. Campi Flegrei caldera, in Southern Italy, is the volcanic area in which such observations have been most evident in recent decades, and is hence an ideal site to afford this basic problem in volcanology.

Campi Flegrei caldera (Fig. 1) is a volcanic area containing part of the city of Naples, likely involved in the Campanian Ignimbrite eruption (39,000 years BP) and giving rise to the large, caldera forming Neapolitan Yellow Tuff eruption (15,000 years BP) (Rosi and Sbrana, 1987; Deino et al., 2004; Fedele et al., 2008; De Natale et al., 2016). The complex volcanological and tectonic history of the area has produced an intricate interplay of fault systems dissecting the caldera at surface and marking the boundaries between high-density and low-density bodies, with the latter including an intensely onshore fractured zone surrounding the Pozzuoli town and extending to about 2000 m depth and prone to localized seismicity (De Natale and Zollo, 1986; Troise et al., 2003; Capuano et al., 2013). The last eruption in the area occurred in 1538 AD, preceded by a large uplift episode, totaling > 17 m of uplift started about 100 years before the eruption (Parascandola, 1947; Dvorak and Mastrolorenzo, 1991; Morhange et al., 2006; Di Vito

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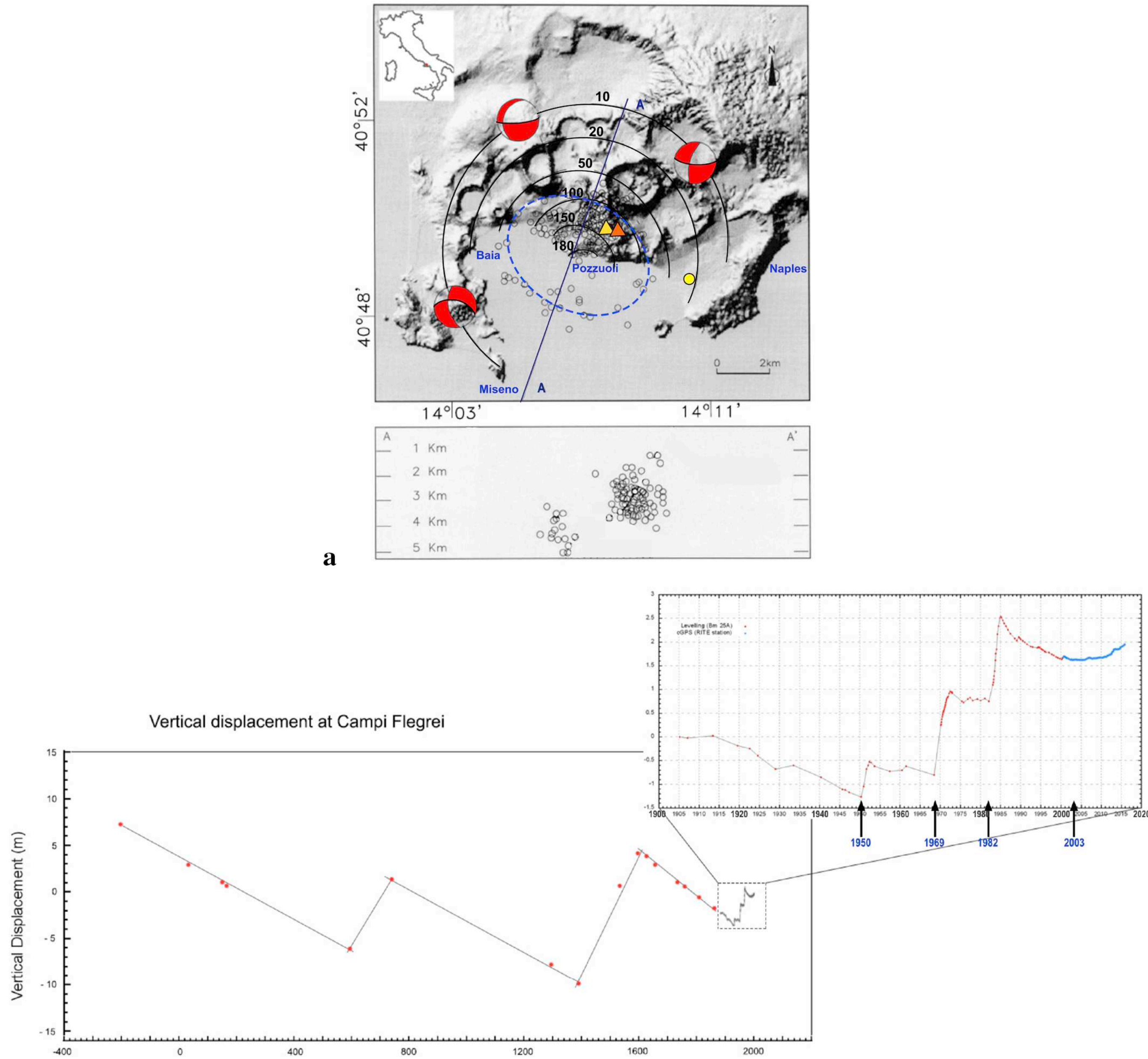

Fig. 1. a) Map of geophysical observations at Campi Flegrei during the 1982–1984 unrest episode. The bottom part of the figure shows the earthquake projections on the section AA′. Contours of vertical displacements (in cm) and earthquake epicentres are shown with black lines and circles. Composite focal mechanisms computed for the different seismic zones are indicated (in red; preferred fault planes in bold blue) (De Natale et al., 1995). Also indicated are the main fumarole areas: Solfatara (yellow triangle) and Pisciarelli (orange triangle), where geochemical analyses in the dataset come from. The yellow circle indicates the location of the pilot hole drilled in 2012 in the framework of Campi Flegrei Deep Drilling Project. The blue dashed line roughly indicates the resurgent area, in which the ground deformation is approximately confined (De Natale et al., 1997; De Natale et al., 2006a, b) Secular ground displacement at Campi Flegrei caldera, measured from marine molluscs borings on the columns of Serapis Temple in Pozzuoli (until 1905), from precision levellings at Pozzuoli harbor (from 1905 to 2000) and from GPS at Rione Terra after 2000. The smaller inset puts in evidence the displacements in the modern era, from 1905 to present (Del Gaudio et al., 2010). Also evidenced, on the inset, are the years of start of the last three large unrest episodes (1950–1952, 1969–1972, 1982–1984) and the year which appears to separate the subsidence started after 1984 and the beginning of the new, on-going slow uplift episode. (For interpretation of the references to colour in this figure legend, the reader is referred to the web version of this article.)

et al., 2016). The area has undergone, in the last 2000 years, large up and down movements, with subsidence dominating at a secular scale (with average rate of 1.7 cm/year) and, in particular, after the 1538 AD eruption (De Natale et al., 2001; De Natale et al., 2006a; Del Gaudio et al., 2010; De Natale et al., 2017; Moretti et al., 2017, 2018) (Fig. 1b). However, starting since the 50's of the last century, the ground movements in the area reversed to uplift, which, since 1950 to 1984 totaled > 4 m with peak rates, in 1983–1984, of about 1 m/year. After about 20 years of relatively fast subsidence following the 1984 peak of vertical ground displacement, uplift started again around year 2004 (Fig. 1b), at rates comparable to those of subsidence (on average 4 cm/year), but much lower than previous uplifts. Both the post-1984 subsidence and the subsequent and still ongoing uplift phase, show minor peaks of uplift followed by a fast recovery of the whole uplift (the so-called 'mini-uplift' episodes: Gaeta et al., 2003; Troise et al., 2007) (Fig. 1b).

The scientific interest and the civil protection concern, for such macroscopic unrest episodes in a very populated area, have stimulated multi-disciplinary research, which allowed to refine our knowledge about the volcanism in this area. In particular, it is now well known that the magma feeding system consists of a permanent, large molten body in the form of a wide (> 200 km^2) and thin (about 1 km) sill, located in the 7–10 km depth range and extending below the whole Neapolitan volcanic area, which includes Vesuvius, Campi Flegrei and, probably, Ischia volcanoes (Zollo et al., 1996, 2008).

At Campi Flegrei, occasional shallow magma accumulation (at about 3–4 km of depth) can occur, as a consequence of magma intrusions from the deeper source. An increasing number of geophysical and geochemical studies corroborate such a two-reservoirs model (De Natale et al., 2006a; Mangiacapra et al., 2008; Arienzo et al., 2010; De Siena et al., 2010; Aiuppa et al., 2013; Moretti et al., 2013a,b; Mormone et al., 2011; Esposito et al., 2011; Zollo et al., 2008; Trasatti et al., 2011), which was adopted as a reference frame to identify the processes behind the present CFc activity, including unrest (e.g., Aiuppa et al., 2013; Moretti et al., 2013a). Very recent 3D numerical modelling of the CFc thermal state (Di Renzo et al., 2016) also confirm that the magmatic system in the last 15 ka is recurrently characterized by the activation of such two reservoirs. In addition, the distribution of 3D gravimetric anomalies confirms the recurrent occurrence of small and shallow high-density bodies consistent with intrusions along faults periphery and the main structural lineaments, which have been preferential pathways for migration of aqueous fluids, gases, and magmas (Capuano et al., 2013). The distribution of anomalous bodies agrees well with previous and independent results from surface wave tomography (Guidarelli et al., 2002, 2006). The presence of deep magma reservoirs of regional extension, fed by mantle-derived fluids and magmas infiltrating from mantle depths is consistent with previous geochemical-petrologic hypotheses (e.g., De Vivo et al., 2010; Pappalardo and Mastrolorenzo, 2012; Moretti et al., 2013b) and with strong geophysical evidences of a low-velocity crustal layer at around 7–10 km of depth (Zollo et al., 1996; De Natale et al., 2001, 2006b; Guidarelli et al., 2006; Zollo et al., 2008; Nunziata et al., 2006). This thin, wide spread magma layer is connected with deeper magma roots at a regional scale (De Gori et al., 2001; Costanzo and Nunziata, 2017). In particular, below the Campi Flegrei district, these latter authors inferred the presence of a low shear-wave seismic layer extending from the depth of 11–12 km down to the Moho discontinuity (km of depth), consistent with the presence of a reservoir fed from a deep source in the upper mantle, from which the pockets of magma may rise. Body-wave seismic tomography reveals a continuous high velocity subducting slab beneath the Campanian volcanoes (De Gori et al., 2001; Panza et al., 2007), which is subject to a progressive detachment southward and explains the slab enriched mantle source of the Campanian volcanism in agreement with the complex features of the lithosphere-asthenosphere system along the Tyrrhenian margin (Panza et al., 2007).

The intrusion of magma at shallow depth during recent uplift episodes has been largely hypothesized, starting from the '70s and 80's (e.g. Corrado et al., 1977; Berrino et al., 1984; Dvorak and Berrino, 1991; Ferrucci et al., 1992; Troise et al., 2007). Ferrucci et al. (1992) found evidence for a sharp rigidity contrast at 3.5–4 km of depth, interpreted as the top of a sill-like intrusion, whereas De Siena et al. (2010) reported some indirect interpretation, from attenuation tomography, of a small melt-rock patch below 3 km. Besides these two papers, no direct evidence for shallow magma intrusion has been found. On the contrary, results of a large and detailed seismic tomography experiment carried out in 2000 (SERAPIS, see Judenherc and Zollo, 2004; Battaglia et al., 2008; Zollo et al., 2008) pointed out the absence of significant amounts of melt at shallow depth (i.e. within the resolution limit of 1 km-size patches), yet inferring a wide, thin magma layer located at about 7.5 km of depth (Zollo et al., 2008). Several recent papers, however, ascribe the on-going uplift to repeated injections of magma at shallow depth, in particular during the peak uplift rate of 2012–2013 (D'Auria et al., 2015; Chiodini et al., 2017; Giudicepietro et al., 2017). More specifically, Macedonio et al. (2014) proposed a model in which sill-like magma intrusion and lateral expansion at the discontinuity between two layers can generate uplift at surface, followed by subsidence when magma stops intruding and spreads laterally. Such a model was claimed to explain the sequences uplift/subsidence recorded at the Pozzuoli bay, associated both to the peak uplift episodes (in particular during the 1982–2004 period), and to the mini-uplifts (see Giudicepietro et al., 2017). However, Troiano et al. (2011) demonstrated that the inflow of deep, hot magmatic fluids in the shallow aquifer, can explain well both uplift and subsequent subsidence, the latter due to the outflow of pressurized gases from the caldera more permeable rocks. Furthermore, Moretti et al. (2018) used a thorough thermodynamic assessment of fumarolic gases to show how to discrimate whether the observed warming up of the hydrothermal system is due to magma intrusion or infiltration of deep CO_{2-}-rich gases inducing vaporization of the hydrothermal bottom zone. In principle, discriminating magma intrusion from hydrothermal effects in the uplift episodes at Campi Flegrei, is crucial to understand how calderas work worldwide, and even more for a reliable assessment of eruption hazard in the area. The volcanological literature of the last 50 years, however, is plenty of contrasting interpretations, either in terms of shallow magma intrusion or of hydrothermal perturbations related to injection of deep magmatic gases. This paper aims to critically review the recent literature, at least the most focused one to solve this important question, and present some new data and analyses, in order to give a more convincing answer to a problem involving the forefront volcanological research as well as crucial arguments of civil protection.

In order to do this we will review recent interpretations of Campi Flegrei unrest, and in particular some models which have been used, in recent literature, to support the hypothesis of multiple magma intrusions at shallow depths. Once we critically discuss the pitfalls of these recent models, we review the interpretation of geochemical and geophysical data (in particular ground deformations) showing what are the really robust evidence for shallow magma intrusion, and what is the likely effect of hydrothermal perturbations due to injection of deep fluids into the shallow aquifers, in different unrest periods. We will consider both the 1982–1984 large unrest, characterized by very high uplift rate (up to 1 m/year) and followed by later subsidence till 2000–2003, and the post-2006 uplift, still ungoing at a rate much smaller than '80s episode. In order to separate, in the 1982–2003 ground deformation, the effects of shallow magma intrusion from the effects of deep fluid (non magmatic) injection, we use theoretical modelling using a permeability model inferred on peculiar leak-off tests in the Campi Flegrei Deep Drilling Project pilot hole well (De Natale and Troise, 2011; Carlino et al., 2014; De Natale et al., 2016), integrated by measurements obtained in the deeper part of the aquifers (3 km) by AGIP-ENEL geothermal exploration in the '70s and '80s (AGIP, 1987).

As a final result, we will give insight into the main causes of unrest phenomena at each period, and in particular, starting from 1980, about when shallow magma intrusions have occurred and when the unrest must be likely interpreted in terms of hydrothermal perturbations.

1.1. The Campi Flegrei unrest episodes: hydrothermal vs magmatic effects

Studies of slow movements at Campi Flegrei began with observations of sea-level markers on Roman coastal ruins, which were sensitive to large, secular deformation (Breislak 1792; Forbes 1829; Niccolini 1839, 1845; Babbage 1847; Lyell 1872; Gunther 1903). Parascandola (1947) presented the first reconstruction of historical ground movements, later modified by Dvorak & Mastrolorenzo (1991), and, more recently, by Morhange et al. (1999) and Del Gaudio et al. (2010). Ground deformation occurred at Campi Flegrei at different time scales (Fig. 1b) and was recorded in different ways (secular, from marine ingression levels on ancient buildings; last century, by precision levellings; last 18 years, by GPS). The Pozzuoli area has been characterized, since Roman times at least, by subsidence at a rate of about 1.5–2.0 cm/year, interrupted three times at most: perhaps in the Middle Ages (following the view of Morhange et al. 1999); certainly about 40 to 100 years before the last eruption in 1538 and since 1950 to present (Del Gaudio et al., 2010) and finally since 1950 to present. After the 1538 eruption, in fact, subsidence started again at the previous rate, until the 1950, when fast uplift started again and led to three peak episodes: 1950–1952, 1969–1972 and 1982–1984. The total uplift since 1950 to 1984 was > 4 m. The most studied unrest was the 1982–1984 one, when 1.86 m of uplift were recorded (e.g. De Natale et al., 2006a). The 1982–1984 unrest was also accompanied by seismic swarms of low magnitude (M_{max} = 4.0), whose frequency reached very high values, with a peak of 610 events in few hours on April 1st 1984. After 1984, about 20 years of subsidence followed, at a generally decreasing rate, with an average rate of about 4 cm/year (i.e. ~80 cm of total subsidence; Fig. 1b). The last two episodes of unrest gave rise to concern about a possible impending eruption. The Rione Terra of Pozzuoli town, at the most deformed part of the caldera, was evacuated in 1970, and the whole town of Pozzuoli (about 40,000 people) was evacuated at the end of 1983 (Barberi et al. 1984). Moretti et al. (2018) first showed the striking symmetry between the main 1984–2003 subsidence trend, and the following and still ongoing uplift (Fig. 2), both characterized by the same but reversed exponential dependence.

Superimposed to such subsidence and the subsequent uplifts pattern, several sharp and short uplift episodes occurred, with peaks in 1989, 1994, 2000, 2006 and 2012–2013 (Gaeta et al., 2003; Troise et al., 2007; D'Auria et al., 2015) (Fig. 2). Mini-uplifts (so called after Gaeta et al., 2003) have then occurred after 1984 with a very regular recurrence time of 5–6 years, amplitudes of 4 to > 10 cm, and a duration of few months each.

The last mini-uplift episode occurred in 2012–2013 (see Fig. 2), but it has been not recognized as such till now. In fact, it added up to an uplift pattern and then it ended with a flat or only slightly subsiding trend, whereas the other ones were superimposed to a decreasing trend (e.g., 1989 and 1994 mini-uplifts), or to a nearly flat background (2001 episode) and only once to a slight initial uplift (e.g., the 2006 mini-uplift, the first one with only partial recovery of the uplift) In all the

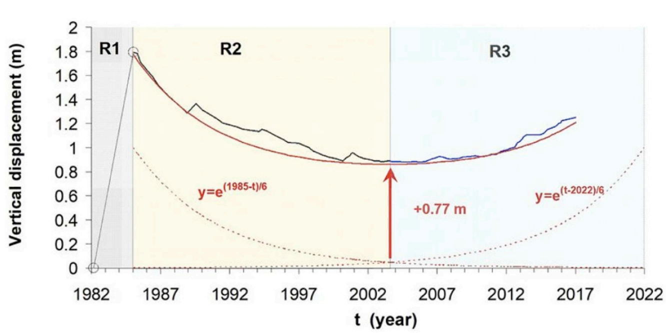

Fig. 2. Baseline displacement (red-solid line) involving the post-1984 subsidence and the ongoing unrest (the zero is set at the beginning of the 1982–84 uplift). Observed vertical displacement at the point of maximum deformation (port of Pozzuoli) is shown by the blue line. Error bars on measured displacements are ± 4 mm for leveling (until year 2000) and ± 3 mm for continuous GPS (after year 2000; from INGV-OV surveillance bulletins: http://www.ov.ingv.it/ov/en/campi-flegrei/275.html). Three deformation regions are distinguished (R1, R2 and R3). Note the specular symmetry of the subsidence phase (post-1984) and the new, on-going uplift phase. They can be both fitted by the same exponential functional form, with the exponent sign changed (after Moretti et al., 2018). (For interpretation of the references to colour in this figure legend, the reader is referred to the web version of this article.)

previous episodes (besides the 2006), after the peak uplift the subsequent subsidence recovered almost completely the previous uplift, making it difficult to recognize the 2012–2013 increased uplift rate as a similar episode. Some researchers, in fact, interpreted the increase in uplift rate during the 2012–2013 as an episode of magma intrusion at shallow depth (D'Auria et al., 2015; Chiodini et al., 2017). The temporary, sharp increase of uplift rate likely affected the decision of Civil Protection authorities who, on December 2012, declared the 'Yellow' alert level (the first step after the 'Green' or quiescent one). The pattern of ground deformation during both the recent uplift and subsidence episodes resembled a bell shape, centered at Pozzuoli harbor (Fig. 3a and b). Another key observation, valid for both ground uplift and subsidence at Campi Flegrei, is the remarkably constant shape of ground deformation, which has remained practically unchanged in the last 48 years, at least (i.e. since the first vertical levellings in the seventies), during both up and down deformation, with very different amount (De Natale and Pingue, 1993; De Natale et al., 1997; Gottsmann et al., 2006). Ground deformation at this area shows a very sharp deformation decay from the most deformed point (Pozzuoli harbour, inset of Fig. 3a), decreasing to about 30% of maximum uplift at 3 km of distance (Fig. 3b). The striking constancy of the deformed area is accompanied by the same remarkable constancy of the seismogenic volume; both the deformed area and the seismic volume do not enlarge, and remain confined during the whole unrest.

Since the '70s of the last century, when the first recent unrest in the area was observed and measured, two main lines of explanation for such phenomena have been proposed. The first, and most obvious one in a volcanic area, relied on the effect of magma intrusion at shallow depth (Corrado et al., 1977; Berrino et al., 1984; Bianchi et al., 1987; Dvorak and Berrino, 1991; Bellucci et al., 2006; Amoruso and Crescentini, 2011; D'Auria et al., 2015; Giudicepietro et al., 2017). The second line of interpretation was based on hydrothermal effects, without any contribution from direct magma rising (Casertano et al., 1976; De Natale et al., 1991; Bonafede and Mazzanti, 1997; Gaeta et al., 1998; Todesco et al., 2003; Bodnar et al., 2007; Troiano et al., 2011). Between these two diverging interpretations, several papers, mainly in recent times, proposed a mixed magmatic-hydrothermal models to produce the unrest episodes (De Natale et al., 2001, 2006a; Gottsmann et al., 2006; Battaglia et al., 2006; Troise et al., 2007; De Natale et al., 2017). The main problem encountered when trying to explain the whole pattern of ground deformation observed at Campi Flegrei with a purely magmatic model (i.e. only with magma intrusion at shallow depth) was to explain the observed long lasting subsidence following the peak uplift of 1984, as well as the sharp subsidence observed after the small uplift episodes occurred in 1989, 1994, 2000, 2006 (the so called 'mini-uplift', after Gaeta et al., 2003). In fact, once the magma is intruded at shallow depth, it is not realistic to imagine it may return back to higher depths causing the observed deflation (De Natale et al., 1991, 2001). This problem, first evidenced by De Natale et al. (1991) has been long recognized as as a serious one, then favouring the interpreations based on purely hydrothermal or mixed magmatic-hydrothermal ones. However, Macedonio et al. (2014) claimed that a magma sill fed by liquid magma from below could explain both the uplift phase and the subsequent subsidence. In favour of purely hydrothermal models came also, in the first years of 2000, the results from the reflection tomography Project Serapis (Judenherc and Zollo, 2004), which were able to find a large magma sill located at about 8 km, probably underlying the whole Neapolitan volcanic area (Zollo et al., 1996), but no evidence for shallow magma batches, in the depth range (3–4 km) required to explain the observed ground deformation during unrest. These results seemed to contrast with findings of Ferrucci et al. (1992) who found evidence of converted P-SV phases at about 3–4 km of depth, recorded at the Pozzuoli harbor seismic station, from a deep Sothern Thyrrenian subduction earthquake, as well as from shots of an active seismic experiment in the Gulf of Pozzuoli. In the last decade, however, the problem of discriminating magmatic from hydrothermal

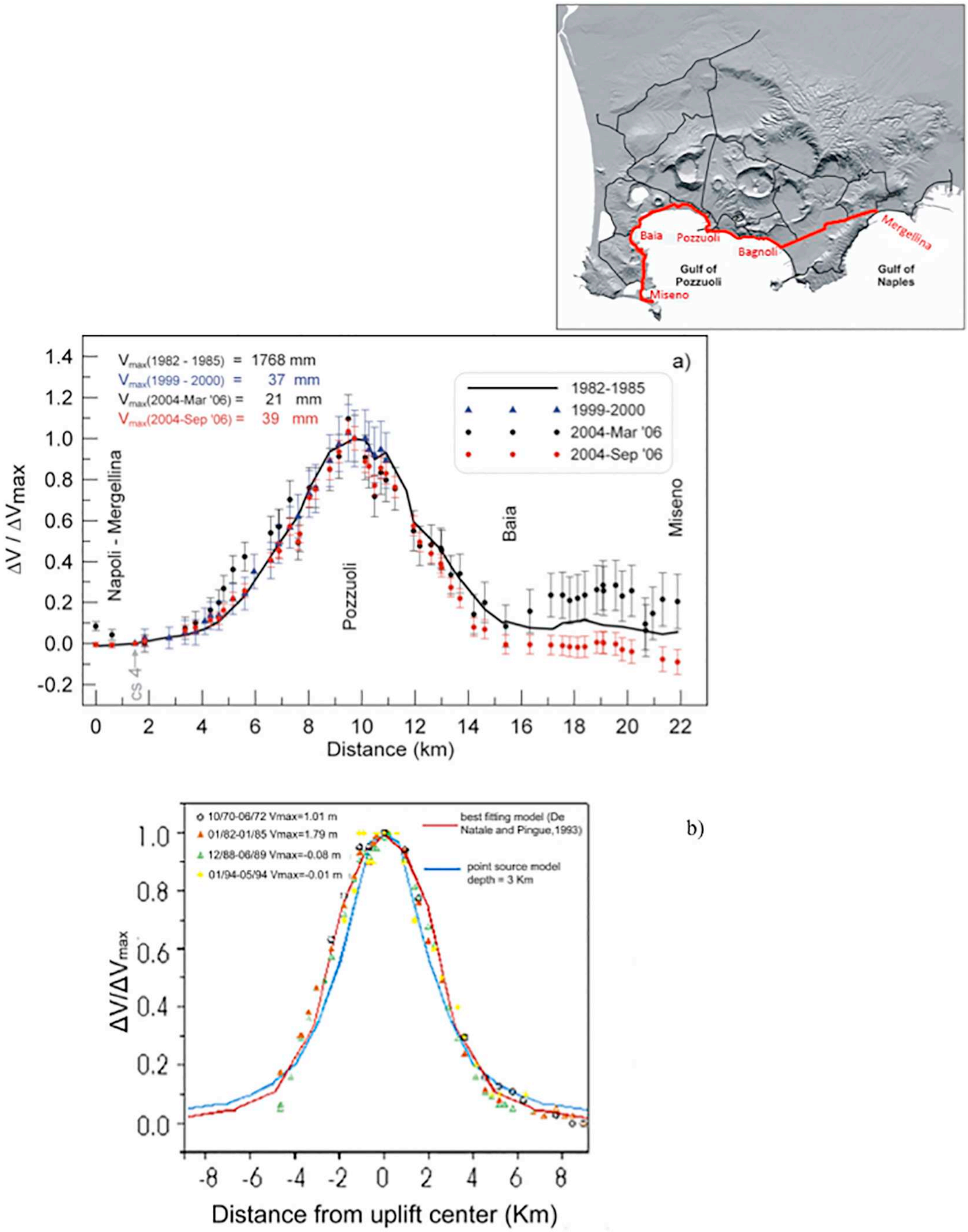

Fig. 3. a) Normalized absolute value of ground deformation measured by precision levellings, along the main leveling route from Naples to Miseno (red line), at several periods from 1982 to 2006; b) Normalized absolute value of ground displacement data, computed at several periods and shown as a function of the distance from the maximum deformed point (Pozzuoli harbor) (De Natale and Pingue, 1993). The maximum deformation (positive or negative) for each period is indicated. Note the remarkable constancy of the shape of ground deformation pattern, for strongly variable amounts of total displacement. (For interpretation of the references to colour in this figure legend, the reader is referred to the web version of this article.)

effects has been mainly afforded by the use of geochemical observations. These observations, in the first years of 2000 were interpreted as due to deeper gas injection in shallower aquifers (Chiodini et al. 2003); however, in last decade, the same group of authors interpreted the new uplift episode, started around 2005–2006, as mainly due to consecutive magma intrusions at shallow depth (Chiodini et al., 2017 and references therein; D'Auria et al., 2015; Giudicepietro et al., 2017) producing the gas composition discharged from surface fumarolic vents. In the last years, the magmatic interpretation has been dominant, with some exceptions (Troiano et al., 2011; Moretti et al., 2013a, 2017, 2018).

In the next part of this paper, we first review what can be inferred from a rigorous analysis of the extremely rich geochemical data set, dating back 37 years. Then, we will try to quantify the magma volumes involved in the shallow intrusions episodes. Finally, we will further focus on the problem of the interpretation of the ongoing unrest, by critically reviewing the recent literature on this argument.

1.1.1. *Geochemical data interpretation*

Geochemical data have been collected, with good frequency and continuity, since 1982. This 36 years long data set then constitutes a very rich source of information on unrest phenomena, mainly because such a period encompasses the large uplift episodes of 1982–1984, the subsequent subsidence 1984–2003, and the on-going uplift. Several papers have been devoted to the analysis of geochemical data since the 80's. We recall here some of them that are the most interesting for our goal of discriminating magmatic from hydrothermal unrest mechanisms. Tedesco et al. (1988) found no evidence for magmatic gas contribution to the unrest episodes; De Natale et al. (1991) first tried to propose a unified model by joint interpretation of geochemical and geophysical data, and also favoured an interpretation in terms of hydrothermal perturbation, mainly based on the observation that subsidence was in progress at that time. Starting from year 2001, several papers from a group of authors (Chiodini et al., 2016, 2017 and references therein; Todesco et al., 2003; Caliro et al., 2007) have been devoted to this problem. Earliest papers (Chiodini et al., 2003; Todesco et al., 2003; Caliro et al., 2007) pointed to the effect of deeper gas injection below the aquifer system, which would have produced both ground uplift and subsequent increased degassing outporing from surface. In more recent papers, however (Chiodini et al., 2015, 2016), the same authors ascribe the recent uplift to the activity of a shallow decompressing magma chamber (3–4 km), without disregarding the role of occasional shallow magma intrusion episodes (e.g., D'Auria et al., 2015) that would renew the whole process. Upon magma decompression, the huge release of magmatic steam into the hydrothermal system would produce in the latter an important pressure and thermal increase, yielding steam condensation (Vanorio et al., 2011 and references therein). Such a removal of steam by condensation is required in this model, because the gases sampled in the main fumaroles show decreasing H_2O but increasing CO_2 contents, which is not consistent with the exsolution of a shallow decompressing magma. Moretti et al. (2017, 2018), however, demonstrated that geochemical data are consistent with the presence of shallow magma intrusion (at 3–4 km of depth) only in 1982–1984. In that period, in fact, the H_2O/CO_2 ratio (Fig. 4a) of fumarole fluids showed a sharp peak, accompanied by the concomitant sharp increase of redox conditions (e.g., peak in H_2O/H_2 Fig. 4b; Moretti et al., 2017) and by the sharp peaks of H_2S/CO_2, N_2/CO_2 and N_2/He (Fig. 4c–e) determined by the close injection of magmatic gases. On the contrary, since 2000–2003 there is evidence (monotonous decrease of all ratios, see all panels of Fig. 4) for the extinction (likely due to magma solidification by cooling) of the shallow intrusive magma body (Woo and Kilburn, 2010; Moretti et al., 2013a,b; Moretti et al., 2018).

Moretti et al. (2018) further showed that the pressure and temperature conditions in the host rocks surrounding main ascent pathways of Campi Flegrei fumaroles are not allowing a bi-phase (liquid + vapor) mixture of water and carbon dioxide, but only a single gas phase. This single-gas phase rises to surface as a superheated vapor from the bottom of a hydrothermal system which is nowadays dried, i.e., supercritical. Therefore, no steam condense in the host rocks surrounding the rising gas plume, and the decreasing H_2O/CO_2 measured in fumarole emissions implies that the actively degassing magma must be a deep one (main reservoir located at about 8 km of depth), excluding the presence, since 2000–2003 to now, of any degassing shallow magma intrusion.

The main cause of the diverging interpretations of geochemical data by the two different research groups stands in the way pressure is treated in the geochemical analysis of data: Chiodini and coworkers make the 'a priori' assumption of liquid-vapor equilibrium, whereas Moretti et al. (2017, 2018) do not make any 'a priori' assumption, consider the system representative of a superheated vapor and use the whole analytical data set to compute P and T from chemical equilibria involving gas species. The arbitrary assumption of liquid-vapor equilibrium was supported by the interpretation of the results of the earliest numerical thermo-dynamical models elaborated by Chiodini et al. (2003) and Todesco et al. (2003) for the hydrothermal system surrounding the Solfatara fumaroles, on which the axisimmetric computational domain was ideally centered. According to geochemistry, such models give a single-gas zone phase only close to the bottom fluid injection point and in the very shallow portion, wehereas most of the

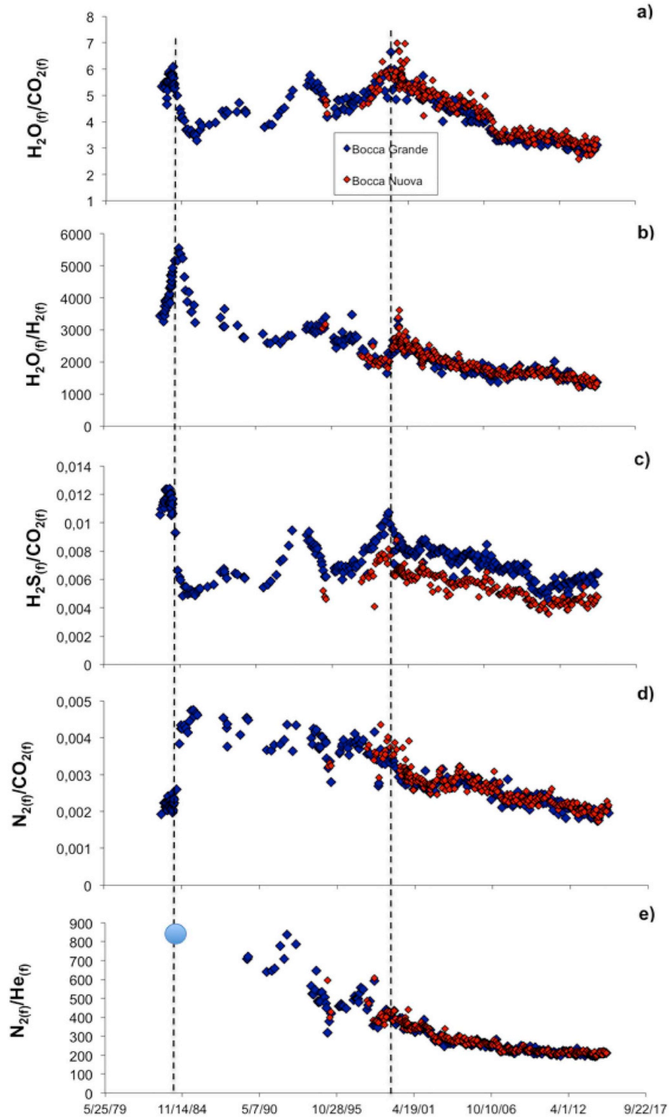

Fig. 4. Chronograms of relevant gas ratios measured at Solfatara crater fumaroles (named Bocca Grande and Bocca Nuova). All the selected ratios (a to e) display sharp peaks during the 1982–84 crisis and smaller ones in 1989, 1994 and 2000 (mini-uplift episodes). On the contrary, decreasing trends mark the ongoing unrest which started in 2000–2004. Note the sharp peaks of H_2O/CO_2 and H_2S/CO_2 around 2000, which likely marks the final stage of the shallow magma sheet cristallization process, characterized by large output of H_2O. The light blue circle in the panel e represents the value extrapolated by Chiodini et al. (2014) at the end of 1984. The time of maximum uplift and the time of likely cristallization of the shallow magma sheet are approximately indicate, on all panels, by the dashed lines. (For interpretation of the references to colour in this figure legend, the reader is referred to the web version of this article.)

rising plume is biphasic (Chiodini et al., 2003; Todesco et al., 2003; Caliro et al., 2007). We then carried out the same simulation, based on the same numerical program THOUGH2 (Pruess, 1991), with the same parameters (rock properties, fluid composition and enthalpy) reported in such studies. Fig. 5a exactly reproduces the above mentioned results on the hydrothermal plume constitution and show that these are obtained because a temperature of ~320 °C is computed at the plume bottom (1.5 km deep), in the sector where the injection of the CO_2-rich fluid occurs (Chiodini et al., 2003; Todesco et al., 2003; Caliro et al., 2007). The same authors, however, declared to have used the expected value for bottom temperature, which is 350 °C according to Chiodini et al. (2003) and Todesco et al. (2003) and which corresponds to the temperature of the H_2O-CO_2 fluid mixture indicated by

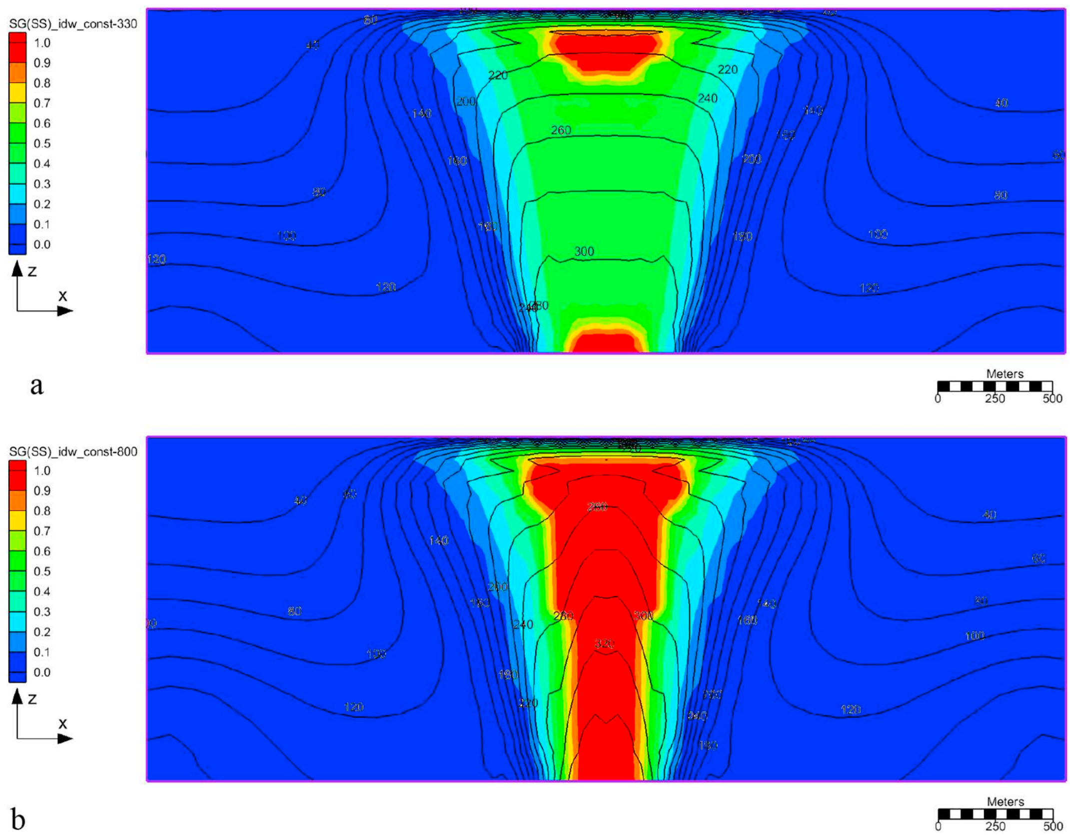

Fig. 5. Thermal model of the Solfatara aquifer. Central vertical (XZ) plane through the computational domain (X = Y = 15,000 m; Z = 1500 m, from which the shown domain 4000 m × 4000 m × 1500 m has been extracted) and results (Temperature in °C and gas fraction, SG) of THOUGH2 numerical simulations, for: a) bottom fluid enthalpy of 330 kJ/Kg, as in Chiodini et al. (2003) and Todesco et al. (2003); b) bottom fluid enthalpy of 800 kJ/Kg (this study). Note that, in a), the bottom temperature in the central part is 320°C, and not 350°C as claimed by the authors. It is, on the contrary, correctly imposed in b). The resulting gas fraction, shown in colour, is 1.0 (single phase) in the whole central column of b).

geothermometers (Caliro et al., 2007). This discrepancy between the claimed boundary condition at the model bottom and the resulting one, very evident in all figures of the thermal model published since 2003 to present, originates from the fact that the input enthalpy cannot correpond to the sum of pure and separate H_2O and CO_2 components (i.e., to an ideal mixture), because excess and positive enhtalpies arise from the interaction of H_2O and CO_2 in the real mixture (e.g. Moretti et al., 2018 and references therein). We then performed the same simulation by considering the excess enthalpy, which introduction allows obtaining the correct temperature of 350 °C in the injection area at the bottom of the computational domain (Fig. 5b). More important, the new output shows that the upgraded enthalpy determines a new structure of the ascending hydrothermal plume, now given by a single-gas phase from depth to surface extending in average 500 m around the central fumarolic pathway.

In agreement with the findings of Moretti et al. (2017, 2018), fumaroles are then hot control points locally tapping deep hydrothermal conditions, which can be related to the chemistry of gases emitted at surface without accounting for secondary effects such as steam condensation. On the other hand, deformation occur in a much larger volume, sourrounding the single-gas plume, and dominated by biphasic hot and/or boiling aquifers (so well below the critical point of water) that develop pore pressures and then ground displacement (Moretti et al., 2018).

1.2. Ground deformation modelling: constraints and interpretations

In this paragraph, we try to discriminate, from the total ground uplift recorded in the period 1982–1984, the amount of uplift likely due to shallow magma intrusion, whose main evidence comes from geochemical data as explained before, from that produced by hydrothermal effects. To this aim, we must recall that the hypothesis that a large part (or all) of uplift can be due to hydrothermal effects has been very frequently proposed in literature, since the 70's (Casertano et al., 1976). Such an hypothesis was given new strenght, after the 80's, by De Natale et al. (1991), and became the dominant model after the 'Serapis' tomography experiment (Zollo et al. 1996, 2008) did not find any trace of molten magma at shallow depths; a large magma sill was only inferred at 7.5–8.0 km of depth. The most of papers published between 2001 and 2012, interpreted Campi Flegrei uplift as exclusively or mainly due to perturbations of the hydrothermal system (De Natale et al., 2001; 2006a,b; Todesco et al., 2003; Chiodini et al., 2003; 2007; Lima et al., 2009). As already well explained by De Natale et al. (1991) the main observation calling for the effect of gas-water overpressure as the main cause of uplift was the large subsidence occurred after 1984. Such subsidence could be naturally explained by inflation-deflation of gas or of water, whereas it is almost impossible to explain it with a magma source, because the only way to release the magma pressure would be an eruption, since magma cannot outpore from the system given its high viscosity. However, some recent papers have proposed a model in which shallow magma injection can produce both uplift and subsequest subsidence. The original paper which claimed this interpretation is by Vanorio and Kanitpanyacharoen, 2015; D'Auria et al. (2015) used that model to ex-plain, in terms of shallow magma intrusion, the relatively higher rate uplift episode occurred in the period 2012–2013. The model developed by Macedonio et al. (2014) states that a certain amount of subsidence can occur even at the end of magma intrusion episodes, but however its background physics is not compatible with total uplift recovering.

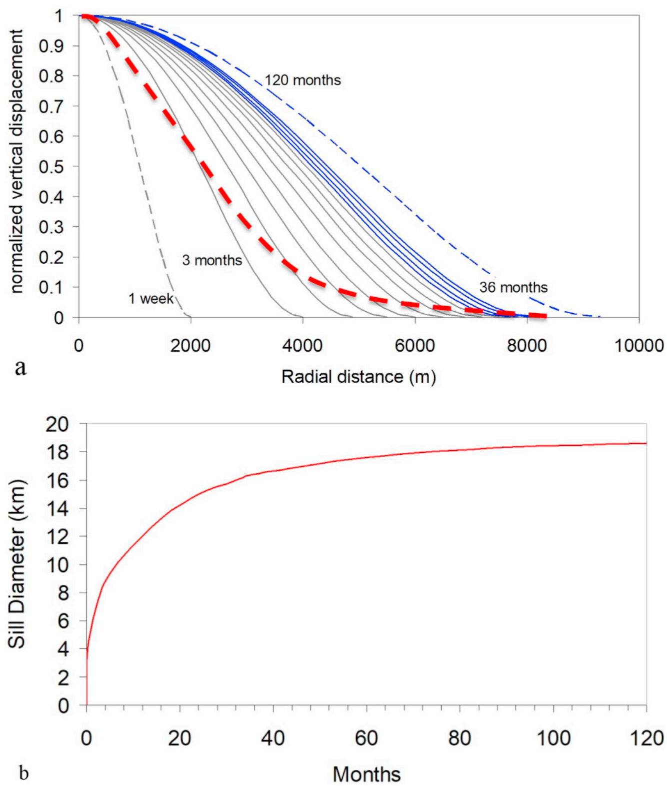

Fig. 6. a) Normalized vertical displacement versus radial distance inferred from the model of Macedonio et al. (2014). Solid lines are every 3 months, (between 3 and 36 months). Black lines correspond to the period of constant magma inflow (300 l/s), which stopped in correspondence of the rightmost solid black line (24 month). In this model, the displacement along the sill is equal to the displacement caused at surface. Note the large progressive lateral spreading of the sill, which would reflect in a specular enlargement of the deformed area at surface, with large changes of the shape of surface deformation; the red curve shows the observed radial decay of real vertical data, which is markedly constant during the time; b) computed sill diameter as a function of time. (For interpretation of the references to colour in this figure legend, the reader is referred to the web version of this article.)

Actually, the model by Macedonio et al. (2014) is not applicable to explain the uplift-subsidence patterns observed at Campi Flegrei nor the seismicity patterns during unrest, as we are going to demonstrate here.

Looking at the Fig. 3 in Macedonio et al. (2014), redrawn here as Fig. 6a, it appears very evident that, in such a model, the sill rapidly spreads laterally; when the magma injection at the center of intrusion stops, the injected magma accommodates by a further lateral expansion of the sill, so that the maximum uplift slightly decreases (subsidence). In the framework of this model, the surface displacement is equal to the sill thickness, shown in Fig. 6a as a function of distance from the surface projection of the sill center, at the times 0, 7 days, 97 days, 127 days, and so on, each 30 days until 1080 days, and then at 3600 days (almost ten years). The last two steps occur just after and well after the stop of magma injection. Fig. 6b shows the reconstructed time behavior of the sill diameter with the same parameters and time evolution used by Macedonio et al. (2014). It is clear that, following the lateral expansion of the sill, the shape of vertical deformation (but it occurs equivalently for the horizontal one) changes dramatically with time, with a strong enlargement of the deformed area. This happens because the model is not laterally constrained, so that magma injection is more easily accommodated by lateral expansion of the sill. This behavior is obviously due to the fact that the shallow layer is only leaned over the rigid half-space; so, there is not tensile strenght, and the magma can easily spread laterally by decreasing the pressure while the volume increases. For these reasons, the claims by Giudicepietro et al. (2017) that such a model can explain both uplift and subsidence observed during the Campi Flegrei unrest episodes, hold only regarding the time behavior of ground deformation above the center of sill-like intrusion, neglecting the strong changes of the ground deformation patterns. On the contrary, the clearest observation which characterizes all the ground deformation episodes at Campi Flegrei, since 1969 to present and whatever the maximum deformation amount and its sign (i.e. uplift or subsidence), is the remarkably constant shape of the ground deformation, at least the vertical one (see De Natale and Pingue 1993; De Natale et al., 1997; 2006a,b). This striking observation has been interpreted, by several authors (De Natale and Pingue, 1993; De Natale et al., 1997; Folch and Gottsmann, 2006), as due to the confining effect of ring faults, surrounding the caldera or the resurgent block (Beaducel et al., 2004). However, previous models show that the 'confining' effect of ring faults works only for pressure sources whose lateral extension is lower than the area spanned by the deeper part of the ring faults. On the contrary, pressure sources located below the ring fault system, and extending out of ring area, produce ground deformations much more similar to those occurring in an homogeneous, unfaulted medium. A progressively expanding sill like the one hypothesized by Giudicepietro et al. (2017) obviously would enlarge, after a short time, much more than the area enclosed by the ring faults, thus excluding any possible confining effect which would make more constant the ground deformation shape. Another compelling observation, ruling out the hypothesis of a significantly expanding sill, is the remarkable constancy of the seismic volume (Troise et al., 2003, see Fig. 1), which does not show any elargement, neither during the 1982–1984 uplift in which the ground level rised of about 1.8 m. Actually, a laterally expanding sill would produce, as also recognized by Macedonio et al. (2014) in their stress calculation, a large stress accumulation at the sill edges, thus producing a progressive lateral spreading of the seismicity, following the lateral expansion of the sill. All these considerations makes it evident that, at Campi Flegrei, it is not possible to hypothesize an intrusion mechanism involving the progressive lateral expansion of a sill. This evidence in turns excludes that subsidence following an uplift episode can be associated to magma intrusion.

1.3. Quantifying magma intrusion during 1982–1984 unrest

Once this statement, hidden by recent literature but here demonstrated to be valid for Campi Flegrei, is assumed, it is also simple to evaluate the maximum amount of magma which intruded shallow crust in the period 1982–1984. It should be in fact a volume able to produce the permanent ground uplift, following 1982, left after the subsidence 1985–2005. As it is clear from Fig. 1b, residual uplift was about 1.0–1.1 m. Such a residual uplift could be due either to residual injection of deep fluids from below, in a purely hydrothermal model, or to the original magma intrusion after the uplift due to the injection of deep fluids into the shallow aquifers has been recovered (see De Natale et al., 2001, 2006a; Troiano et al., 2011). Interestingly, the amount of uplift 1982–1984 recovered by the ground subsidence 1985–2005, as well as the rate of subsidence, is in good agreement with a thermal-fluid-dynamical model taking into account the measured, in-situ permeability of the Campi Flegrei substructure (De Natale et al., 2016) and the inferred rate of deep fluids injection during 1982–1984 in line with measured by gas flux emitted from the main fumaroles (Caliro et al., 2007 and references therein). As already demonstrated by Troiano et al. (2011) the models of ground deformation due to deep fluid injection are strongly affected by two critical parameters: in situ permeability and injection rate. While injection rate can be roughly estimated by measuring the flux of the main gases at the most important emission points, in situ permeability can only be measured by appropriate tests operated in drilled holes at the depths we want to characterize (AGIP, 1987; Zamora et al., 1994; Vanorio et al., 2002; Rabaute et al., 2003; Giberti et al., 2006; Vinciguerra et al., 2006; Peluso and Arienzo, 2007; Piochi et al., 2014). In our case, we use leak-off tests operated in the

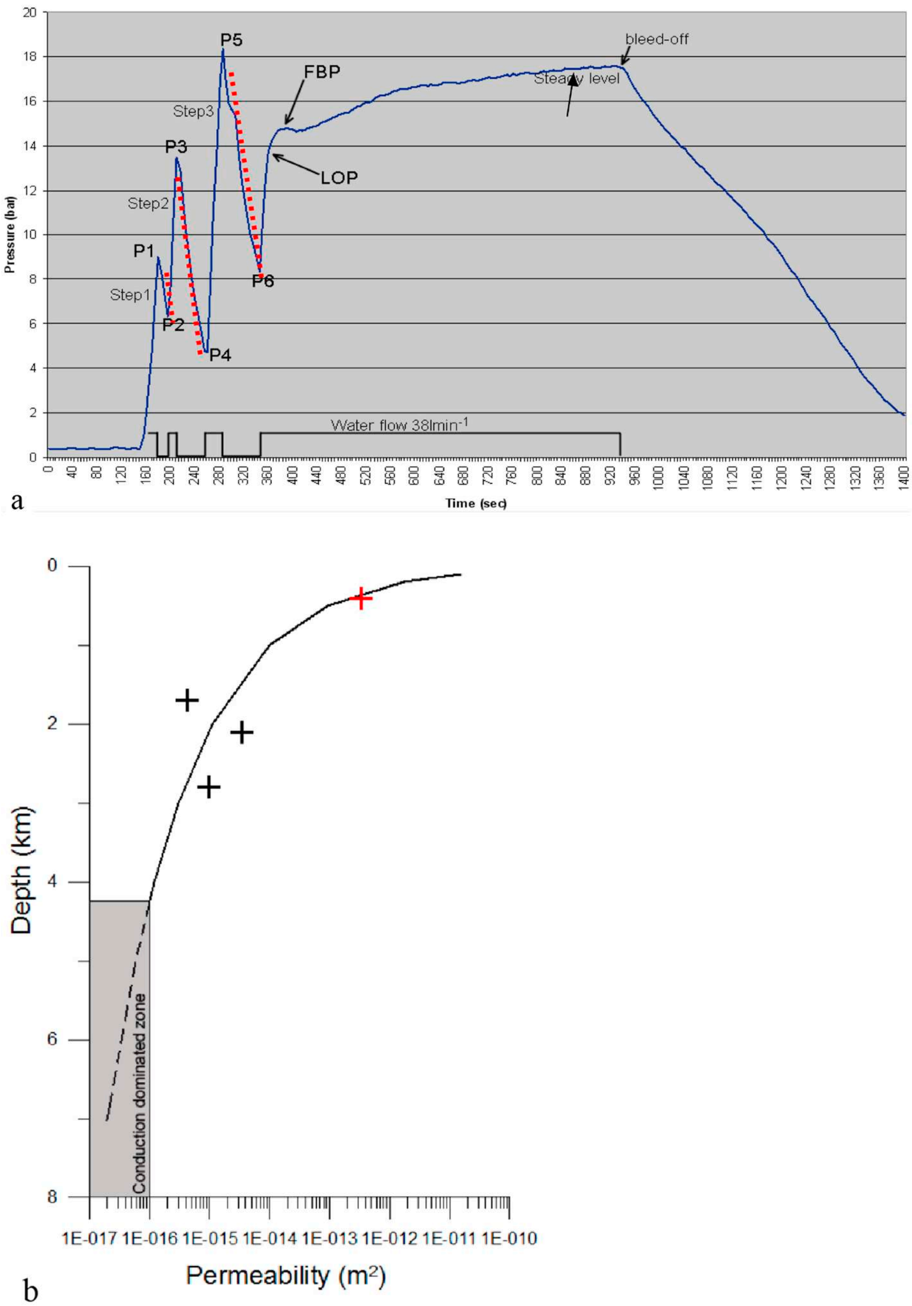

Fig. 7. a) Plot of the Leak-off test operated in the CFDDP pilot hole, at a depth of 500 m. The function in the lower part of the figure represents the water flow in the well. Differently from a normal Leak-off test, we made three runs, in the phase of pressure rise, by instantaneously start flow, taking it costant for some tens of seconds, and then suddenly stop it. The dashed red lines mark the pressure decays after the three runs of pumping and stopping, from which slope the permeability is computed. In this way, we were able to further determine the permeability of the surrounding rocks at 500 m., from the decay of pressure after stopping flow. The other points on the Leak-off curve, are: Leak-Off Pressure (LOP), Formation Beak-down Pressure (FBP), steady level and bleed-off (see Carlino et al., 2015, for an exhaustive explanation); b) plot of the theoretical permeability as a function of depth used in this work (from Manning and Ingebritsen, 1999). The best curve has been obtained by interpolating permeability data (shown by crosses) at different depths. The red cross indicates the permeability value obtained by CFDDP Leak-off tests (see Fig. 3), whereas black crosses indicate data obtained by ENEL-AGIP geothermal explorations in the ‘80s (AGIP, 1987). (For interpretation of the references to colour in this figure legend, the reader is referred to the web version of this article.)

Campi Flegrei Deep Drilling pilot hole experiment (De Natale and Troise, 2011; Carlino et al., 2015).

1.3.1. Permeability determination from in-situ experiments during CFDDP pilot hole drilling

Drilling of the CFDDP pilot hole was carried out in 2012, to accomplish stratigraphy and petrology determinations (De Natale et al., 2016), as well as to perform several experiments to determine ‘in situ’ stress and permeability parameters (Carlino et al., 2015). In particular, we used a modified leak-off test procedure (LOT) to simultaneously determine stress parameters and permeability at depth. The experiment was carried out at the depth of 500 m (bottom hole) before casing the hole with a liner. Water was pumped at a constant rate of 38 l/min in the well that was hermetically closed, in several time intervals; the resulting plot of overpressure measured at the well head is reported in Fig. 7a. The general envelope of the overpressure as a function of time has been used for the ordinary leak-off test (Carlino et al., 2015) aimed at determining the minimum principal stress. During the rising path of the well head overpressure, before the formation breakdown pressure (FBP in Fig. 7a), water injection has been suddenly stopped and then reactivated three times. When water injection is instantaneously stopped, the well head pressure decreases at a rate which only depends on the permeability of the open rocks at the bottom hole. This procedure was repeated three times to get a more robust determination of permeability. The three resulting values are however very close one another, showing an average permeability value of $3 \times 10^{-13}\,m^2$. This determination, obtained at 500 m of depth, has been integrated at larger depths with some reliable permeability measurements obtained by ENEL-AGIP during their geothermal explorations carried out between 1975 and 1985 (AGIP, 1987). From these measured values, a suitable permeability curve (K) as a function of depth (Z) has been computed, based on the simple parametric relations found by Manning and Ingebritsen (1999): $\text{Log K} = -14 - 3.2 \times \text{Log Z}$. Fig. 7b shows the permeability-depth function so determined.

Table 1
Values of the parameters characterizing the physical properties of rocks (density, specific heat, thermal conductivity and porosity) are chosen based on literature and data wells (e.g. Rosi and Sbrana, 1987; AGIP 1987).

Rock physical property	
Density (kg m3)	2100
Thermal conductivity (W m-1 K-1)	2,8
Specific heat (J kg −1 K -1)	1000
Porosity	0.2

2. Results

We refer here to the method used by Troiano et al. (2011), for details on the computation of a time-dependent model of surface vertical displacement during Campi Flegrei 1982–1985 unrest and subsequent subsidence. It is based on the computation of the changes in pressure and temperature at the grid points representing the whole rock volume of the Campi Flegrei caldera, and then of the surface displacement generated by such pressure and temperature changes. The pressure and temperature changes produced, at each time step, by the deep fluid injection are computed by the THOUGH2 program (Pruess, 1991). The resulting surface displacements are computed by the COMSOL Multiphysics finite element program. We again refer to Troiano et al. (2011) for any detail about the used method. Here, we show, in Table 1, all the used parameters.

The resulting vertical displacement at the central point of the model surface (which represents the point of maximum vertical displacement at Pozzuoli harbor), with the given values for the fluid injection rate at bottom of the aquifer and for the permeability as a function of depth, are shown in Fig. 8. The shadowed area shown in figure represents the time interval (three years) of the injection before abrupt interruption. The return to zero of ground uplift after stopping injection matches very well the observed subsidence after the 1984 peak. The rate of subsidence for a sudden interruption of fluid injection is controlled by permeability; so, the good match with observations confirms that the used permeability function is a good approximation of the real one. The maximum obtained ground uplift, with adopted injection and permeability parameters, is about 0.7 m, which is the amount of uplift subsequently recovered by the 1984–2003 subsidence, after about twenty years. The evidence that, with realistic parameters, the maximum inferred uplift can be made similar to the recovered part of the uplift, and the rate of subsidence can be made the same than observed, further corroborates the conclusion that the part of ground uplift recovered by subsequent subsidence was due to the influx of deep fluids, other than magma, in the shallow aquifers. We want to stress, however, that the presented one is just one of the possible models: whereas small changes of most parameters does not change significantly the results, changes of the permeability function and/or of the fluid injection rate, can significantly affect them. However, we want to focus on the feasibility to explain the part of uplift subsequently recovered in terms of deep fluid injection and resulting perturbations in the shallower aquifers. Once discriminated the amount of hydrothermal effects, we can infer, by subtracting them from the total uplift, the amount of uplift likely due to shallow magma intrusion: it is about 1.0–1.1 m. From such a value, we can give a corrected estimate of the amount of intruded magma, assuming a realistic intrusion depth. The depth of magma intrusion, according to several results, should be in the depth interval 2.5 to 4 km (Dvorak and Berrino, 1989; Ferrucci et al., 1992; Bellucci et al., 2006; Woo and Kilburn, 2010). Since we are dealing with a sill intrusion, well approximated by a penny shaped opening crack, we can use the results of Woo and Kilburn to model the average opening and the volume change of the crack. We recall here the used model, which was originally obtained by Fialko et al. (2001) for a penny shaped crack embedded in a homogeneous elastic medium, in the approximation that the ratio ½ < r/R < 4/3, where r is the crack radius, R is the depth, n and m are, respectively, the Poisson ratio and the medium rigidity, ΔP is the pressure change in the crack:

$$\Delta U_{\max} \cong \frac{8(1-\upsilon)}{3\pi}\frac{\Delta P}{\mu}\left(\frac{r}{R}\right)^{\frac{5}{3}}r$$

By fitting vertical displacement data, these authors found the best fitting model for a depth $R = 2.75$ km and source radius $r = 2.1$ Km. As shown by Woo and Kilburn (2010) the volume change can be computed from the formula:

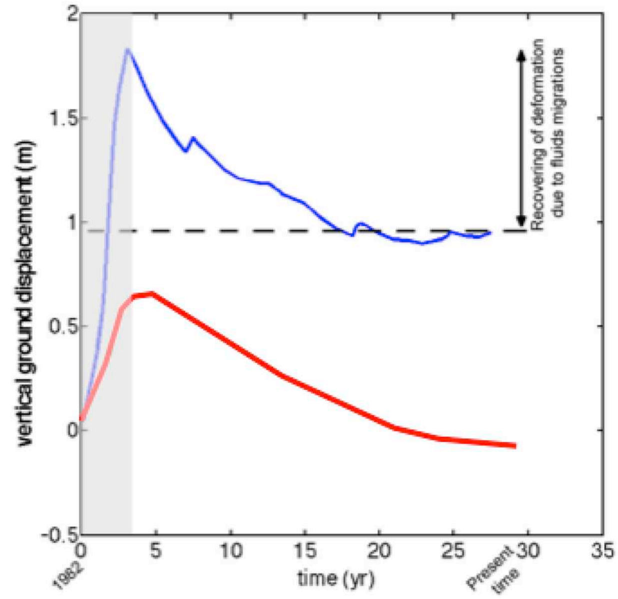

Fig. 8. Comparison between maximum ground deformation computed by an axisymmetric fluid-dynamical model (in red) and the maximum vertical deformation observed in the period 1982–2014 (in blue). In the model, injected fluid flow at the base of the system (at a temperature of 350 °C) is suddenly started, at a rate of 6000 t/d representing the observed increase of total gas flow at surface (see Troiano et al., 2011), and applied for 3 years (gray shadow area marks the time of injection), after that the injection rate is suddenly stopped. The permeability used, as a function of depth, is that computed by in-situ tests and shown in Fig. 7b. Surface ground deformation is obtained by computing, with Comsol numerical program, the effect due to changes in the internal pressure and temperature (see Troiano et al., 2011). (For interpretation of the references to colour in this figure legend, the reader is referred to the web version of this article.)

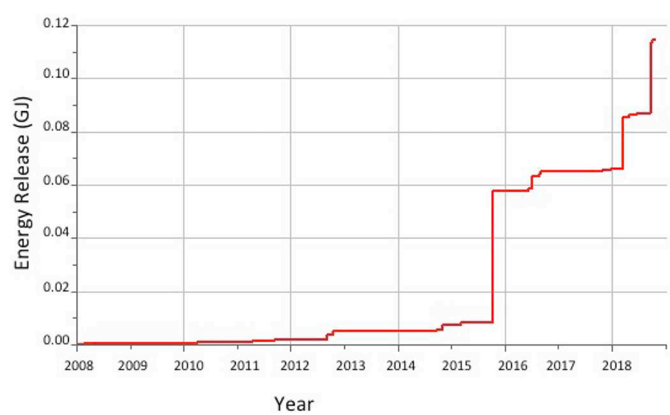

Fig. 9. Plot of cumulative energy release from earthquakes occurred at Campi Flegrei caldera in the last ten years (2008–2018). The clear increase of slope of the curve in the last 3–4 years clearly indicates an increase of earthquake energy.

$\Delta V \cong \frac{8(1-\upsilon)}{3}\frac{\Delta P}{\mu} r^3$

In order to justify 1 m of net uplift, using an average rigidity of 5×10^9 Pa, the overpressure in the crack must be $\Delta P = 6.7$ MPa, so that the resulting $\Delta V = 0.025\ km^3$.

This is the approximate amount of magma volume intruded at shallow depth (in this computation 2.75 km).

We stress this magma volume represents a rough estimate of the true intruded volume. This is not only due to the intrinsic approximations of the sill model, but also to the fact that, in a layered medium, the intruded volume required to produce a given surface uplift can be different (generally larger e.g. Folch et al., 2000; Manconi et al., 2007).

2.1. The nature of unrest episodes

Magma intrusion at shallow depth has been often hypothesized to explain the strong unrest episodes taking place at Campi Flegrei caldera in the period 1969–1984 (i.e. Corrado et al., 1977; Berrino et al., 1984; Dvorak and Berrino, 1991; Bellucci et al., 2006). Other models have been however developed, able to explain the observed ground deformation episodes in terms of thermal-fluid-dinamical perturbations in the shallow geothermal system (e.g. De Natale et al., 2001; Chiodini et al., 2003; Troiano et al., 2011; Gottsmann et al., 2006). It is of fundamental important, both for scientific research and for civil protection, to shed light on the mechanisms producing the observed unrest at different periods. In particular, the most important question, given the long lasting, ongoing unrest, is to discriminate the episodes and the amount of magma intrusion at shallow depths, from other effects mainly linked to deep fluid injection and related perturbations in the shallow aquifers. Important insight, to answer this question, is given by geochemical data. They show marked differences between the 1982–1984 and the subsequent periods (Moretti et al., 2017, 2018). During 1983–1984, a sharp peak in the H_2O/CO_2 and other indicators (Fig. 4), and an evident increase of the oxiding species (Moretti et al., 2017) strongly suggest the occurrence of shallow magma intrusion. Several recent interpretations, besides Moretti and co-authors, point out shallow magma intrusion at that time (Chiodini et al. 2016 and references therein; De Siena et al., 2010; Amoruso and Crescentini, 2011; D'Auria et al., 2015, Giudicepietro et al., 2017) In this paper, we give a more accurate estimation of the volume of such intrusion, by considering the amount of ground uplift 1982–1984 that was not recovered by later subsidence. The interpretation of the part of ground uplift recovered by later subsidence in terms of an hydrothermal effect due to perturbation of shallow aquifers, had formerly been proposed by De Natale et al. (1991, 2001, 2006a,b); however, some recent papers (Macedonio et al., 2014; D'Auria et al., 2015; Giudicepietro et al., 2017) questioned such interpretation by claiming that the subsidence could be explained by a purely magmatic model, involving shallow sill intrusion. In this paper, we showed that such a purely magmatic model cannot explain the Campi Flegrei unrest. In fact, in order to produce later subsidence after initial uplift, an intruded sill should largely expand laterally, and would consequently generate a significant, progressive lateral extension of the deformed area, which is in strong contrast with the observations. Riaffirming that the subsidence can only be produced by a later effect of hydrothermal inflation, allows us to compute, from the residual uplift left aftter subsidence, an intruded magma volume in the period 1982–1984 of $0.025\ km^3$. This volume is roughly equivalent to the amount of magma extruded in the last, 1538 eruption.

After 1984, there is no further evidence, from geochemical data, of other periods of sharp increase of the ratio H_2O/CO_2, or of the oxidizing species, which would indicate new magma intrusions (Moretti et al., 2017, 2018). We observe, on the contrary, a sharp decrease of the ratio H_2O/CO_2 starting around 2000. This also the same period when the subsidence stops and the new uplift episode, still on-going, starts. Such a decrease is consistent with a major change in the depth of the magma reservoir from which the gas are exsolved. This in turns suggests that the shallow magma intrusion, whose depth can be estimated at about 3 km, had completely solidified around 2000, and the prevailing injected gases came from the deeper magma chamber, located at about 8 km of depth. A solidification time of about 20 years, for a thin sheet also much thicker than 2 m, is absolutely reasonnable, as it has been shown by Moretti et al. (2018).

The on-going uplift, started between 2004 and 2005, is characterized by an average uplift rate of about 5 cm/year, with some peak values, like the period 2012–2013, up to 15–16 cm/year. These uplift rates are much smaller than those recorded during the 1970–1972 and 1983–1984 periods, in which they reached peak values close to 100 cm/year, i.e. about one order of magnitude larger. Moretti et al. (2018) further evidenced the sharp simmetry of the subsidence recorded after 1984 and the present uplift, well fitted by a simple mathematical function shown in Fig. 2. In this fitting curve, the medium point marking the end of subsidence and the starting of uplift is in the year 2003. Both the general subsidence and the uplift trends, however, are characterized by sharp and short lasting uplift episodes which were identified by Gaeta et al. (2003) and named 'mini-uplifts'. At the time of the Gaeta et al. paper, such mini-uplifts had occurred in 1989, 1994, 2000; i., e. with a rather periodic occurrence rate of 5–6 years. More recently, another mini-uplift occurred in 2006, which also marked the beginning of the new, on-going uplift phase (Troise et al., 2007). Also the uplift phase was characterized by rather periodic occurrence of episodes of faster uplift which, since superimposed on a background uplift trend, have not been recognized till now as 'mini-uplifts'. The most important of such events occurred in 2012–2013, about six years after the 2006 mini-uplift, thus preserving almost the same period of the previous mini-uplifts. The 2012–2013 event has been interpreted by some authors as due to an episode of shallow magma intrusion (e.g. D'Auria et al., 2015; Chiodini et al., 2016 and references therein), and was the main reason why the Italian Civil Protection issued, in December 2012, the first step of volcanic alert in the area ('Yellow' level). However, based on our previous results ruling out the possibility that, at Campi Flegrei, an episode of shallow magma intrusion can produce a subsidence phase following the uplift, we must conclude that all the mini-uplifts, in which the additional uplift is then recovered, cannot reflect shallow magma intrusions. The 2012–2013 event was characterized by an increase of the uplift rate, reaching about 16 cm/year, later decreasing to the previous uplift rate; then, it was simply a mini-uplift episode, with uplift rate of about 8 cm/year, superimposed to a background uplift rate. From Fig. 2 it is clear that well recognisable mini-uplift episodes occurred in 1989, 1994, 2000, 2006, 2012; interestingly, a minor episode could be also noted in 2016, about exactly four years after the starting of the 2012 event, although the recovery of uplift does not seem complete. The nature of mini-uplifts can be interpreted basically in two ways: the first is the periodic increase of deep fluid injection, the second is the almost periodic self-sealing of the porous medium in which deep fluids are injected. The first mechanism seems more appropriate for mini-uplifts occurred till 2006, which were modeled this way by Chiodini et al. 2003). For the 2012 and 2016 episodes, however, there was no evidence of any modulation of the deep gas injected. An almost periodic self-sealing, mainly caused by salt precipitation in the porous matrix, during an almost constant gas flow from below, could also cause the observed pulsed character of mini-uplift. Discriminating between the two mechanisms is, however, out of the scope of this paper, but the main point was to assess that mini-uplifts cannot be caused by magma injection at shallow depth.

Once stated that mini-uplifts cannot indicate shallow magma intrusion, we must also conclude there is no evidence that the present, ongoing uplift is driven by shallow magma intrusion. There are, on the contrary, many opposite evidences: one of them is the already mentioned observation, evidenced by Moretti et al. (2018), that the general trends of the post-1984 subsidence and the post-2003 uplift are specular, thus indicating the same mechanism, but with opposite sign. Since the mechanism of subsidence is very likely due to the deflux of previously injected deep fluids, then the on-going uplift should be due to a new period of injection, at an almost constant rate; on the contrary, if the present uplift would be due to shallow magma injection, we should hypothesize an almost continuous injection of magma into a shallow reservoir, at an almost constant rate as required by the almost constant uplift rate observed since 2007. However, the most striking observation ruling out shallow magma intrusion during the present unrest is the lack of sharp peaks of H_2O/CO_2, which would indicate the arrival and/or degassing of magma at shallow (3–4 km) depth. Such evidence, very clear in the period 1983–1984, is not only absent in following periods, but is completely contrary since in the last 18 years we observe a continuous decrease of the H_2O/CO_2 ratio. Chiodini et al. (2016) tried to interpret such evident discrepancy by hypothesizing that H_2O is lost in the shallow crust by significant condensation. That hypothesis, which was based on the assumption that the conditions of temperature and pressure along the fumaroles allow the coexistence of liquid and gaseous water. Such an assumption, always implicitely or explicitely used in the papers on Campi Flegrei by Chiodini (2016 and references therein) and co-authors (e.g. D'Auria, 2015) was in turn based on the simulations of the thermal model of the Solfatara-Agnano area (the one in which the most important fumaroles are located), (Chiodini et al., 2003; Todesco et al., 2003). However, we demonstrated, in this paper, that their modelling and related results were affected by a highly approximated boundary fluid enthalpy, so that using the same parameters they wanted to test, the correct result is that temperature and pressure conditions of the plume feeding fumaroles are such to not allow liquid/gas mixture, but only a superheated vapor. Our present result confirms, in the most direct way, what already shown in another way by Moretti et al. (2018) and formerly hypothesized by Moretti et al. (2013a, 2017).

In conclusion, all the evidence point out that, in the period 1982–1984, an episode of shallow magma intrusion cause a large and fast uplift episode, whereas the present, on-going unrest, has started after the solidification of the 1982–1984 magma intrusion, and is driven by injection, at the aquifer bottom, of deep fluids coming from the main, deeper magma chamber, located at about 8 km of depth (Zollo et al., 2008). The continuous decrease of the H_2O/CO_2 ratio as well as other indicators (Fig. 2) in the main fumaroles indicate a different gas signature starting about in 2000, just after the almost complete solidification of the 1982–84 sill intrusion at shallow depth. This signature reflects the involvement of the gas flux coming from the deeper magma chamber, which is progressively reaching a steady state upon interaction with the basal hydrothermal system (Moretti et al., 2013a, 2017, 2018).

2.2. Insight on the possible evolution of present unrest

The evidence that shallow magma intruded in the ‘80s is not active anymore because cooled, and that no further shallow magma intrusion seems to have occurred after 1984, pose strong constraints on any interpretation of the on-going unrest. The absence of recent intrusion episodes, and the absence of molten magma at shallow depth certainly implies a lower risk of eruption as the end of the present unrest. Although the considerations about the shallow magma migration and/or presence at shallow depths is classically the main volcanological argument to assess the likelyhood of an eruption (Selva et al., 2012; Marzocchi and Bebbington, 2012), Kilburn et al. (2017) presented more specific arguments, well grounded on the fracture mechanics, to use for eruption forecast. Their model indicates that, at the end of 1984, i.e. at the time of maximum uplift, the rocks were entering the ‘critical’ stage, in which any further incremental stress is accomodated by fracture, rather than by deformation. This implies, as a first consequence, that seismicity should increase (in number and magnitude) as the ground uplift increase, approaching to the ground level of 1984. Furthermore, when the ground level will reach the 1984 level (which was about 0.5 m higher than today) the seismicity will likely increase at the same level it was in that period. The evolution of seismicity in the last decade already confirms an increasing trend, more pronounced in the last years. In 2015, the earthquake occurred on October 7, $M_d = 2.5$, had been the largest one since 1985; in 2018, on Semptember 18, an even larger earthquake of estimated magnitude 2.5, but locally felt much stronger, occurred. Fig. 9 shows the plot of seismic energy release as a function of time, in the last 10 years. It is very clear the increasing trend, more pronounced in the last years.

So, the main problem if the uplift does not stop in few years, will be the strong increase of seismicity. We recall that, in the 1983–1984 period, we recorded > 10,000 earthquakes, with maximum magnitude M = 4.2 and peak frequency of about 200 events/h (De Natale and Zollo, 1986).

Regarding the implications for a possible eruption, Kilburn et al. (2017) pointed out that, once the ground level will reach the 1984 threshold, the system could enter the critical behavior, with high probability of large scale fracturing, which in turns could open the way to an eruption. However, although such an evolution only depends from the stress accumulation and changes in rehology of rocks, if magma is not present at shallow depth the hypothesis that the on-going unrest can end with an eruption is not very likely. Furthermore, the critical threshold computed by Kilburn et al. (2017) does not take into account the effect of very high temperatures at shallow depth that, given the very high geothermal gradient of the area (Carlino et al., 2012) could strongly affect the rehology of the rocks (Carlino and Somma, 2010), making it more ductile and hence less prone to extensive fracturing. Such a behavior is strongly suggested by the very high uplift occurred before the last, 1538 eruption (Di Vito et al., 2016), which was much larger than the conservative threshold used by Kilburn et al. Such evidence from the past eruption further corroborates the hypothesis that the possibility that the present unrest ends up with an eruption in few years, if the uplift rate continues at the present level, is unlikely.

3. Discussion and conclusions

Discriminating magma intrusions from hydrothermal effects causing unrest at large calderas is one the most important problems for volcanological research and for civil protection purposes. Campi Flegrei caldera is the place where the interplay of the two possible effects is more evident, and where it is most important to discriminate each of them due to the extreme urbanization and population of the area. We have here analyzed 38 years of data, showing a clear difference between what occurred in the large unrest of 1982–1984, characterized by very high uplift rates (up to 0.7 m/y), and in the present, still on-going one. The large unrest of the ‘80s, leading to a maximum observed uplift of 1.86 m in less than three years, show clear evidence for magma intrusion. The amount of maximum uplift, after subtracted the likely effects of hydrothermal perturbations due to deep, non magmatic fluids injection, indicates a volume of magma intruded at shallow depth (around 3 km) of 0.025 km^3. This magma, intruded in form of a thin sheet (few meters of thickness) has solidified in < 20 years. Geochemical data confirm the signature of a significant change in the composition of gas emitted at fumaroles, which indicates it is flowing from a much deeper magma. The change in character of gas emission is roughly coincident (from 2000 to 2003) with the end of the subsidence period started at the end of 1984. Geochemical data also confirm the absence, after the 1984, of any signal for new magma intrusion at shallow depth. Moreover, we have here demonstrated that some recent hypotheses of recent magma intrusion at shallow depth are biased by some incorrect assumptions and/or modelling. In conclusion, there is evidence that the large uplift of 1982–1984 has been driven by both magma intrusion, accompanied by injection of deep, non-magmatic fluids, at shallow depth. The later subsidence, ended around 2000–2003, was likely due to the stopping of deep fluid injection at shallow levels. The gas emission in the area, till 2000, was still reflecting the processes occurring in the shallow intruded magma, with some episodes of deeper gas injection accompanying the mini-uplift episodes. The present, on-going unrest started around 2000–2003 is, on the contrary, likely to be due to the increased injection of gas coming from the deeper magma chamber, in absence of any further shallow magma intrusion and/or residuals of previously intruded molten magma. This episode is markedly different from the large uplifts of 1969–1972 and 1982–1984 mainly for the much lower uplift rate (0.08 m/year) almost one order of magnitude less; and is strongly different also in the geochemical anomalies of fumarole gas. The pattern of up and down movements recorded from 1982 is made even more complex by the likely modulation of permeability in the area, due to both self-sealing caused by high fluid flows in pore and fractures, and the fracturing due to earthquake occurrence and pressure build-up. The results of this work have important implications on the possible evolution of the present unrest. In fact, while they confirm that a shallow magma intrusion episode occurred in 1983–1984, they also indicate that that intruded magma sheet has been solidified, and the present unrest is not driven by continuous, or else episodic new magma intrusions. The hazard implications of a scenario in which shallow magma intrusions continuously occur, would be actually extremely critical. On the other side, our results showing that no shallow magma is involved in the recent unrest are relatively less critical. However, the three episodes of large unrest occurred recently testify that a change with new magma intrusion episodes is always possible, although in that case we would likely observe an abrupt change to high uplift rates. Approaching the ground level of 1984, even with the present, low rate uplift unrest, implies however the high probability of more intense and frequent seismicity, as predicted by the recent physical stress model of Kilburn et al. (2017).

In conclusion, the study of Campi Flegrei unrest put in evidence that the complex phenomena of caldera unrest, often showing uplift and subsidence occurrence without interbedded eruptions, must be analyzed by rigorous approaches jointly considering geophysical and geochemical data. Using well suited and well grounded methods, the effects of magma intrusion and of hydrothermal perturbations can be at some extent discriminated, thus allowing a much more reliable inference on the likely evolution of on-going unrest. This is really valuable, not only to shed light on the caldera volcanism, which is the most explosive and less known on the Earth; but even to reliably direct civil protection measures, that are crucial, particularly in highly populated calderas.

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