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Review Article

Interpretation of Recent Unrest Events (Bradyseism) at Campi Flegrei, Napoli (Italy): Comparison of Models Based on Cyclical Hydrothermal Events versus Shallow Magmatic Intrusive Events

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Several recent models that have been put forth to explain bradyseism at Campi Flegrei (CF), Italy, are discussed. Data obtained during long-term monitoring of the CF volcanic district has led to the development of a model based on lithological-structural and stratigraphic features that produce anisotropic and heterogeneous permeability features showing large variations both horizontally and vertically; these data are inconsistent with a model in which bradyseism is driven exclusively by shallow magmatic intrusions. CF bradyseism events are driven by cyclical magmatic-hydrothermal activity. Bradyseism events are characterized by cyclical, constant invariant signals repeating over time, such as area deformation along with a spatially well-defined seismogenic volume. These similarities have been defined as "bradyseism signatures" that allow us to relate the bradyseism with impending eruption precursors. Bradyseism is governed by an impermeable shallow layer (B-layer), which is the cap of an anticlinal geological structure culminating at Pozzuoli, where maximum uplift is recorded. This B-layer acts as a throttling valve between the upper aquifer and the deeper hydrothermal system that experiences short (1-10^2 yr) timescale fluctuations between lithostatic/hydrostatic pressure. The hydrothermal system also communicates episodically with a cooling and quasi-steady-state long timescale (10^3-10^4 yr) magmatic system enclosed by an impermeable carapace (A layer). Connectivity between hydrostatic and lithostatic reservoirs is episodically turned on and off causing alternatively subsidence (when the systems are connected) or uplift (when the systems are disconnected), depending on whether permeability by fractures is established or not. Earthquake swarms are the manifestation of hydrofracturing which allows fluid expansion; this same process promotes silica precipitation that seals cracks and serves to isolate the two reservoirs. Faults and fractures promote outgassing and reduce the vertical uplift rate depending on fluid pressure gradients and spatial and temporal variations in the permeability field. The miniuplift episodes also show "bradyseism signatures" and are well explained in the context of the short timescale process.

1. Introduction

About a billion people, roughly 15% of the world population, live in areas of volcanic risk because many population centers developed around dormant volcanoes that have not erupted in recent years. The Neapolitan volcanic area is one such high volcanic risk area [1], with about three million people living within an area with a radius of ~25 km in a volcanic province that includes active volcanoes such as Vesuvius, Campi Flegrei (CF), and Ischia [2] (Figure 1(a)).

The CF is currently the area at greatest risk within the Neapolitan volcanic area as it is actively experiencing the
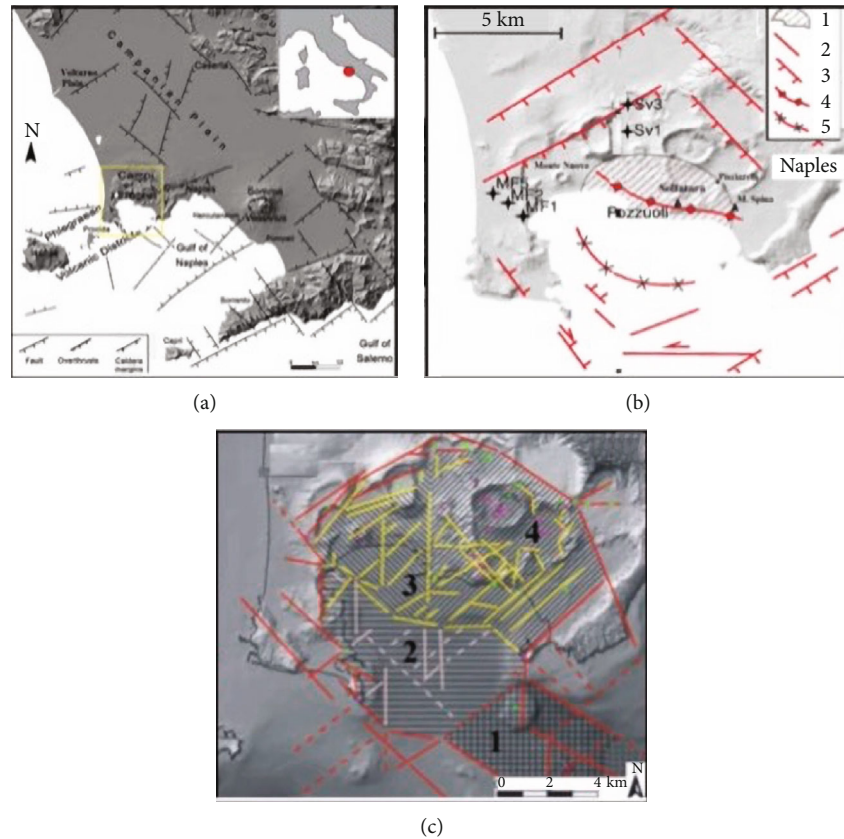
(a)

(b)

(c)

Figure 1: (a) Structural sketch map of the Campanian Plain and surrounding Apennines. (b) Structural map of the CF–Pozzuoli Bay area: (1) area affected by an uplift of 100–50% during the 1982–84 bradyseism; (2) fractures and lateral faults; (3) normal faults; (4) Pozzuoli anticline; (5) Pozzuoli Bay syncline. Black stars geothermal wells: MF = Mofete; SV = San Vito (from Milia et al. [17]; Milia and Torrente [18]). (c) Structural sketch map of CF caldera: (1) sector deformed during caldera collapse; (2) undeformed to subsiding portion of the caldera floor; (3) eastern sector of the resurgent portion of the caldera floor; (4) western sector of the resurgent portion of the caldera floor; red lines = regional faults activated during caldera collapse; yellow lines = regional faults reactivated during resurgence of a portion of the caldera floor; white lines = regional faults reactivated during subsidence of a portion of the caldera floor (modified from Capuano et al. [96]).

phenomenon of ground uplift and subsidence called brady-seism (from the Greek, meaning "slow movement"), a term introduced by ancient Greeks who populated the area before the Romans [3]. For this reason, the region is continuously monitored using a variety of geophysical (seismicity), geo-chemical (composition of emitted gasses and fluids), and volcanological tools in order to better constrain and under-stand the volcanic hazard. The geology (sensu lato) of CF has been intensively studied for the last century by applying modern ideas and methods, and yet, significant differences of opinion exist regarding the fundamental processes leading to bradyseism over the past few thousand years. The need to better understand CF volcanic activity is fundamental to protect the population from hazards linked to explosive vol-canic eruptions and to understand the role of seismicity as a possible precursor of a potential future eruption. CF experi-enced many bradyseism episodes since at least Greek-Roman time, but only once, in 1538 CE, after a ground uplift of ~7 m, was there a recorded eruption (Monte Nuovo) [4]. After the 1538 CE eruption, a long period of subsidence ensued until 1950 when a phase of rapid uplift of about 0.8 m started, followed by about twenty years of subsidence

that lasted until 1969 [3]. There were two other dramatic bra-dyseism phases in 1969–1972 and 1982–1984, always starting with uplift followed by an amount of subsidence that did not balance the preceding uplift [3], i.e., the subsidence phase did not return the surface to its initial preuplift level (Figure 2).

Unrest in 1982-84 was especially concerning; it resulted in ~1.8 m of uplift and ~16,000 associated earthquakes, most of which were of low magnitude [5, 6]. After a ~22-year period of subsidence, a new phase of slow uplift began at the end of 2005 and continues to the present time. From 1984 to today, several sharp and short bradyseism episodes (displacements generally ~0.1 m or less) have been regis-tered: in 1989, 1994, 2000, 2006, and 2012–2013 (Figure 2). Since 1950, all bradyseism episodes have followed very sim-ilar ground deformation patterns, with only uplift and subsi-dence rates and magnitudes showing variation [7]. Inflation events are accompanied by seismic crises that define a spa-tially fixed seismogenic volume during unrest with a circular shape (radius ~6 km) around the town of Pozzuoli, where the greatest uplift is observed. The magnitude of the uplift and seismicity both decrease in intensity toward the margins of the circular uplift region [5–11].
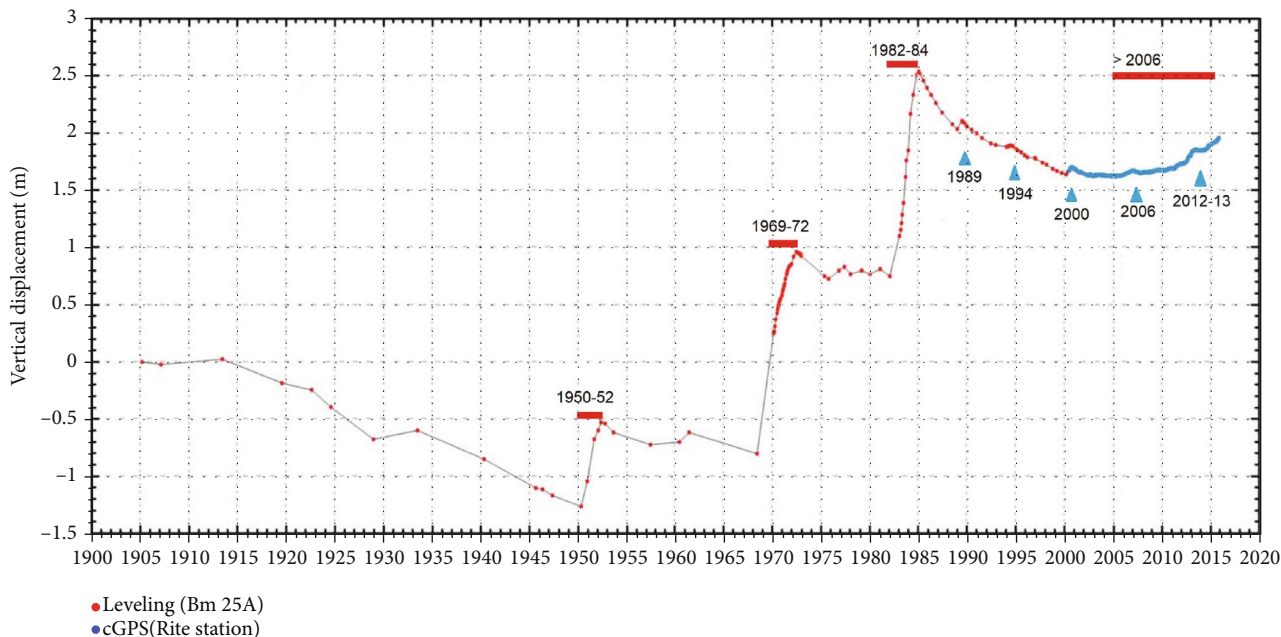
FIGURE 2: Ground displacement at Pozzuoli Porto since 1905, measured by precision leveling (until year 2000, in red) and at the close GPS benchmark of Rione Terra (since 2000, blue dots) (from Moretti et al. [60]).

In this review, we relate the recent CF bradyseism events to cyclical magmatic-hydrothermal processes that are characteristic of the cooling and crystallization of epizonal, water-bearing magmas, as first described by De Vivo and Lima [12]. A quantitative conceptual model describing this process was later reported by Bodnar et al. [13], followed by a rigorous thermodynamic model describing the process [14]. Here, we first emphasize that bradyseism displays repetitive cyclical signals without substantial variations in terms of recognizable signatures; the latter can be helpful to distinguish bradyseism from impending eruption precursors. Secondly, by combining the cyclical model with results of other investigations, we show that the bradyseism mechanism is not linked to recent shallow emplacement of magmatic intrusions as proposed in the recent literature. Finally, we show that the cyclical nature of the observed bradyseism is closely related to the complex geological, tectonic, and stratigraphic upper crustal structures that govern spatial and temporal variations of the subsurface permeability.

2. Geological-Structural Pattern, Volcanic History, and Models for Recent Bradyseism at Campi Flegrei

2.1. Geological-Structural Pattern. The CF volcanic complex is located within the upper Pliocene graben structure of the Campanian Plain that developed on the western margin of the Apennine Mountain chain after the opening of the Tyrrhenian Basin (Figure 1(a)). The NW–SE extensional tectonics, connected to the opening phases of the Tyrrhenian Sea, controlled the morphosedimentary evolution of the Campania coast while volcanism was still active. In the period from 700 ka to 400 ka, several deep sedimentary basins formed, including the Volturno Plain, Gulf of Pozzuoli, Gulf of Naples, and Salerno (Figure 1(a)). The boundaries of the CF area onshore and offshore are characterized by high-angle faults. In the CF caldera area, a compressional tectonic regime is active, with an anticline culminating near the town of Pozzuoli and a syncline located beneath Pozzuoli Bay (Figure 1(b)). The rate of fold uplift ranges from 1 to 20 mm/yr [15–19]. Other studies (e.g., [20]) generally report an extensional tectonic regime without going into much detail in the area within the caldera. The structural sketch map of the CF caldera (Figure 1(c)) shows the limit between eastern and western resurgent caldera sectors passing through the town of Pozzuoli, where the anticline culminates and where during bradyseism events, the maximum elevation changes are recorded.

In the Campania Plain, volcanism started with fissure-sourced ignimbrite events (from >300 to 19 ka) [21–23] and is still active in the Neapolitan area (Mt. Vesuvius and CF). In spite of the evidence reported by de Vivo et al. [21] and Rolandi et al. [22], before 2012, most authors reported that the CF caldera was formed by two eruptions, the Campania Ignimbrite (CI) at 39 ka and later by the Neapolitan Yellow Tuff eruption (NYT) at 15.6 ka ([24, 25] and ref. therein). In 2012, a 506 m drill-hole in the Bagnoli plain in the CF eastern sector revealed that the CI volume is much lower than the NYT volume, in contrast to the hypothesis that the CI source was in the CF area, and supporting the hypothesis that the collapse at CF only occurred during the NYT eruption [26, 27]. The latter finding also implies that most of Naples urban area does not lie within the CF caldera. At least 73 phreatomagmatic eruptions younger than NYT are known to occur within the caldera area, grouped into three periods: epoch I 15.6-9.5 ka, epoch II 8.6-8.2 ka, and epoch III 4.5-3.8 ka ([28, 29] and ref. therein, [30]). In the

last 3.8 ka, only one eruption (Monte Nuovo; 1538 CE) occurred [4]. The spatial and temporal migration of volcanic activity after the eruption of the NYT toward the center of the caldera combined with a major reduction in the volumes and an evolution in the chemistry of the erupted products suggests that recent CF activity represents the evolution of a large cooling magma chamber [31].

2.2. History of Bradyseism.

Since Roman times, the Pozzuoli area has undergone several uplift episodes followed always by subsidence. In 1905, the Istituto Geografico Militare (IGM) initiated precise leveling of CF establishing a line along the coast of Pozzuoli Gulf; before 1905, no precise leveling measurements are available [32]. After about 3.8 ka, when the last, large, explosive Monte Spina eruption occurred, a very small eruption of $0.02 \, km^3$ occurred in 1538 CE, forming Monte Nuovo (New Mountain) [4]. The 1538 eruption was preceded by about one hundred years of ground uplift marked by sea recession, and the area of greatest uplift was centered on the site of the eruption (3 km west of Pozzuoli) [3]. After the 1538 eruption, slow, regular subsidence occurred. The latter has been well documented after 1819 with the first measurements of sea level above the Serapis Temple floor, an archaeological site in Pozzuoli. More recently, bradyseismic events occurred in 1950-1952, 1969-1972, and 1982-1984 (Figure 2). The latter was of great concern for the inhabitants of Pozzuoli due to the magnitude of the uplift of about 180 cm, followed by 80 cm subsidence in the following 20 years [32]. At the end of 2005, slow uplift started again at rates comparable to those of subsidence (on average ~4 cm/yr) and the uplift continues today at an accelerating rate (Figure 2). During post-1984 subsidence and during the subsequent and still ongoing uplift, minor abrupt episodes of uplift, followed by minor subsidence and fast recovery of the whole uplift, have been registered in 1989, 1994, 2000, 2006, 2012-2013, and 2016 (Figure 2) [33]. The 2012–2013 uplift rate was faster (~16 cm/yr) than the others and has been studied in more detail [34]. In early 1982, uplift was quick (~100 cm/yr) and earthquakes reached significant magnitude in 1983. During 1984, the uplift continued, and the seismicity was very intense, with 33 events with $M \geq 3.0$ and six events with $M \geq 3.8$ (Figure 3) [6]. At the end of 1984, after an uplift of 1.79 m (Figure 2), subsidence started and seismicity quickly decreased and the composition of fumaroles at Solfatara, a volcano near Pozzuoli, recorded a greater contribution of magmatic gas [35].

2.3. Bradyseism Associated with Shallow Magma Intrusions.

Since 2010 almost all published studies interpreted CF unrest to be the result of intrusion of magmas to shallow depths of ~3 km (e.g., [5, 33, 36–49]). For a more extensive review on bradyseism interpretations, see Cannatelli et al. ([50] and ref. therein).

In the 2000s, De Vivo and Lima [12], Battaglia et al. [51], and Bodnar et al. [13] were the first to speculate that migration of fluid within the caldera hydrothermal system was the cause of ground deformation and consequent unrest, rather than intrusion of magma. Battaglia et al. [51] came to this conclusion by inverting leveling, trilateration, and gravity measurements collected between 1980 and 1995; they modeled the location, geometry, and density of the deformation source. Densities obtained were compatible with fluid density. Other studies [36, 38, 41, 43, 44, 52] also utilizing deformation and microgravity data found different densities mostly attributed to differentiated magma. Woo and Kilburn [41] reported that a sill-shaped magma intrusion, coming from the deeper reservoir and emplaced at a depth of ~3 km, gave rise to the recorded ground deformation for 1982-1984 unrest, similar to previous studies [36, 38]. Subsequent subsidence was attributed, by the latter authors, to temporary pressure changes in the aquifer caused by magmatic intrusion, but without details. Woo and Kilburn [41] corroborate the interpretation of Bellucci et al. [53] that all CF bradyseismic events are caused by sill intrusions, centered in the vicinity of Pozzuoli, the location of maximum uplift. Since 2010, most published papers agree with the Woo and Kilburn [41] model and have focused on developing more convincing explanations for the subsidence that always follows uplift [33, 54–61]. Noteworthy is the Macedonio et al. [55] model that associates uplift with sill magmatic injection that at the center of intrusion stops and rapidly spreads laterally; the injected magma accommodates by a further lateral expansion of the sill, so that the maximum uplift slightly decreases causing subsidence. For the 1982-84 event, the interpretation of Troise et al. [33] agrees with the Woo and Kilburn [41] model but argues that the sill-shaped magma intrusion caused an uplift of ~110 cm and the subsequent uplift of ~70 cm (to achieve the total uplift of ~180 cm) was due to the influx of deep fluids which over the next 20 years led to a slow subsidence. The miniuplifts, as defined by Gaeta et al. [62], registered during both subsidence and uplift (Figure 2), have been interpreted by most in the same way [44, 57–59, 63–65]. The 2012-2013 unrest was also interpreted to be the result of a shallow magma intrusion, and this triggered the Italian Civil Protection to increase the volcanic alert level in CF area ("Yellow" level) in December 2012. Troise et al. [33] did not support this interpretation and hypothesized that the miniuplifts can be caused by a periodic increase of deep fluid injection or periodic self-sealing of the porous medium in which deep fluids are injected, without explaining the details.

Chiodini et al. [66] measured the fumarolic gas composition and CO_2 fluxes in the Solfatara and Pisciarelli areas during the last 10 years and argued that magmatic fluids related to magma intrusion at depth cause volcanic unrest because they heat and pressurize the CF hydrothermal system which in turn transfers energy to the host rocks triggering low magnitude seismicity and gas emissions. The geobarometric and geothermometric relations have been determined by mass balance considering that H_2O and CO_2 react to form H_2, CO, and CH_4. They calculated that the pressure-temperature increase at the top of a vertically elongated (0.3–2 km deep) gas front is responsible for low magnitude earthquakes and surface CO_2 gas emissions registered in the last years.

The involvement of shallow magma intrusions is based on the decreasing H_2O/CO_2 ratio in gas composition of the Solfatara volcano fumaroles since 2000 [57–61] (Figure 4(a)).
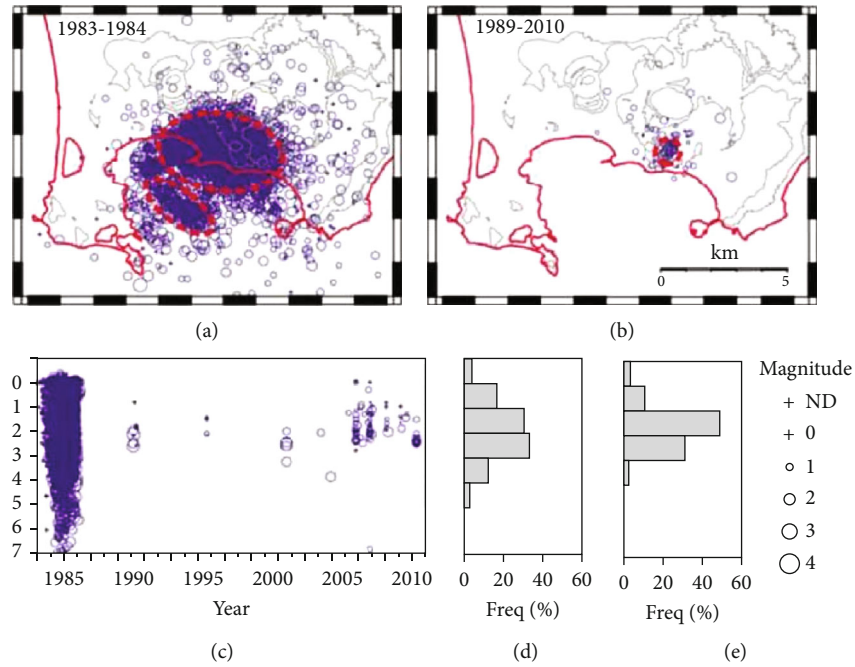
FIGURE 3: (a) Blue circles indicate epicenters of earthquakes recorded between 1983 and 1984; the dashed red ellipses mark the two main clusters. (b) The same for the period 1989–2010, the dashed red circle marks the cluster. (c) Hypocenter depths as a function of time. (d) Percentage of hypocenters for each depth range for the period 1982–1984 (the depth scale is the same as for (c)). (e) The same for the period 1989–2010 (from D'Auria et al. [6]).
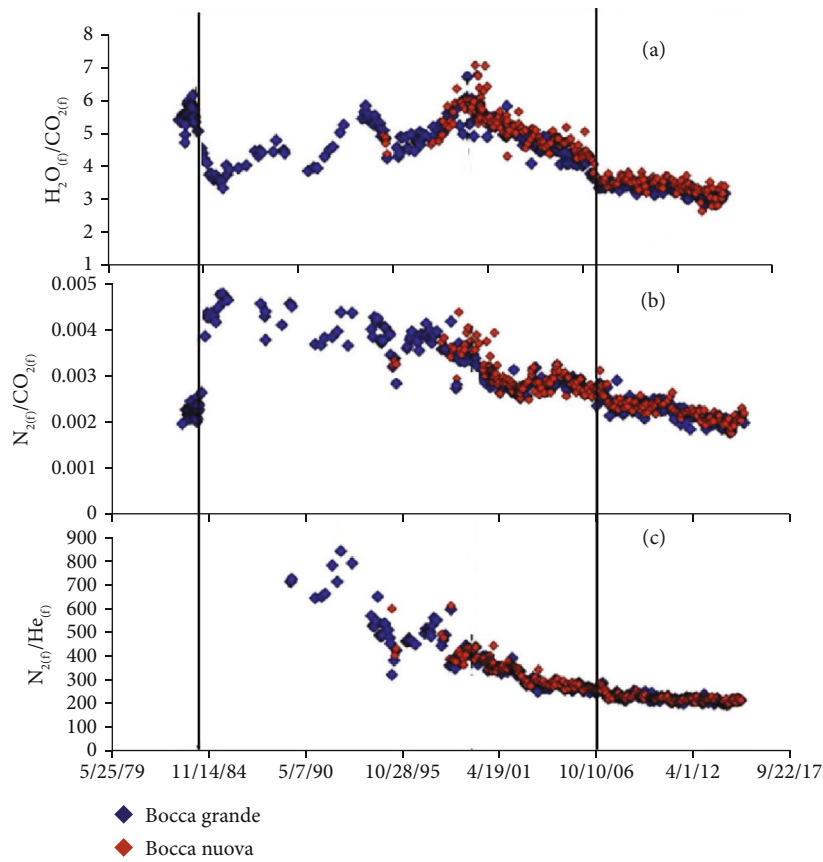

FIGURE 4: Gas ratios measured at Solfatara Bocca Grande and Bocca Nuova fumaroles: (a) H_2O/CO_2, (b) N_2/CO_2, and (c) N_2/He (modified from Troise et al. [33]).
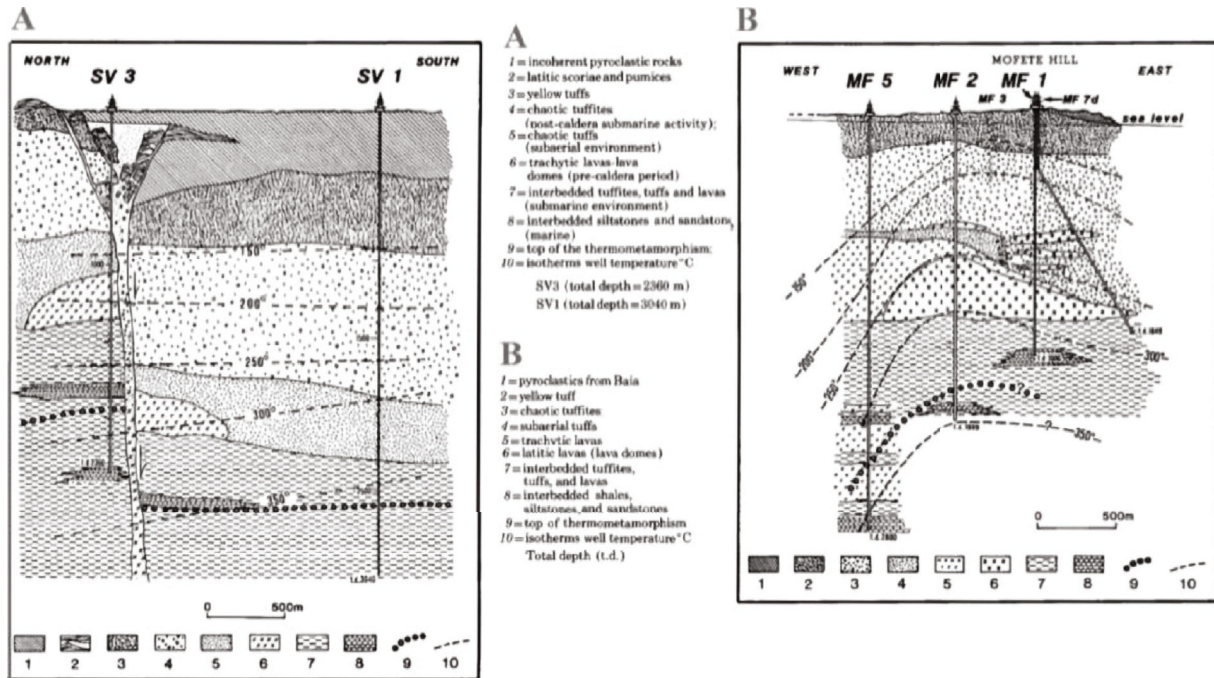
FIGURE 5: (a) Cross-section including SV1 and SV3 wells. (b) Cross-section including MF1, MF2, and MF5 vertical wells. MF3 and MF7d wells are also shown. See Figure 1 for location (from Agip [70]; De Vivo et al. [74]).

A very recent model by Nespoli et al. [67] differs from others mainly in that a magmatic intrusion is not required to explain bradyseism. These workers apply a physical model which considers that a Thermo-Poro-Elastic (TPE) inclusion, with an assigned geometry, is responsible for deformation induced by mechanical effects of both pore pressure and temperature changes of the fluids which pervade a poroelastic region, embedded in an elastic matrix. To explain the CF bradyseism phase during the 1982-1984 period, Nespoli et al. [67] recognized the presence of a layer with time-dependent permeability below the shallow aquifer, based on tomography studies performed by Calò and Tramelli [68]. Uplift occurs when fluids cannot pass through this impermeable layer and continues until the increase of pore pressure triggers fractures, increases the permeability, and allows fluids to escape, leading to subsidence. Calò and Tramelli [68] not only highlighted the existence of a seismic layer, which separates the shallow aquifer from the deeper part of the caldera, but also concluded that this layer plays an important role in bradyseismic events.

2.4. Stratigraphy, Permeability, and Cooling State. Several wells drilled in the CF area [69–72] for a geothermal exploration program by State Co AGIP-ENEL Joint Venture, and a recent review ([73] and ref. therein) summarizes the crustal structure at CF, revealing both the local stratigraphy and small-scale variability to a depth of ~3 km. Figure 5 shows the stratigraphic sequence in the 3000 m deep San Vito and Mofete wells ([68] and unpublished data). The uppermost 2000 m is composed of recent volcanoclastic products. In the San Vito boreholes (Figure 5(a)), a systematic increase in metamorphic grade with depth is observed

below about 2 km. The thermal profile associated with this metamorphic aureole suggests that magma was present at a shallower depth (4-5 km) in the past compared to today ($\geq$7.5 km). The isotherms (Figures 5(a) and 5(b)) show that measured borehole temperatures are lower than paleotemperature obtained by fluid inclusion studies on core from the San Vito and Mofete wells [68]. The lithological characteristics of the well stratigraphic succession (Figure 5) define a permeability structure with two important characteristics. The first is the anisotropy in permeability resulting from the sedimentary layering, with permeability parallel to bedding exceeding the orthogonal component. The second characteristic is the vertical variation of the mean permeability governed by low-permeability layers [73, 74].

Based on stratigraphic relations, from the surface downward the permeability oscillates from high (coarse clastics and volcanoclastics) to low (transgressive siltstones and claystones) to high (debris flows) to low (marine calcarenites and siltstones) to high (fluvial conglomerates) and finally to low (carbonates, thermometamorphic and plutonic rocks) values. Although the vertical permeability structure has not been measured in situ over the whole section, typical values for similar volcanic, sedimentary, and metamorphic rocks suggest variations in the vertical permeability of five to ten orders of magnitude [75]. In addition, at different depths, isolated aquifers with variable salinity, not related to depth, have been intersected during drilling [69]. Such high salinity is confirmed by fluid inclusion studies of the paleogeothermal system [74]. The latter indicates the poor connectivity between confined aquifers unless compromised by episodes of fracturing in response to transient excursions of fluid pressure. The structure and magnitude of the permeability

field along with episodic fracturing events that *transiently* increase the fracture permeability are an important component of our model to explain bradyseism at CF.

A more recent study [76] on the elastic and mechanical properties of well cores from San Vito (SV1 and SV2) and Mofete (MF1, MF2, and MF5) by high-resolution microstructural and mineralogical analyses reports the existence of an impermeable layer, confirming earlier exploratory drilling studies by the joint venture AGIP-ENEL [69, 71, 72]. Vanorio and Kanitpanyacharoen [76] stress that the deep impermeable layer is a caprock capable of confining the overpressured fluid-bearing underlying formation. The caprock is able to accommodate the strain as fluids accumulate and pore fluid pressure increases until a critical threshold is exceeded (see below). Borehole sections indicate that carbonate country rocks continue to greater depths than the deepest wells.

3. Discussion

3.1. Evolution of a Steady-State Magmatic-Hydrothermal System during Cooling.
The study of well cores showed that the CF volcanic system has been cooling for some unknown period of time that extends to the present time. The pyrometamorphic layer, in the San Vito well, is at 2 km depth, and the thermal profile associated with this metamorphic halo suggests that in the past magma was at a shallower depth of ~4-5 km, compared to its present-day depth of >7.5 km [77]. The cooling and downward migration of isotherms are consistent with the comparison between "fossil" temperatures obtained by fluid inclusions [74] and those obtained by borehole measurements in the geothermal wells of San Vito 3 and Mofete 5. Moreover, Mormone et al. [78] reached the same conclusion by studying two cores, taken at depths of −443 and −506 m, in a 506 m drill hole in the Piana di Bagnoli area located 3 km east of Pozzuoli [26]. Sr isotope ratios, determined on feldspars, and $\delta^{13}C$ and $\delta^{18}O$ on carbonate veins in tuff, along with mineralogical assemblages, give an equilibrium temperature of at least 160°C, much higher than temperatures of 60°C and 80°C measured today at 443 m and at the bottom of the well, respectively.

3.2. A Model for the Cyclicity of Uplift and Subsidence Events.
The results of ground deformation and microgravimetric data models applied at CF [35, 36, 41, 43, 51, 52, 55, 59, 79] give different densities for the bodies causing the deformation because the crust is treated, for the most part, as an isotropic, homogeneous, and elastic medium. The parameterization of these deformation models, including the choices made for the elastic properties and layer thicknesses, leads to inferred densities that vary considerably. At CF, the presence of structural discontinuities and/or elastic heterogeneity, viscoelasticity, and the anisotropy in permeability [73] can confound any simple inversions to obtain densities. In addition, at CF, there are differences in displacement ratios and between uplift and subsidence phases that clearly show that simple models based on homogeneous and elastic crust are gross simplifications of a complex geological environment.

As described by Lima et al. [14], bradyseism results from the complex interplay of two dominant processes operating on very different timescales. The long (10^3–10^4 yr) timescale process is associated with the deep, steadily cooling magmatic system (>7.5 km deep) with the concomitant migration downward of the solid–melt boundary of the mush zone of the crystallizing magma body. The short timescale (1-10^2) process is associated with the hydrothermal system beneath an impermeable layer B that episodically undergoes failure by fracture (seismicity), connecting the lower lithostatic reservoir with the upper hydrostatic aquifer. Bodnar et al. [13] present a quantitative conceptual model describing this process, and Becker et al. [80] describe fluid-related processes in this environment. Thus, bradyseism is attributable to the variations of the short timescale system related to the confined aquifer. Figure 6 summarizes bradyseism phases during the transition from magmatic to epithermal conditions in a subvolcanic environment based on the model of Fournier [81]; for more details, see Bodnar et al. [13] and Lima et al. [14]. In 1982, unrest started along with low magnitude seismicity (Figure 6(a)), associated with vesiculation (or second boiling; [82]) of residual magma that generated magmatic fluids that remained geopressured (lithostatic) (isobaric crystallization) or even super-geopressured (isochoric crystallization) until fracture propagation enabled the expulsion of magmatic fluids (Figure 6(a)). Unrest continued throughout 1983 with an increasing uplift rate and intense seismicity, often clustered in swarms [83] (Figures 3(a) and 3(c)), due to the geopressured magmatic fluids breaching the brittle carapace (Figure 6(b), layer A), marking the brittle-plastic transition zone (Figure 6(b), see also Becker et al. [80]). Magmatic and hydrothermal fluids mix but remain in a lithostatic pressure regime because they are confined by the relatively impermeable layer B (unit B, in Lima et al. [14]). At this stage, unrest reached the ground deformation peak (Figure 6(b)). At the end of 1984 (Figure 6(c)) after an uplift of 1.79 m, hydrofracture propagation in layer B enables the mixed magmatic and hydrothermal fluids to escape, causing intense seismicity, clustered in swarms, at more superficial levels (Figure 6(c)) [6]; at this stage, the fluid transitions irreversibly from lithostatic to hydrostatic conditions with the expansion of fluids and boiling. The solubility of aqueous silica decreases significantly along this path, leading to silica precipitation and an episode of permeability loss by fracture sealing. Thermodynamic and transport properties of aqueous geofluids change rapidly and dramatically in regions of pressure-temperature-composition space around the critical point of H_2O [84, 85], as seen in geothermal well isotherms at a depth of ~3 km, where the temperature is around 400°C. At the end of 1984, subsidence started and seismicity decreased. At this stage (Figure 6(c)), chemical-physical variations involve the system at a large scale, and fracture healing closed most of the numerous hydrofractures associated with ~16,000 earthquakes recorded during the 1982-84 bradyseism event. Since 1984, subsidence slowly continued for about 22 years, likely because layers A and B were not completely sealed, with magmatic gases still being detected in the Solfatara fumaroles even if they gradually decreased over time (Figures 4(b) and
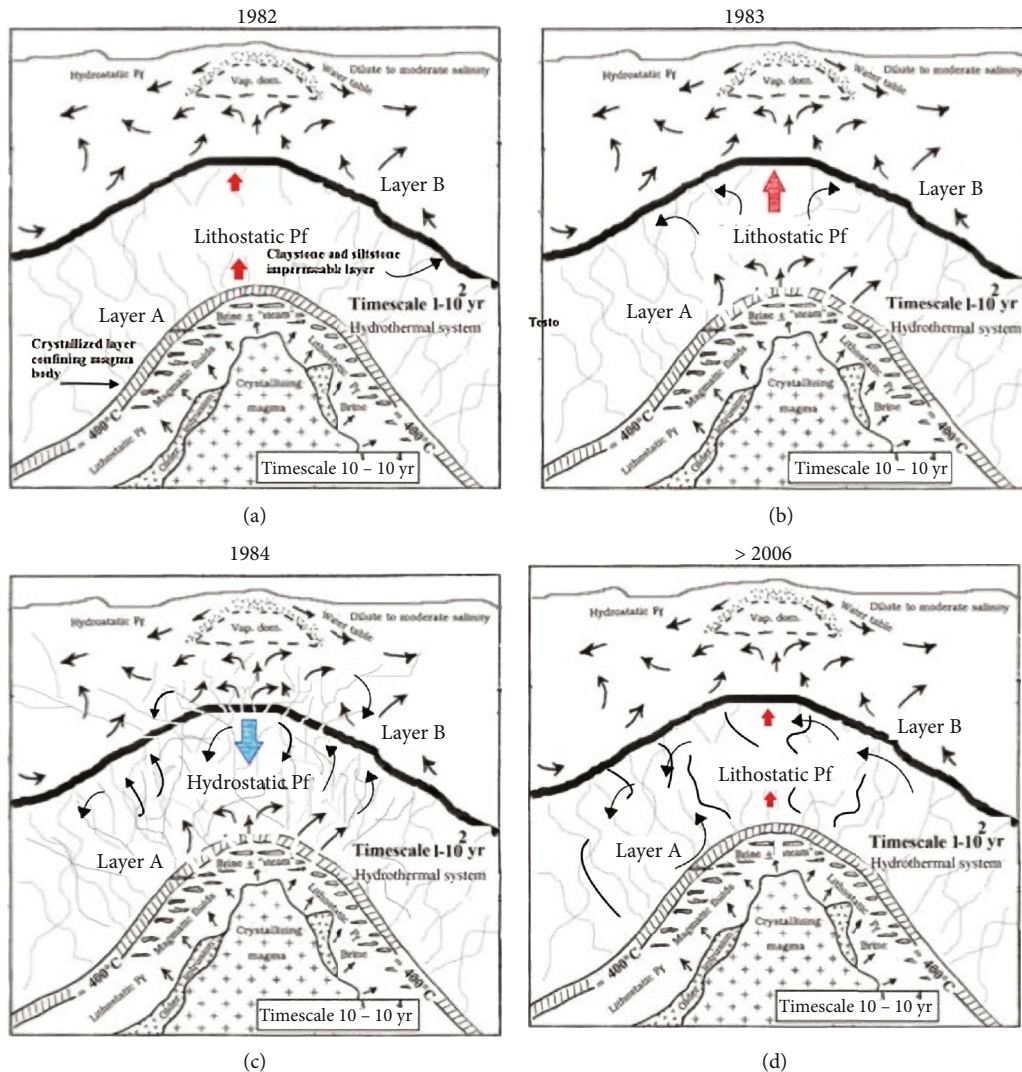
FIGURE 6: (a) In 1982, uplift started: long timescale magmatic system was sealed by a low permeability carapace (layer A) which blocks hot magmatic fluids that accumulate; short timescale hydrothermal system was closed by low permeability layer (layer B). (b) In 1983, uplift continues: the carapace (layer A) breaks allowing magmatic fluids to enter the overlying confined hydrothermal system. (c) In 1984, impermeable layer B breaks as well triggering subsidence. Pressure regime changes producing faulting, brecciation, hydrothermal alteration, boiling, and silica deposition. Subsidence continues until layer B close by self-sealing. (d) Since 2006 until today, layer B is sealed, and uplift goes on (modified from Fournier [81]).

4(c)). When the system is again completely sealed and the deeper and more shallow aquifers are no longer connected, subsidence stops even if it does not balance uplift completely because the deformation at the macroscopic scale is highly inelastic. Since the end of 2005 (Figure 6(d)), layer B has been sealed, and the vertical deformation recorded at the surface has slowly resumed. Bradyseism is governed by the impermeable layer B, a caprock formed by fibrous minerals and intertwining filaments that confers shear and tensile strength, contributing to its ductility and increased resistance to fracture [76]. The presence of this layer has been highlighted also by tomographic studies of Calò and Tramelli [68]. Following compression/decompression, layer B can act as a mechanical toggle such that connectivity between hydrostatic and lithostatic reservoirs is episodically established. This process is a common mechanism associated with ore deposition in the porphyry systems (see Lima et al. [14], for more details). When connectivity is established and fluid expansion occurs, the temperature increase associated with fluid decompression will lead to a decrease in the solubility of silica in the fluid [86].

Earthquake swarms during bradyseism should be the manifestation of hydrofracturing and crack propagation which act as conduits or paths of egress for aqueous fluid during decompression (volumetric expansion of fluid) [87–89]. In the future, we anticipate that new and evolving methodologies will be applied at CF to better distinguish between different sources of seismicity. As an example, Butcher et al. [90] applied statistical and quantitative approaches to identify a stable source of long-period (LP) seismicity at Cayambe Volcano, Ecuador, that is likely associated with a shallow hydrothermal system.

Minor episodes of bradyseism have occurred every 5-6 years since 1989 (Figure 2), and associated seismicity, both during uplift and following subsidence, is explained in the context of the short timescale process. At the end of 2005, a new unrest phase started. Compared with 1982, the uplift rate is much slower; in 15 years, the total uplift is ~75 cm, and seismicity is moderate, of low magnitude and with shallow epicenters (Figures 3(a) and 3(c)). It is possible that after 40 years, bradyseism is attenuating, as hypothesized by Lima et al. [14]. However, there is also the possibility that fluids escape from the system through the numerous existing fractures and faults in the CF area (Figure 1(c)) as in the nearby Solfatara-Pisciarelli active areas (Figure 3(b)). In this way, the energy for unrest is reduced and the cooling of the magmatic system is accelerated, pushing the magma-host rock mushy contact zone deeper through time as heat advection by magmatic-hydrothermal migration takes place.

Figure 4 shows the variation in the composition of fumarolic gases recorded at Bocca Grande (BG) and Bocca Nuova (BN) at the nearby Solfatara volcano; both N_2/CO_2 (Figure 4(b)) and N_2/He (Figure 4(c)) ratios show a decreasing trend since 1984, as it should be if the connectivity between the two systems (magmatic and hydrothermal) is decreasing and considering that N_2 and He are likely high-temperature magmatic gases; after 2006, the decrease is very slow. The current N_2/CO_2 ratio is similar to the one measured in 1982 when unrest started. CO_2 can have different origins since volcanic, magmatic, and hydrothermal gases are dominated by water and CO_2. In addition, at CF, CO_2 can be generated by decarbonation reactions in the wall-rocks as well as in the CF basement (see Figure 3 in Lima et al. [14]). The H_2O/CO_2 ratio (Figure 4(a)) decreased rapidly in 1984 when subsidence started. These variations were interpreted [14] to reflect magmatic fluid contribution from the deep magmatic system that follows hydrofracking of both impermeable A and B layers (Figure 6(c)). Afterwards, variations in the H_2O/CO_2 ratio seem to be related to minor unrest events. After 2006, when new unrest started, the H_2O/CO_2 trend decreases more slowly. Increasing H_2O/CO_2 trend is one of the greatest concerns of researchers who interpret it as a signal of a probable eruption (e.g., [57, 60, 63]). Chiodini et al. [66] investigated relationships among hydrothermal temperature-pressure, fluid flow, and earthquakes at CF (always, based on the assumptions that (a) the caldera was formed by CI 39 ka ago—in contrast with the evidences of de Vivo et al. [21], Rolandi et al. [22, 23], and De Natale et al. [26, 27]; (b) bradyseism is caused by magmatic intrusions; and (c) an eruption is imminent) and interpreted the impact of hydrothermal fluid P-T changes on earthquake occurrence and fluid emissions. An increase of the CO_2 emission at the surface reflects an increase in the amount of magmatic fluid entering the base of the hydrothermal system, leading to heating and pressurization. The latter contrasts with unrest cyclicity that explains uplift by the episodic opening and closing of a multitude of fractures related to the transient switch between lithostatic and hydrostatic conditions on a short (1-10 yr) timescale.

When considering magmatic and hydrothermal contributions in fumaroles, it is useful to compare the enriched gases in magmas. Both N_2/CO_2 (Figure 4(b)) and N_2/He (Figure 4(c)) do not show any variations. H_2O/CO_2 ratios of magmatic and hydrothermal gases are within the same range [91]. At Solfatara, deeply derived CO_2 commonly occurs in specific areas such as faults, small vents, and steaming ground rather than across the entire volcanic system [91]. Locally at Solfatara-Pisciarelli, changes in CO_2 can be caused by activation of fractures (Figure 3(b)), most likely affecting the carbonates, and low seismicity and tremors at shallow depth can be the result of fluid boiling [88, 89]. On the other hand, in consideration of the complex stratigraphy, permeability, and resistivity [92], a confined aquifer lens locally can also be present. The latter can undergo fluid overpressure for temperature increase, hydraulic fracturing, and closure by silica precipitation in response to phase transition from liquid to vapor at 20 MPa [86]. Likely, this is the case reported for two episodes of seismicity and gas emission that occurred on 7 October 2015 and 6 December 2019 at Solfatara-Pisciarelli area [93], with a sudden increase in the fumarolic tremor amplitude observed during the seismicity episode. The uplift rate decreased immediately after the swarm whereas the fumarolic tremor amplitude remained higher than that observed prior to the swarm. Recently, it has been observed also that shallow volcanic tremors can be related to intense rainfall in the CF area [94].

The model for deformation induced by the Thermo-Poro-Elastic (TPE) inclusion reported by Nespoli et al. [67] shows similarities with our model by attributing bradyseism to the presence of a layer with time-dependent permeability capable of episodically preventing the rise of fluids to the surface. In any case, they do not specify a magma role nor if it is in a steady and cooling state and when and why magmatic fluid flow to the surface increases and why bradyseism shows cyclicity over time.

3.3. Bradyseism Signature. At least since the 1970s, when the first vertical leveling was done at CF, the bradyseism ground deformation pattern has remained invariant: a bell shape curve (Figure 7) both during uplift and subsidence episodes with higher rate during uplift, and the maximum uplift centered near Pozzuoli [7, 33, 52, 95].

The constancy of the deformation area along with the constancy of the seismogenic volume during unrest can be considered to represent a "bradyseism signature." The latter involves cyclical events in the same spatial-temporal pattern as discussed above. Magmatic intrusions cannot meet these requirements; they generally follow fractures or paths of less resistance to ascend, it is unlikely each magma pulse would give the same ground deformation pattern again and again, and the magma injection model cannot explain the subsidence following the uplift without very ad hoc and unrealistic assumptions. In addition, the density model built (Figure 8) using a new 3-D inversion of the available high-precision gravity data, and a new digital terrain and marine model [96] characterizing the shallow caldera structure (down to 3 km depth), highlights a pronounced low-density region in the central portion of the caldera, incompatible with the presence of magmatic intrusions. Moreover,
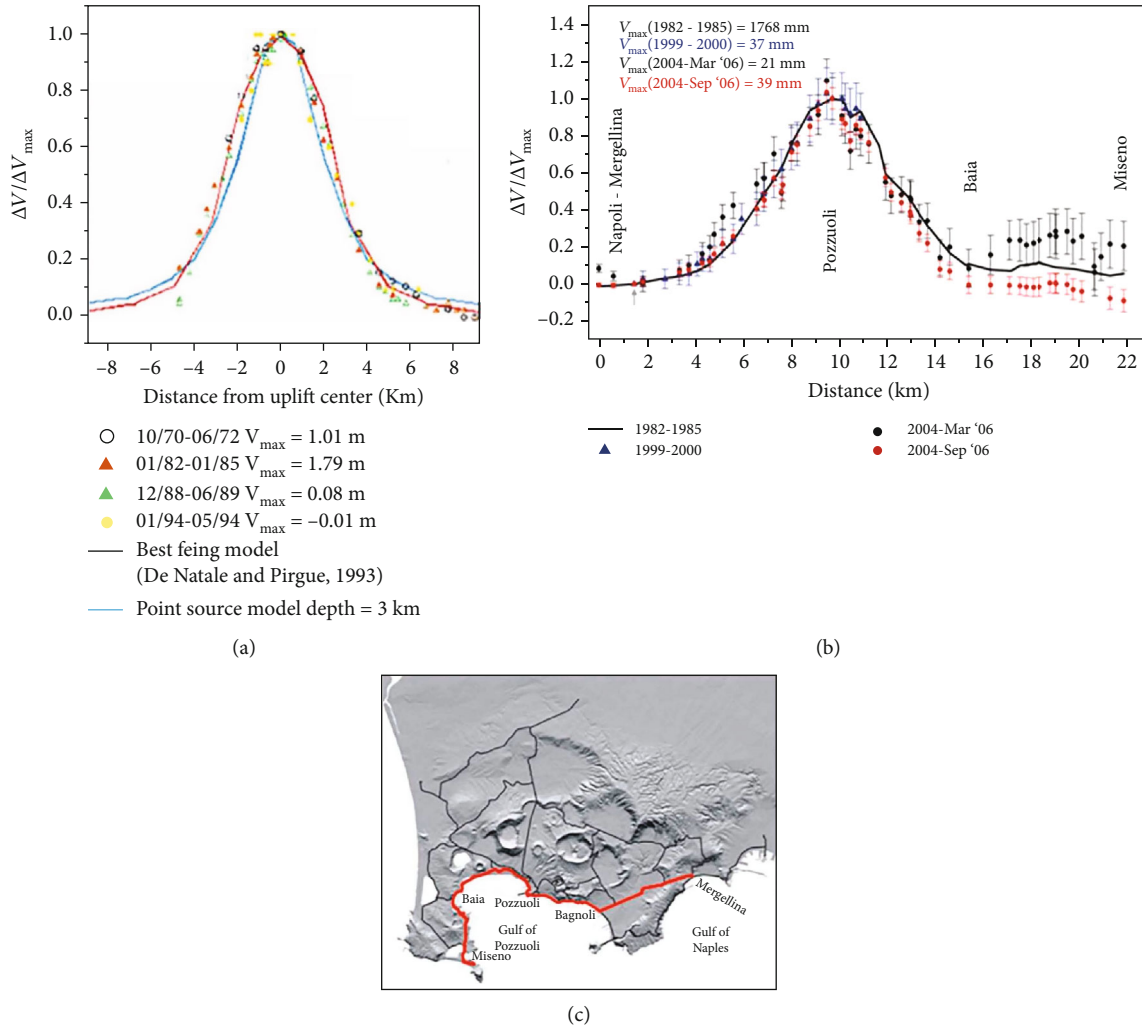
(a)

(b)

(c)

FIGURE 7: (a) Normalized absolute value of ground displacement data shown as a function of the distance for four different unrest periods (De Natale and Pingue [7]). (b) Normalized absolute value of ground deformation measured by precision levelings (along the red line in (c)) from 1982 to 2006. (c) The main leveling route from Naples to Miseno.

during drilling of 11 geothermal wells (AGIP-ENEL joint venture), which reach a depth of 3.2 km, no magmatic intrusions were found. Trasatti et al. [46] argue that the presence of seismicity to a depth of 4.5 km and the relatively cold temperatures ~420°C encountered during drilling to ~2.7 km depth at CF are difficult to reconcile with the presence of magma at ~3 km depth.

Bradyseism cyclicity requires as a prerequisite a stable geological structure capable of preserving the hydrothermal system without substantial variations for centuries. As illustrated in Figure 1(b), an anticline with a low permeability cap layer is present at a depth of ~3 km. The latter has a pozzolanic composition and a fibril-rich matrix as a result of lime-pozzolanic reactions with formation of fibrous minerals contributing to its ductility and increased resistance to fracture [76]. The caprock restricts the vertical ascent of fluid until a transient and short-lived hydrofracturing episode develops, acting as the bradyseism trigger. The dramatic changes in the thermodynamic properties of H_2O near the critical point [84] in conjunction with irreversible

decompression from a lithostatic to a hydrostatic pressure regime facilitate the deposition of silica and the sealing of fractures in a cyclic manner. The active intracaldera compressional tectonic regime explains both anticline formation and fluid accumulation as illustrated in Figure 6. Earthquakes associated with bradyseism are located close to the town of Pozzuoli and in the middle of the bay near the focus of the tectonic folding (Figure 3(a)). The maximum depth of earthquakes is located at the stratigraphic boundary between the carbonates of the Mesozoic–Cenozoic succession and the overlying stratified clastic succession of Pleistocene age (see Figure 3, in Lima et al. [14]).

Identification of the "bradyseism signature" can be useful in distinguishing events that deviate and could be of greater concern for risk prediction. The 1538 Monte Nuovo eruption did not show the bradyseismic signature, even if it was preceded by about 7 m of uplift. Ground deformation and the maximum uplift were not centered at Pozzuoli but on the site of the eruption (~3 km west of Pozzuoli) [3, 4].
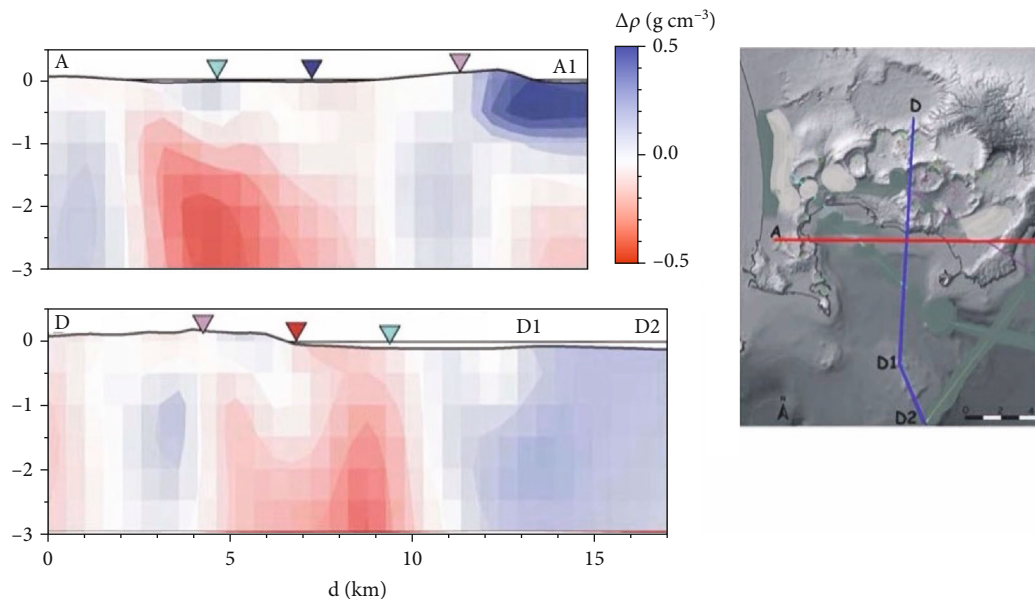
FIGURE 8: Selected vertical density cross-sections. Locations of the plotted sections are shown in the small map on the right. Color scale represents density contrasts (g/cm³) (modified from Capuano et al. [96]).

3.4. Comparison to Other Caldera Systems. Caldera unrest poses a challenge for eruption forecasting because it may persist for weeks or centuries at large calderas and ground deformation and gravity changes must be correctly interpreted for hazard evaluation. It is critical to differentiate variations of geophysical observables related to volume and pressure changes induced by magma migration from shallow hydrothermal activity associated with hot fluids of magmatic origin rising from depth. To discriminate signals generated by hydrothermal perturbations from those related to magma movement towards the surface is not easy, few ground surface displacement (GSD) models account for significant complexities such as heterogeneities in hydrological and mechanical properties of matrix and geological features like faults, which might influence both the path of ascending magma and the subsurface circulation of hydrothermal fluids. As observed in several large calderas, hydrothermal fluid injection, circulation, and gas formation can generate complex and temporally and spatially varying GSD patterns of deformation rates, magnitudes, and geometries. Cocco et al. [97] proposed a numerical model for evaluating the variations in geophysical parameters associated with the perturbation of the hydrothermal system, using the CF caldera as an example. Throughout simulations of the hydrothermal system dynamics, they observed that heterogeneities in hydrological and mechanical properties as well as the presence of faults within caldera forming volcanoes substantially affect the hydrothermal circulation of hot fluids and the consequent variation in geophysical signals. These results indicate that it is very unlikely that cyclical events in a restless caldera can give rise to the same geophysical signals repeatedly over time as seen for CF "bradyseism signature." Unrest events occurred within several large calderas, including Rabaul, Aira and Iwo-Jima (Japan), Long Valley (California, USA), Yellowstone (Wyoming, USA), Kilauea, and Mauna

Loa (Hawaii, USA) ([98] and ref. therein). In these calderas, we are not aware of invariant ground deformation patterns both during uplift and subsidence episodes with higher rate during uplift, and the maximum uplift centered always in the same restricted area, for many different deformation events. For example, at Yellowstone caldera, geodetic measurements that now include semipermanent Global Positioning System (GPS), continuous GPS, and interferometric synthetic aperture radar (InSAR) have shown the entire caldera floor to be mobile in space and time [99, 100] even if observations of cyclical crustal deformation are key to evaluating the hazards of this active volcanic system and to improve understanding of the relation between time-dependent deformation and magma migration and help to differentiate between hydrothermal and magmatic sources [101].

4. Conclusions

A decade ago, the authors proposed a magmatic-hydrothermal model to explain bradyseism at CF, since over the years the latter has not lost its validity, we reviewed the abundant literature which in the meantime has been published to corroborate our model and to demonstrate that the recent CF bradyseism events are related to cyclical magmatic-hydrothermal processes associated with cooling and crystallization of epizonal, water-bearing magmas. Firstly, we emphasize that bradyseism displays repetitive cyclical signals without substantial variations. At CF, a bell shape ground deformation pattern in which the magnitude of uplift and subsidence decay rapidly with distance from the location of maximum uplift/subsidence defines the "bradyseismic signature" (Figure 7). A constancy is also noted for the seismogenic volume during a bradyseismic episode. The maximum uplift is always centered at Pozzuoli. If

bradyseism always gives the same signature, it is logical to expect that the generating mechanism is the same and repetitive over time. The models that explain bradyseism due to shallow magmatic intrusions do not support these characteristics that are invariable over time. The results of both seismic tomography [68, 77] and the density model built using a new 3-D inversion of the available high-precision gravity data [96] also exclude the presence of magma at shallow levels. "Bradyseismic signature" can be very helpful to distinguish bradyseism from impending eruption precursors. Secondly, by combining the cyclical model with results of other investigations, we show that fresh magma injection is not a prerequisite for bradyseism to occur and that, in the last decades, bradyseism has not been accompanied by shallow magmatic intrusions.

Finally, we highlighted that the cyclic nature of the observed bradyseism is closely related to the complex geological, tectonic, and stratigraphic upper crustal structures that govern spatial and temporal variations of the subsurface permeability. Bradyseism should be governed by the impermeable layer B that is part of an anticlinal geological structure capable of preserving the hydrothermal system without substantial variations for centuries, acting as the throttling valve between the upper aquifer and the deeper hydrothermal short timescale system ($1\text{-}10^2$ yr) under alternatively lithostatic/hydrostatic pressure. Connectivity between hydrostatic and lithostatic reservoirs is episodically turned on and off causing alternatively uplift and subsidence depending on whether permeability by fractures is established or not. Earthquake swarms should be the manifestation of hydrofractures that allow fluid expansion and silica precipitation sealing cracks and isolating again the reservoir. Since 2006, unrest starts again slowly, showing a "bradyseism signature" with an uplift of ~75 cm in about 15 years, possibly due to the fluids escaping from the system through fractures as evidenced by the increased Solfatara and Pisciarelli fumarole flows (Figures 1(c) and 3(b)). In this way, the energy for unrest is reduced and the cooling of the magmatic system would also be accelerated, pushing the liquidus deeper and deeper. Lima et al. [14] applied rigorous thermodynamic constraints to evaluate ground deformation at CF and determined that if the system were completely sealed, a maximum uplift of 40 m for the entire CF would be feasible. In active volcanic environments similar to CF, uplifts of similar magnitudes associated with hydrothermal fluids have been reported previously. As an example, Wicks et al. [100] reported 28 inches (0.71 m) of uplift over a 5-year period associated with deep hydrothermal fluids at the Yellowstone caldera.

The miniuplift episodes are well explained by the short timescale process. The cyclicity is also highlighted comparing magmatic and hydrothermal contributions in Solfatara-Pisciarelli fumaroles; both N_2/CO_2 (Figure 4(b)) and N_2/He (Figure 4(c)) variations and N_2/CO_2 ratio in Solfatara fumarolic gases returned to the same values recorded in 1982, when the system was in the same closed condition as today.

According to the model we propose, at present, the probability of an eruption at CF should be the lowest that it has been in the last 500 years and should be expected to decrease further over time because of cooling of the magmatic system. The scenario could change only if new magma arrives from greater depths and supplies the CF deep magma chamber. In this case, the ground deformation pattern should change radically compared with the typical "bradyseism signature."

Conflicts of Interest

The authors declare that they have no conflicts of interest.

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