A Re-interpretation of Long-Term Deformation at Campi Flegrei caldera, Italy and Perceptions of the Causes of Caldera Unrest

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Where information has been derived from other sources, I confirm that this has been indicated in the thesis.'

Signed: .... Date: ....

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Abstract

Campi Flegrei caldera has been in a state of unrest since 1950. Between 1950-1984 the centre of the caldera was raised c. 3.8 m by three rapid uplifts (c. 1 m yr^{-1}). The third episode was followed by subsidence until 2004 (c. 0.9 m), when a slow uplift began that continues to Present (2019). The causes of the deformation are debated but common to all conventional models is that they cannot account for the change in the characteristics of deformation after 1984. This research focuses on identifying a potential cause for the change and on understanding the perceived role of the hydrothermal system in ground movements amongst scientists.

By combining the results of a review of the caldera's magmatic-hydrothermal system and behaviour with an analysis of the distribution of seismicity in relation to hydrothermal reservoirs, new model constraints were defined and deformation trends reinterpreted. Perceptions of unrest were investigated through a survey of 62 Italian scientists. The primary result is a new model for deformation that considers ground movements since 1950 to represent a single evolutionary sequence. It was recognised here that conditions in 1984 were favourable for an increase in bulk permeability in hydrothermal reservoirs below the deforming area as the crust was progressively fractured and faulted over successive uplifts. Based on this observation, post-1984 ground movements are attributed to a redistribution of pore pressure as reservoirs continue to adjust to the mechanical changes in the crust. Amongst surveyed scientists, there was a general perception that for the hydrothermal system to contribute to uplift it must be pressurised by either a magma intrusion or an injection of magmatic fluid, and that this is a pre-requisite in order for subsidence to occur as a result of pore pressure loss.

The model can act as an end-member scenario for evaluating the evolution of uplift, whilst improved understanding of the perceived controls on deformation is a step towards improving communication of unrest.

Impact Statement

Campi Flegrei caldera in Italy has a population of 360 000 people and is generally considered to be one of the highest risk volcanoes in the world. Since 1950, it has been in a state of unrest characterised by four episodes of caldera-wide uplift. Between 1950-1984 three short-term episodes of uplift (2-3 years) cumulatively raised the town of Pozzuoli, in the centre of the caldera, by c. 3.8 m and repeatedly triggered evacuations of up to 40 000 people causing significant socio-economic impacts. Uplift in 1982-1984 was followed by a period of subsidence. Then in 2004 the most recent episode of uplift began, which continues to Present (2019). Similar behaviour is known to have occurred in the c. 100 years before the last eruption at Monte Nuovo in 1538 and suggests a reactivation of the magmatic system. There is thus a social need to improve the understanding of the processes of unrest to reduce the potential for false alarms and failed forecasts in the future.

Unrest is caused by the interaction of pressurised fluids and the crust. A long-standing debate at Campi Flegrei as to the causes of uplift has centred on whether they result from the pressurisation of magma, the hydrothermal system, or a combination of both. This thesis builds on previous work and uses long-term trends in monitoring data to define a new model for unrest since 1950, that can account for the full sequence of ground movements for the first time. It considers deformation after 1984 to be caused by pore pressure changes in the hydrothermal system triggered by the exceedance of a critical threshold of fracturing due to stretching of the crust by successive intrusions of magma between 1950-1984. As such, ground movements since 1950 can be considered to represent a single long-term evolutionary sequence of conditions in the crust. The thesis also evaluated the extent to which the academic debate as to the causes of ground movements exists amongst scientists who may be involved in the scientific operational response to unrest.

The results of the thesis have practical applications in the improved interpretation of unrest at Campi Flegrei. As a result, they can lead to (i) more realistic hazard assessment, (ii) improved forecasts of the evolution of an unrest episode and whether or not it is likely to end in eruption, and (iii) improved hazard communication. They can also form the basis of the definition of future unrest scenarios that can be used for developing strategies for caldera unrest management and emergency planning.

Finally, long-term sequences of unrest (~10-100 years) before intra-caldera eruptions are common at large calderas globally. The results for Campi Flegrei can therefore provide a benchmark for evaluating the roles of the magmatic and hydrothermal systems in unrest at large calderas in general.

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Abbreviations

AGIP Azienda Geologica Italiana Petroli
AMS Agnano Monte Spina
BDT Brittle-Ductile Transition
BET Bayesian Event Tree
BG Bocca Grande
BN Bocca Nuova
CFDDP Campi Flegrei Deep Drilling Project
CI Campanian Ignimbrite
CNR Consiglio Nazionale delle Ricerche
COTs Commercial Off The shelf components
DCP Dipartimento della Protezione Civile (Department of Civil Protection)
DDS Diffuse Degassing Structure
DRE Dense Rock Equivalent
ENEA Italian National Agency for New Technologies, Energy and Sustainable Economic Development
ENEL National Electric Agency
FC Fumarola Circolare
IGGI Istituto di Geochronologia e Geochimica Isotopica Pisa
IGM Italian Military Geographic Institute
INGV Istituto Nazionale di Geofisica e Vulcanologia
INGV-OV Istituto Nazionale di Geofisica e Vulcanologia - Osservatorio Vesuviano
IoT Internet of Things
Irea Istituto per il rilevamento elettromagnetico dell'ambiente (Irea)
NDIR Non-Dispersive Infra-Red
NYT Neapolitan Yellow Tuff
RFZ Ring Fault Zone
SAFEN Società Anonima Forze Endogene Napoletane
TDS Total Dissolved Solids
UCL University College London
UWI University of the West Indies
VAL Volcano Alert Level
VT Volcano-Tectonic

Chapter 1

Introduction

Large calderas with diameters of 5 km or more are among the highest risk volcanoes on Earth (Acocella et al., 2015). Following their formation, they frequently become the sites of intra-caldera eruptions that are commonly preceded by ~10-100 years of unrest, characterised by episodic uplift and seismicity (Newhall and Dzurisin, 1988; Acocella et al., 2015). Long-term trends in ground deformation suggest that episodes may belong to a single evolutionary sequence where the probability of an eruption increases over time (Newhall and Dzurisin, 1988; Robertson and Kilburn, 2016; Kilburn et al., 2017). However, individual uplifts may result from the pressurisation of magma (e.g. McKee et al., 1984; Dzurisin et al., 1990; Wicks et al., 2006; Chang et al., 2007; Parks et al., 2012; Montgomery-Brown et al., 2015; Tizzani et al., 2015), magmatic or hydrothermal fluids (e.g. Dzurisin et al., 1990; Dzurisin et al., 1999; Caliro et al., 2004; Chiodini et al., 2015; Hildreth, 2017; Moretti et al., 2017), or some combination thereof (e.g. Gottsman et al., 2006; Woo and Kilburn, 2010; Hutchinson et al., 2016). Purely magmatic or hydrothermal fluid sources do not increase the likelihood of a magmatic eruption. A critical challenge for hazard assessment is identifying whether a discrete episode will evolve into an eruption.

Campi Flegrei caldera, Italy has a population of c. 360 000 people and is in a state of unrest that began in 1950. Since this time, it has undergone four episodes of uplift (1950-1952, 1969-1972, 1982-1984 and 2004-Present) that have elevated the centre of the caldera by c. 3.6 m and repeatedly triggered the evacuations of up to 40 000 people (Barberi et al., 1984). Debate exists as to the causative processes of the uplifts and the relative contributions of the magmatic and hydrothermal systems during each episode (e.g. Gottsman et al., 2006; De Natale et al., 2006; Moretti et al., 2017). The proposed mechanisms have conflicting implications for hazard (e.g. Casertano et al., 1976; Corrado et al., 1977; Bianchi et al., 1987; Orsi et al., 1999a; De Vivo and Lima et al., 2006; Woo and Kilburn, 2010; Chiodini et al. 2015a; Moretti et al., 2017) and the development of scenarios of future unrest. A step towards reducing the likelihood of false alarms and failed forecasts in the future, is understanding the role of the hydrothermal system in unrest. This thesis examines how the potential contribution of the hydrothermal system to ground movements may change over time in the context of a long-term unrest sequence, and the perceptions of its role in unrest amongst scientists who may be involved in the operational response in the event of future volcanic crisis.

1.1 Caldera-hosted Hydrothermal Systems

Calderas are complex volcano-tectonic structures formed by subvertical subsidence into a magma reservoir that has been drained during an eruption, or by subsurface flow along intrusions (Scandone, 1990; Branney and Acocella, 2015). Large calderas, such as Campi Flegrei are produced by one or more eruptions expelling more than 10 km^3 DRE (Dense Rock Equivalent) of material (Walker, 1984; Newhall and Dzurisin, 1988; Cole et al., 2005). The resulting topographic depression is approximately sub-circular, bounded by ring faults that accommodate the subsidence, and typically filled with pyroclastic material. The morphology is controlled by the depth and geometry of the magma reservoir, the amount of subsidence, pre-existing structural discontinuities, rock strength and the stress field (Lipman, 2000; Cole et al., 2005; Acocella, 2007). Following collapse, intra-caldera eruptions often occur at intervals of 10^1-10^3 years and are usually located along faults, especially at the caldera margins (Lipman, 2000).

Commonly, large calderas host long-lived (10^3-10^6 years) hydrothermal systems, the surface manifestations of which may include; hot springs, mud pots, fumaroles and geysers (Cathles et al., 1997; Hochstein and Browne, 2000; Branney and Acocella, 2015). They develop in fractured and faulted volumes within the caldera, often around the collapse margins, where permeability is sufficient for groundwater to circulate to depths where it becomes heated due to the presence of an underlying magma reservoir that is usually located at depths greater than 5 km (Wohletz and Heiken, 1992; Stimac et al., 2015; Branney and Acocella, 2015; Garden et al, 2017). As the fluids are heated, they become buoyant and move upwards relative to the hydrostatic gradient as a hydrothermal upflow. Under quiescent conditions the fluids transport heat and mass towards the surface at a rate determined by the buoyancy force and the permeability of the host rock (Elder, 1981; Norton, 1984).

High temperature and chemical gradients caused by hydrothermal fluid flow commonly result in the alteration of the host rock to characteristic alteration assemblages that reflect local temperature, permeability and chemical conditions (Browne, 1978; Henley and Ellis, 1983; Norton, 1984). The different assemblages can be divided into the argillic, chlorite-illite (phyllitic transition), Calc-Aluminium silicate (propylitic) and thermo-metamorphic alteration zones (Fig. 1.1). The argillic and chlorite-illite zones are characterised by clay minerals and form a low permeability caprock that isolates the main hydrothermal circulation from surface groundwaters. The hydrothermal reservoirs typically broadly coincide with the Calc-Aluminium silicate zone where the alteration, in combination with the precipitation of minerals from circulating fluids (e.g. carbonate and silica) causes the lithology to become brittle (Stimac et al., 2015). In this region fractures can be maintained, creating sufficient permeability for convection ($\geq 10^{-16}$ m^2, Hayba and Ingebritsen, 1997; Manning and Ingebritsen, 1999). Fluids are dominantly meteoric, near neutral-pH and Na Cl rich Henley and Ellis, 1983; Nicholson, 1993; Jasim et al., 2019). They circulate at temperatures of c. 220-350 °C (e.g. Rosi and Sbrana, 1987; Rowland and Sibson, 2004; Caliro et al., 2005), at close to hydrostatic pressures (Fournier, 1999; Stimac, et al., 2015). H_2O is usually in the liquid phase, although vapour-dominated regions can develop where temperatures are sufficiently high and the upflow is isolated from surrounding groundwaters (Ingebritsen and Sorey, 1988; Goff and Kanik, 2000). The reservoir is recharged by distal meteoric waters towards the base of the system where they are heated and mix with magmatic fluids (i.e. volatiles released from cooling or crystallising magma bodies). The limit of meteoric circulation generally coincides with the 400 °C isotherm, where there is a rapid loss of permeability due to mineral precipitation that seals fractures (Fournier, 1987: Fournier, 1991). This results in an increase pore pressures up to lithostatic values limits fluid flow so that heat transport becomes dominated by conduction rather than advection (Manning and Ingebritsen, 1999). In silicic systems this temperature also coincides with the Brittle-Ductile Transition (BDT), that causes a further loss of permeability (e.g. Fournier, 1999). A general conceptual model of the structure of caldera hydrothermal systems is given in Figure 1.2.

Throughout this thesis the term hydrothermal fluids will be used in reference to mixtures of meteoric and magmatic fluids, whilst the hydrothermal system or hydrostatic pore pressure regime will be used to describe the region of the crust where hydrothermal fluids flow at hydrostatic pressures. The terms magmatic system and magmatic pore pressure regime refer to the underlying crust where fluids are dominantly magmatic and at lithostatic pressures.

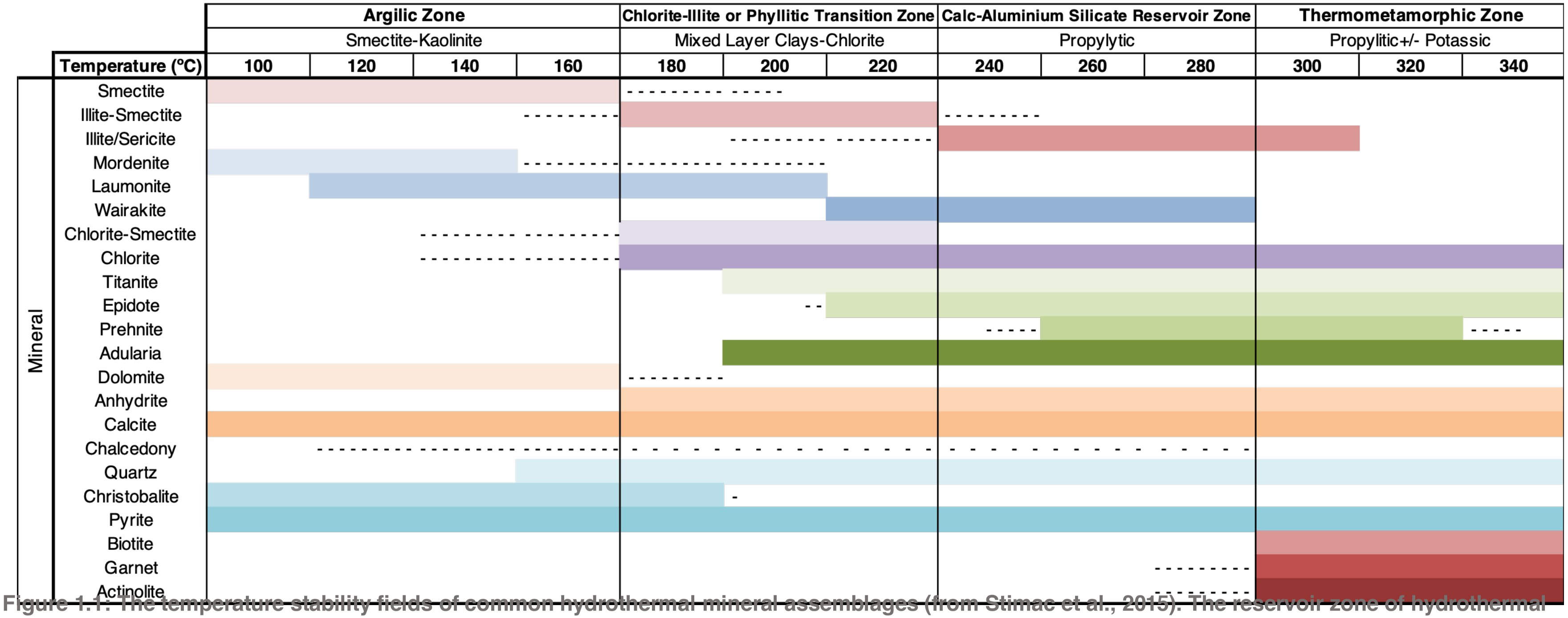


Figure 1.1: The temperature stability fields of common hydrothermal mineral assemblages (from Stimac et al., 2015). The reservoir zone of hydrothermal systems typically coincides with the calc-aluminium silicate, or propylitic zone. Alteration of this causes the lithology to become brittle, allowing fracture-controlled permeability to develop and fluid flow.

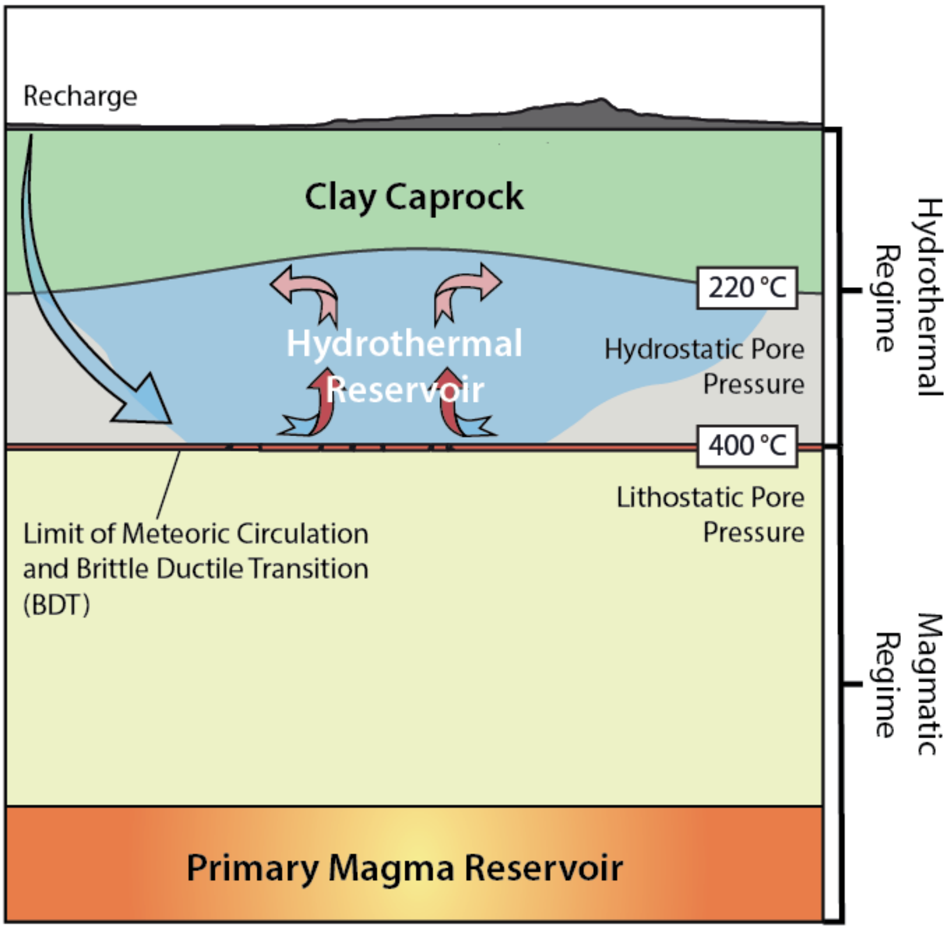


Figure 1.2: General conceptual model of a caldera hosted magmatic-hydrothermal system. The primary magma reservoir is the heat source that drives convection in the hydrothermal reservoir, which is located between a low permeability zone that isolates it from the underlying magmatic system, and a clay-rich caprock formed by hydrothermal alteration.

1.2. Caldera Unrest

Volcano unrest is "*the deviation from the background or baseline behaviour of a volcano towards a behaviour which is a cause for concern in the short-term because it might prelude an eruption*" (Phillipson et al., 2013). Episodes occur at around 20 calderas in a given year, vary in duration from hours to years, and are more likely to be non-eruptive than pre-eruptive (Newhall and Dzurisin, 1988; Phillipson et al., 2013; Acocella et al., 2015). Most frequently, unrest is recognised from ground deformation (e.g. uplift, subsidence, tilt) and swarms of volcano-tectonic (VT) earthquakes (typically less than M 3). Changes in the flux, temperature and composition of hydrothermal fluids at the surface may also be observed (Newhall and Dzurisin, 1988; Acocella et al., 2015; Pritchard et al., 2019). Associated hazards include ground deformation, ground shaking, accumulation of gases and hydrothermal explosions, even where unrest does not evolve into an eruption (Potter et al., 2012).

Rapid changes in ground level and increases in the rates of VT seismicity can trigger an emergency response in a population, such as at Long Valley caldera, USA in 1980-1984 (Hill, 2006; 2017), Rabaul, Papua New Guinea in 1983-1985 (McKee et al., 2017) and Campi Flegrei in both 1970-1972 and 1983-1984 (Barberi et al., 1984). The resulting socio-economic impacts can last for years after unrest has ended. Economic losses may result from the financial costs of evacuation and mitigation measures, business interruption, increases in insurance premiums, loss of tourism and a decline in investment (Mader and Blair, 1987; Johnston et al., 2002; Benson, 2006; Potter et al., 2012; 2015). Social impacts are generally related to the disruption of lives and livelihoods (Potter et al., 2012). Previously, long-lasting emergencies, such as the examples above, have also been characterised by the development of mistrust between the public and those responsible for the operational response (i.e. scientists and emergency management), as a result of high levels of uncertainty as to the evolution of unrest, media speculation and rumours (Barberi et al., 1984; Potter et al., 2012; Hill et al., 2017).

Caldera unrest may be magmatic or non-magmatic in origin and is caused by the interaction of pressurised fluids (e.g. magma, exsolved magmatic volatiles or hydrothermal fluids) and the crust (Potter et al., 2012). No formalised definitions exist for these terms but throughout this thesis magmatic unrest will be used to refer to that which involves magma transport to shallower depths (e.g. Pritchard et al., 2019). Non-magmatic unrest describes pressure and/or volume changes in either hydrothermal or magmatic fluids. Generally, it is not possible to differentiate the nature of the source of an unrest episode, except where it evolves into an eruption (Newhall and Dzurisin, 1988; Acocella et al., 2015; Pritchard et al., 2019). Numerical modelling has shown that comparable rates and magnitudes of uplift may plausibly be produced by either a magmatic or hydrothermal source (e.g. Casertano et al., 1976; Hurwitz et al., 2007; Hutnak et al., 2009; Fournier and Chardot, 2012), whilst increases in fluid pressure can induce seismicity, irrespective of its nature. Pressure variations and alterations to flow paths caused by changes in the stress field can also result in changes in the rate and spatial extent of outgassing, as well as the geochemical characteristics of hydrothermal activity (Caliro et al., 2005; Lowenstern et al., 2006; Todesco et al., 2008; Chiodini et al., 2010; Cardellini et al., 2017). Where unrest is magmatic, the presence of a hydrothermal system can modify the signals of unrest monitored at the surface. For example, the opening and closing of fractures during magma intrusion can change flow paths, and thus heat and mass transport in an overlying hydrothermal system. This necessarily results in a redistribution of pore pressure, resulting in ground deformation that may be significant enough to change or amplify that due to the magma (Hurwitz et al., 2007; Todesco et al., 2008; Jasim et al., 2018). Geochemical signals of magma in fluids discharging at the surface (e.g. fumaroles and thermal waters) may also be masked as acidic magmatic gases such as SO_2 and HCl are removed by hydrolysis and scrubbing reactions in the hydrothermal system (Symonds et al., 2001; Rouwet et al., 2017). A further challenge for hazard evaluation at large calderas is that they can undergo decadal sequences of episodic unrest without eruption (~10-100 years), characterised by caldera-wide ground movements, and swarms of low-energy VT events, of variable intensity (Newhall and Dzurisin, 1988; Acocella et al., 2015).

Representative behaviours of long-term deformation at large calderas include that at Yellowstone caldera, USA, Long Valley, and Rabaul (Acocella et al., 2015). At Yellowstone, four periods of uplift have been observed between 1923-1985, 1995-1998, 2004-2010 and 2016-2017 that have cumulatively raised the ground level by c. 0.2 m (Fig. 1.3a). Characteristic of the uplifts are slow rates of displacement, with mean rates of c. 1.5-2 cm yr^{-1} (maximum of 7 cm yr^{-1} in 2004- 2010), and each is followed by subsidence of a similar magnitude (Wicks et al., 2006; Chang et al., 2007; Hurwitz and Lowenstern, 2014; Pritchard et al., 2019). Generally, changes in the direction of ground movements coincide with peaks in VT seismicity that are thought to represent the opening of fractures and transport of fluids (Smith et al., 2009). The cyclical inflation and deflation of the caldera is considered to represent the accumulation of fluids and their transport away from the pressure source respectively but it has not been possible to differentiate the nature of the fluids involved (Dzurisin et al., 1994; Waite and Smith, 2002; Hurwitz and Lowenstern, 2014). Proposed mechanisms for the uplifts include: the pressurisation of magmatic fluids as they are exsolved from a magma below an impermeable layer (Fournier, 2004; Shelley et al., 2013), the pressurisation of magmatic fluids below an impermeable layer following a magma intrusion (Dzurisin et al., 1994; Dzurisin et al., 2012), and a purely magmatic source (Wicks et al., 2006; Chang et al., 2007; Chang et al., 2010).

Long-term deformation at Long Valley since 1979 (Fig. 1.3b) contrasts from that at Yellowstone in that uplift is permanent (c. 0.85 m) and episodes of uplift are succeeded by periods of relative stability of the ground level, with no significant subsidence (Montgomery-Brown et al. 2015; Hildreth, 2017). Mean rates of displacement have varied from <1 to 10 cm yr^{-1}, except between 1979-1985 when uplift occurred at rates of up to c. 20 cm yr^{-1} triggering an emergency response of the population (Langbein, 2003; Hill, 2006; Montgomery-Brown et al. 2015; Hill, 2017). The permanent uplift requires the emplacement of a permanent strain source but, as for Yellowstone, a lack of definitive evidence for magma transport has led to the proposal of multiple mechanisms for uplift. They include: magma intrusion (Newman et al., 2001; Langbein, 2003, Battaglia and Hill, 2009; Montgomery-Brown et al., 2015), the pressurisation of a mixture of magma and hydrothermal fluids (Liu et al., 2011), bubbles rising within a magma volume (Linde et al., 1994), and the pressurisation of magmatic fluids below a low permeability horizon underlying the hydrothermal system during 'second-boiling' of a rhyolitic magma (Hildreth, 2004; 2017). Second-boiling refers to volatile release (CO_2 + H_2O) during late stage crystallisation (Hildreth, 2017). Hildreth (2017) argues against a magmatic source for the unrest and suggests that any magma in the subsurface has solidified on the basis that there is no evidence for intra-caldera eruptions in the last 500 000 years, the hydrothermal circulation is relatively low-temperature (100 °C between 2-3 km depth), the lack of clear evidence for a magma body in tomographic imaging, and the absence of thermal, seismic and geochemical signals of intrusion below the uplifted area. However, a gravity increase of 66 ± 151 µGal between 1982 and 1999 has been interpreted as an indicator of a mass increase with a density in the range of that for silicate magma (Battaglia and Hill, 2009).

The third case of long-term deformation at Rabaul between 1971-mid-September 1994 is characterised by continuous uplift that progressed at variable rates (Fig. 1.3c). It was accompanied by swarms of VT events (max ML 5.2) that were concentrated at depths of <3 km in the ring fault zone (McKee et al., 1984; McKee et al., 1985; Mori and McKee, 1987, Mori et al., 1989) and culminated in the VEI 4 eruption of the cones of Tavurvur and Rabaul (McKee et al., 2017). Between 1971 to mid-1983 deformation proceeded at c. 8 cm yr^{-1}, then in 1983-1985 there was a period of rapid deformation, similar to that at Long Valley between 1979-1985, with a mean rate of displacement of 5 cm per month, and peaks of up to 10 cm per month (McKee et al., 1984; Newhall and Dzurisin, 1988). During this period the central caldera was raised c. 0.8 m and rates of VT seismicity accelerated from a few hundred events per month up to 14 000 in April 1984 (McKee et al., 1984; McKee et al., 2017. The intensity of the unrest resulted in an emergency response that included the evacuation of the town of

Rabaul, with associated financial losses of $22.2 million (Benson, 2006). The rapid uplift was considered at the time to be related to the accumulation of magma and to potentially be precursory to an eruption (McKee et al., 1984). However, from 1985 through to late 1986 activity generally declined, returning to pre-1983 levels. Seismicity and uplift rates again accelerated after May 1992 until 1994 when the number of VT events decreased (McKee et al., 2017). On the 18th of September 1994 a 27-hour period of rapid uplift of c. 6 m and intense seismicity occurred that marked the final approach to eruption (McKee et al., 2017).

The full 1971-1994 deformation sequence at Rabaul is thought to represent the repeated intrusion of magma into the shallow crust (McKee et al., 1984; McKee et al., 1989; Saunders, 2001). Localised uplift of c. 1.5 m at the Rabaul Golf course between 1972-1973 is attributed to the pressurisation of the shallow hydrothermal fluids (Crick, 1975 - cited in Saunders, 2001) but the hydrothermal system is not considered to have contributed significantly to caldera-wide deformation, although a decline in the uplift rate in 1994 that coincided with the appearance of hybrid earthquakes may indicate the depressurisation of the hydrothermal system at Tavurvur, suggesting that it was impacted by the unrest (McKee et al., 2017). The critical features of the caldera-wide uplift are its continuous nature and the rapid intensification immediately prior to eruption. Similar behaviour is also known from historical records to have occurred in the years before previous eruptions in 1878 and 1937. As such it has been suggested that the full 1971-1994 deformation profile represents a single, long-term evolutionary sequence where the probability of an eruption increased as deformation progressed (Acocella et al., 2015; Robertson and Kilburn, 2016).

Figure 1.3: Representative patterns of uplift at Yellowstone (A.), Long Valley (B.) and Rabaul calderas. Graph A and B are from Pritchard et al. (2019), C. is from Acocella et al. (2015).

1.3 Long-term Caldera Unrest at Campi Flegrei

Two periods of historical unrest have occurred at Campi Flegrei; in the c. 100 years before the last eruption at Monte Nuovo in 1538, and from 1950 to Present (2019). Unrest prior to the Monte Nuovo eruption was characterised by episodic caldera-wide uplift that cumulatively raised the centre of the caldera by c. 17 m, felt VT seismicity, and increased degassing at the main fumarolic area of Solfatara-Pisciarelli (Di Vito et al., 1987; Dvorak and Gasparini, 1991; Guidoboni and Ciuccarelli, 2010; Giacomelli and Scandone, 2012). Earthquakes were occasionally strong enough to cause buildings to collapse in the town of Pozzuoli at the centre of the caldera and in 1470-1472 the gas flux from Solfatara-Pisciarelli was sufficient to cause vegetation die-offs in the surrounding area (Guidoboni and Ciuccarelli, 2010). Episodes of rapid uplift are also known to have occurred prior to 1503 and between 1503-1511 from historical records that detail changes in sea level and the emergence of land from the sea (Guidoboni and Ciuccarelli, 2010). Two years before the eruption there was an increase in the frequency of felt earthquakes, then in the 36 hours immediately before there was an intensification of unrest and a localised uplift of c. 6 m at the eventual vent site (Parascandola, 1946; Dvorak and Gasparini, 1991; Guidoboni and Ciuccarelli, 2010). The week-long eruption that followed began on 29th September and was a small (VEI 3) phreatomagmatic event that built the cone of Monte Nuovo (Parascandola, 1946; Guidoboni and Ciuccarelli, 2010; Global Volcanism Program, 2013). Overall, the decadal sequence of caldera-wide deformation and seismicity, and the rapid final approach to eruption may be similar to the long-term unrest that occurred at Rabaul between 1971-1994 (Kilburn et al., 2017).

Since 1950, four episodes of caldera-wide uplift have occurred without eruption in 1950-1952, 1969-1972, 1982-1984 and 2004-Present, cumulatively raising the centre of the caldera by a net c. 3.6 m (Del Gaudio et al., 2010). The deformation has been accompanied by VT seismicity and changes in fumarolic activity of varying intensity (Versino, 1972; Global Volcanism Program, 2013; INGV-OV, 2019). Major uplift and felt seismicity in 1969-1972 (1.77 m) and 1982-1984 (1.79 m) led to evacuations from the town of Pozzuoli. In 1969-1972 3000 people were permanently evacuated from the Rione Terra district, then in 1982-1984 40 000 people were evacuated from the town (Barberi et al., 1984). No eruption alert was issued in either case and the evacuations were justified on the basis of the seismic hazard (Versino, 1972; Barberi et al., 1984). Both emergencies were characterised by high uncertainty as to the evolution of the unrest, as well as public alarm and confusion that was fuelled by media speculation as to the likelihood of an eruption (Barberi et al., 1984; Peterson et al., 1993; Longo, 2018). In the case of the 1969-1972 uplift, a further 20 000 people are reported to have self-evacuated (Global Volcanism Program, 2013), prompted by fear of an eruption and the publication of conflicting scientific interpretations as to the evolution of the unrest (Tazieff, 1977; Imbo, 1979; Barberi et al., 1984). The socio-economic impacts of the emergencies were severe and persisted for years after the emergencies had ended. Economic losses included the costs of evacuation, the loss of workforce as people were displaced, the decline of tourism and property prices, and disruption of the fishing industry as boats could not use the harbour due to the uplift (Peterson et al., 1993; Kilburn et al., 2018). The development of the settlement of Monterusciello, which was built to resettle evacuees after 1982-1984, generated further costs of L. 2000 billion (Scandone and Giacomelli, 2018). Social impacts included the disruption of lives and livelihoods, loss of community due to evacuations, and the psychosocial effects of protracted periods of uncertainty and exposure to seismicity (Chandessais, 1982; Maj et al., 1989; Bland et al., 2005; Longo, 2018). Understanding the deformation behaviour of the caldera is key to improving hazard assessments and developing realistic scenarios of future unrest for mitigation planning.

The long-term deformation profile since 1950 (Fig. 1.4) shows that the pattern of ground movements is intermediate between that at the calderas discussed in section 1.2. The rapid uplifts in 1950-1952 (33.7 cm yr^{-1}), 1969-1972 (57.3 cm yr^{-1}) and 1982-1984 (69 cm yr^{-1}) progressed at rates comparable to those observed during the unrest crises at Long Valley caldera in 1979-1985 and Rabaul between 1983-1985 (section 1.2). Each episode also resulted in a permanent deformation. In contrast to the ground movements at these calderas, however, the 1950-1952 and 1969-1972 uplifts at Campi Flegrei were followed by minor subsidence, then after the 1982-1984 episode the caldera entered a phase of prolonged subsidence that lasted until 2004 of an amount c. 50% of the preceding uplift, suggesting a prolonged loss of pressure in the crust, as follows uplifts at Yellowstone.

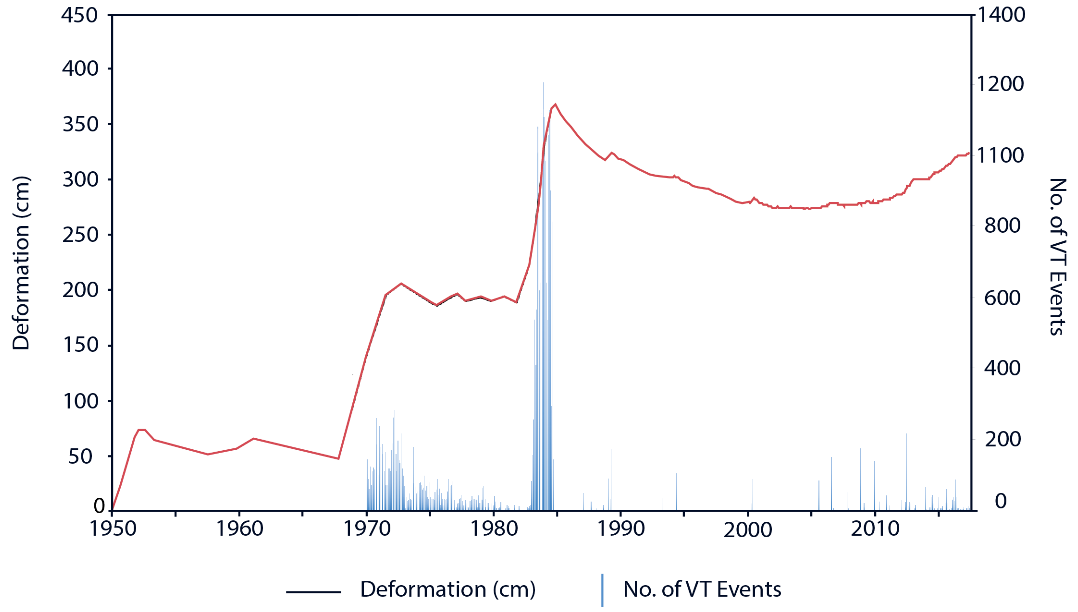


Figure 1.4: The long-term deformation trend at Campi Flegrei since 1950. Data from the Vesuvius Observatory.

The official scientific position during the 1969-1972 and 1982-1984 unrests, was that uplift was likely related to magmatic processes, but there were no definitive indicators of magma movement. As a result, an academic debate developed in the literature as to the nature of the source of the uplifts that has persisted to present. The 1969-1972 episode has been attributed to both a magma intrusion (Corrado et al., 1977; Bianchi et al., 1987; Woo, 2007) and an inflation of the hydrothermal system resulting from either a thermal pressurisation (Grindley, 1974), or tidal forcing (Casertano et al., 1976; Palumbo et al., 1985). Models constrained by the 1982-1984 unrest have variably attributed ground deformation to: magma intrusion (e.g. Berrino et al., 1984; Bianchi et al., 1987; Dvorak and Berrino, 1991; Amoruso et al., 2017), pressurisation of magmatic fluids (e.g. De Vivo and Lima et al., 2006; Bodnar et al., 2007; Lima et al., 2009), pressurisation of the hydrothermal system by magma or injection of magmatic fluids (e.g. De Natale et al., 1991; Battaglia et al., 2006; Troiano et al., 2011), or a combined magmatic-hydrothermal source (e.g. Orsi et al., 1999a; De Natale et al., 2006; Gottsman et al., 2006; Woo and Kilburn, 2010). Source density estimates from inversion of gravity data for the 1982-1984 unrest are ambiguous, ranging from values for supercritical hydrothermal fluids to silicate melts (142 kg m^{-3} - 2500 kg m^{-3}, Battaglia et al., 2006; Gottsman et al., 2006; Amoruso et al., 2008). This disparity is due to differing assumptions as to the crust's density structure and porosity.

Justifications for a hydrothermal source have included the non-eruptive nature of the unrest, changes in fumarole gas chemistry between 1983-1984, the interpretations for which are non-unique, and the necessity of generating large amounts of uplift at reasonable overpressures in the crust. The current prevailing view is that a component of the 1982-1984 uplift must have been related to the hydrothermal system, as a subsequent loss of pore pressure in this region of the crust is considered the simplest explanation for the slow subsidence that followed the 1982-1984 uplift (e.g. De Natale et al., 2006; Gottsman et al., 2006; Chiodini et al., 2003; 2015a, Moretti et al., 2017).

The characteristics of the most recent phase of uplift since 2004 are distinctly different to the preceding episodes. In particular it has progressed slowly, at a mean rate of 3.5 cm yr^{-1} and is approximately the inverse of the preceding subsidence. A peak in the deformation rate that was accompanied by VT swarms in December 2012 prompted the elevation of the Volcano Alert Level (VAL) from Green (Normal) to Yellow (Attention), where it has remained since (DPC, 2019). Interpretations of the cause of the deformation generally assume that it is a result of the pressurisation of the hydrothermal system but invoke processes that have conflicting implications as to whether the probability of an eruption is increasing as the deformation progresses (Chiodini et al., 2015a; 2016), or if it is at its lowest since 1982-1984 (Moretti et al., 2017). In the case of the former, the current uplift is attributed to injections of increasingly H_2O-rich magmatic fluids into the hydrothermal system from a crystallising shallow magma body, whereas the latter expects pressurisation to result from an increase in hot fluids originating at a deeper magmatic source located at c. 8 km depth. The high degree of uncertainty is a critical challenge for realistic hazard assessment and also the communication of unrest between stakeholders.

1.4 Research Aims and Objectives

Conceptual models of caldera unrest episodes form the contextual framework within which monitoring parameters are interpreted and communicated. They can also provide a basis for the development of future scenarios of unrest that may be used for unrest management planning and in the parameterisation of tools for hazard assessment (e.g. Bayesian Event Trees). The lack of direct observations and diagnostic indicators of operating processes in monitoring parameters means that all are inherently subjective. An essential step towards improving the definition of scenarios of the evolution of future unrest, is to develop robust models of the interactions between the magmatic and hydrothermal systems that are compatible with changes in monitoring parameters throughout the full long-term unrest sequence.

Conventional models of unrest at Campi Flegrei have variable starting assumptions and there is no agreement amongst them as to the structure of the crust in the deforming area, in particular the location of magma bodies and hydrothermal reservoirs. Typically, they are constrained by monitoring data from the 1982-1984 uplift onwards and none can account for the timing of the multi-decadal ground oscillation after 1984 (i.e. prolonged subsidence followed by slow uplift), or why similar behaviour was not observed following the uplifts in 1950-1952 and 1969-1972. Common to all is an assumption that uplift episodes are independent events, rather than part of a single long-term sequence. The main aims of this thesis are to:

i. Establish a model of the structure of the magmatic-hydrothermal system and the location of hydrothermal reservoirs based on the current knowledge of the Campi Flegrei subsurface and hydrothermal activity
ii. Determine the knowledge of the past behaviour of the caldera, a chronology of observations of the system and any indicators of changes in the conditions in the hydrothermal system
iii. Re-interpret long-term trends in monitoring data and observations of surface activity to produce a model that can account for the full sequence of ground movements since 1950
iv. Ascertain if a preferred mechanism for ground movements exists amongst scientists who may be involved in a response to a future intensification of unrest, their expected evolution of uplift since 2004 and perceived challenges in communicating unrest

The definition of these aims has been informed by the initial research question:

Is the long-term deformation sequence at Campi Flegrei since 1950 compatible with a single evolutionary sequence where the contribution of the hydrothermal system changes over time, and what is its perceived role in ground movements amongst those who may be involved in the scientific response to a future intensification of unrest?

To address this, the thesis will seek to answer the following:

a. Where in the crust are magma bodies, magmatic fluids and hydrothermal fluids known to be located in the crust?
b. Are the positions of hydrothermal reservoirs stable through time?
c. What is the relationship between the deforming area though time and the location of hydrothermal reservoirs?
d. Does the relationship between deformation and other monitoring parameters change significantly through time?
e. Did hydrothermal activity change significantly during and between uplift episodes?
f. What is the control on pore pressure in the hydrothermal system?
g. Are there any long-term changes in the pore pressure control and if so, where in the crust are these changes occurring?
h. Is there a preferred scenario for the controls on ground movements at Campi Flegrei amongst scientists?
i. Is there an agreement as to where magma is currently located in the crust and the expected evolution of ongoing uplift since 2004?
j. What, if any, are the perceived challenges of communicating unrest between stakeholders at Present (2019)?

1.5 Outline of Approach and Structure of Thesis

An interdisciplinary methodology will be applied to address the main research question. A literature review will enable the current understanding of the structure of the magmatic-hydrothermal system to be established and a model of the crust to be defined. To understand the past behaviour of the caldera observations of activity and existing interpretations from scientific publications, observatory reports and the media, will be reviewed in combination with long-term trends in monitoring parameters collated from these sources and existing data catalogues. This will then be used to understand temporal changes in monitoring parameters, define constraints that must be satisfied by conceptual models and to identify if there are any indicators that conditions in the hydrothermal system have changed through time. Integration of the findings of this review with the model of the structure of the magmatic-hydrothermal system will then allow for a reinterpretation of unrest and the definition of a novel conceptual model for post-1984 ground movements that is compatible with the full sequence of unrest since 1950. Social science methods will be adopted in order to ascertain perceptions of scientists as to the role of the hydrothermal system in unrest and challenges in communicating unrest. Data will be collected using questionnaires then interrogated using thematic and statistical analysis methods. The findings of this research may be used in the definition of future unrest scenarios and to better understand the differing mental models of the causes of unrest between scientists and therefore improve its communication.

The thesis is structured into eight chapters. This chapter has provided an overview of caldera unrest, the challenges of interpreting the nature of unrest and the characteristics of decadal ground movements at calderas, with particular relation to Campi Flegrei. The debate as to the cause of unrest at this volcano is introduced, as well as the limitations of conventional conceptual models in accounting for the long-term deformation profile. The aims and objectives of the research are also outlined.

Chapter 2 reviews the current state of knowledge of the Campi Flegrei magmatic-hydrothermal system. It establishes the tectonic and structural setting of the volcano, its eruptive history, and then synthesises the literature related to the structure of the magmatic-hydrothermal system to produce a schematic model of the crust. The objective of this chapter is to provide the geological and ideological context from which the rest of the thesis follows.

Chapter 3 reviews existing models of ground movements after 1982 and evaluates the extent to which they can account for the observed deformation. Chapter 4 then describes the methodology applied in re-interpreting monitoring data and developing a new model for ground movements. Chapter 5 chronologically reviews long-term changes in deformation, it's relation with other monitoring parameters though time and observations of hydrothermal activity at the surface through time. It then summarises the key observations that must be satisfied by models of the causative processes of unrest.

Chapter 6 presents the primary result of the thesis, which is a new model for ground movements at Campi Flegrei since 1950 that can account for the change in the characteristics of ground movements after 1984 for the first time. The chapter uses the characteristics of seismicity and deformation through time to show that ground movements since 1984 can be explained in terms of a continued adjustment of the hydrothermal system to a bulk permeability change in the hydrothermal system in 1984 that resulted from a progressive increase in fracturing with uplift over successive uplifts between 1950-1984.

Chapter 7 is a self-contained chapter that describes the methodology used to collect and analyse the data used to investigate the perceptions of scientists as to the role of the hydrothermal system in unrest at Campi Flegrei before presenting the results. Chapter 8 concludes the thesis by summarising the main findings of the research and identifying objectives for future research.

Chapter 2

The Campi Flegrei Magmatic-Hydrothermal System

Caldera unrest may be magmatic or non-magmatic in origin and is caused by the interaction of pressurised fluids (e.g. magma, exsolved magmatic volatiles or hydrothermal fluids) and the crust (Potter et al., 2012). The development of realistic conceptual models of unrest that maybe used in the development of unrest scenarios requires a knowledge of: (i) sources of fluid within the magmatic-hydrothermal system; (ii) fluid storage zones; and (iii) controls on fluid flow. In this chapter a generalised overview of the tectonic setting of the volcano and eruptive history is presented before a synthesis of multi-disciplinary information related to the structure of the magmatic-hydrothermal system. The objective of this chapter is to provide the geological context from which the rest of the thesis follows.

2.1 Tectonic Setting

Campi Flegrei is a volcanic field located in the Campanian Plain, a half-graben situated between the Neogene thrust belt of the Southern Apennines and the Tyrrhenian Sea (Fig. 2.1). It is part of the Campi Flegrei Volcanic Zone (CVZ), which includes the centres of Campi Flegrei, Ischia, Procida and a number of submarine vents. The volcanism is the result of back-arc extensional tectonism related to subduction within the Africa-European convergence zone (Zuppetta and Sava, 1991). Beginning in the late Miocene-early Pliocene (c. 8-10 Ma) the Tyrrhenian Sea back-arc began to open as a result of slab roll-back towards the SE of the Ionian plate below the Calabrian Arc (Rosenbaum and Lister, 2004; Piochi et al., 2005; Mattei et al. 2010; Milia and Torrente, 2011; Vitale and Ciarcia, 2018). Subsequent trans-tensional ESE-WNW extension during Pliocene-Quaternary rifting then displaced Mesozoic sediments along high-angle NW-SE and NE-SW normal faults. This led to the formation of horst-graben structures along the Tyrrhenian margin of Italy, including the Campanian Plain (Milia et al., 2003; Piochi et al., 2005). This regional tectonism has been attributed to an anticlockwise rotation of the Italian Peninsula (Scandone et al., 1991; Florio et al., 1999) and to gravitational spreading of the Apennines (Woo and Kilburn 2010 and references therein).

Typical of an extensional setting, the Campanian Plain is associated with lithospheric thinning, high heat flow (up to 200 MW m^{-2}), and is bound by high angle normal faults that are located at the base of the Mesozoic carbonate platforms of Mt. Massico and the Sorrento Peninsula (Fig. 2.1b, Scandone et al., 1991; Piochi et al., 2005; Woo and Kilburn 2010; Milia and Torrente, 2011; Piochi et al., 2014 and references therein). The structure is c. 70 km long and 30 km wide. It is defined by a large negative gravimetric anomaly that is caused by a thick (2-3 km) infill sequence of Pliocene-Quaternary clastic sediments and potassic volcanic units (Rosi and Sbrana, 1987; Barberi et al., 1991; Scandone et al., 1991; Judenherc and Zollo, 2004; Piochi et al., 2014). Volcanic centres are distributed throughout the half-graben, the locations of which are tectonically controlled by NE-SW and subordinate NW-SE normal fault systems that have acted as pathways for magma (Orsi et al., 1996; Acocella et al., 1999; Acocella and Funicello, 2006; Milia and Torrente, 2011). Radiometric dating suggests that eruptive activity began c. 0.8-0.5 Ma in the northern sector of the Campanian Plain at the extinct Ventotene and Roccamonafina volcanoes, then migrated SE to the Neapolitan region by c. 0.36 Ma (Peccerillo, 2005; Piochi et al., 2005 and references therein; Crosweller et al., 2012).

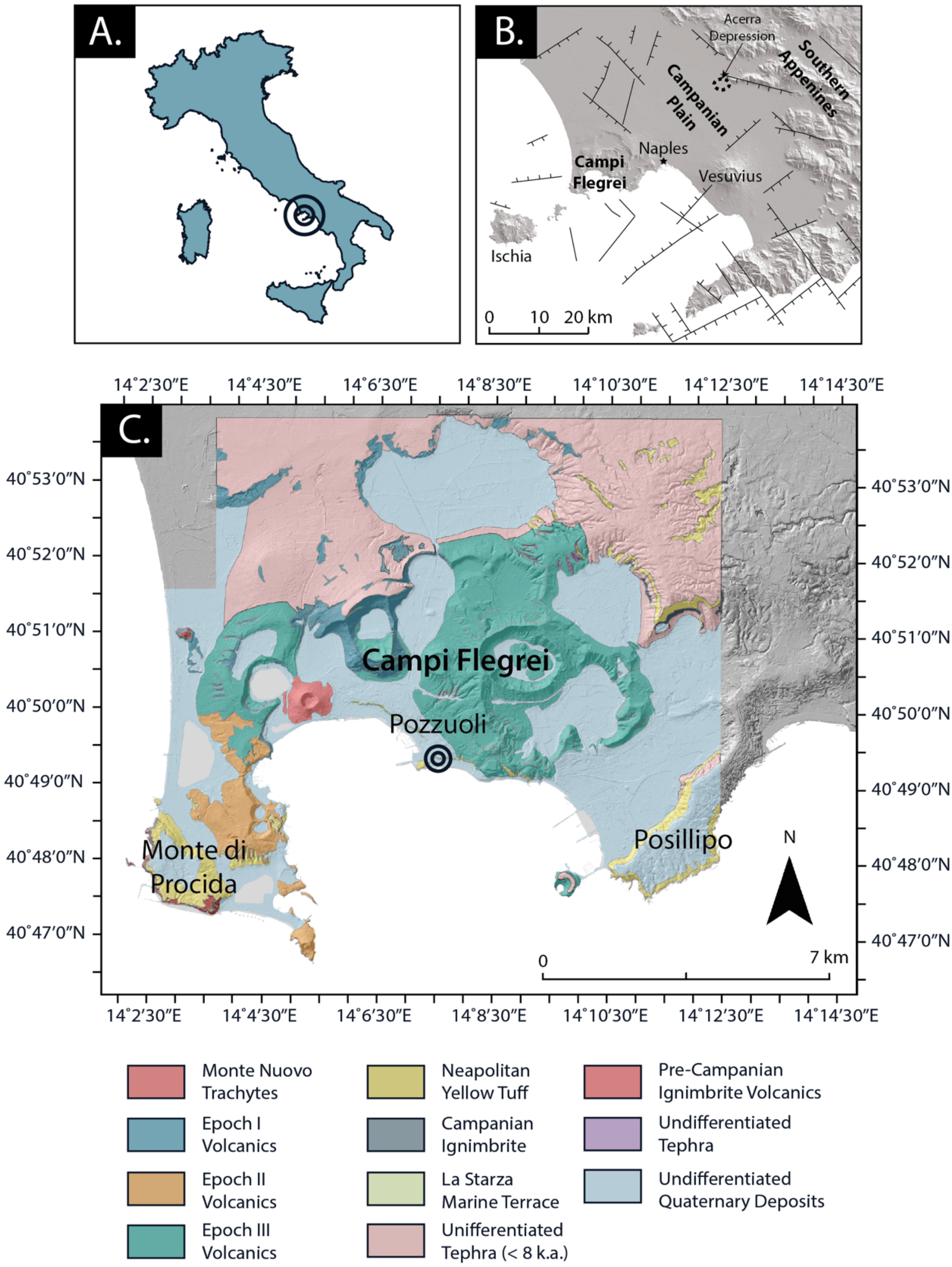


Figure 2.1: Campi Flegrei location map. Boxes A and B show the location of the caldera in Italy and the tectonic setting of the Campanian Plain respectively. Box C is a geological map showing the primary geological units. The town of Pozzuoli is located at the centre of the caldera. Data and base map from the Vesuvius Observatory.

2.2 Eruptive History

2.2.1 Caldera Formation

The Campi Flegrei Volcanic District (CFVD) has erupted at least six ignimbrite eruptions at c. 289-205, 184, 157, 39 and 14.9 ka, and sanidine xenocryst ages suggest an active centre has existed since at least 315 ka (De Vivo et al., 2001; Gebauer et al., 2014; Belkin et al., 2016). Persistent activity is known from the stratigraphic record to have occurred since at least c. 80 ka (Rosi et al., 1983; Pappalardo et al., 1999; Tomlinson 2012) and discrete centres within the modern area of Campi Flegrei have been dated to c. 60 ka (Rosi and Sbrana., 1987). The volcanism has dominantly been explosive and largely occurred from monogenetic vents (Orsi et al., 1996; Pappalardo et al., 1999; Vitale and Isaia and references therein). The deposits are dominated by alkali-trachytic tuffs and tuffites interbedded with silty, arenaceous and marly sediments, with subordinate lava flows and domes (Fig, 2.1c., Rosi and Sbrana, 1987).

The present-day area of Campi Flegrei is dominated by a quasi-circular depression c. 8-12 km in diameter that encloses the Gulf of Pozzuoli and is bound by a continuous topographic high between Monte di Procida and Posillipo Hill onshore. It is widely considered to represent the margins of a caldera that was formed by one or more volcano-tectonic collapses (Rittman, 1950; Rosi and Sbrana 1987; Lirer et al., 1987; Barberi et al., 1991; Scandone et al., 1991; Orsi et al., 1996; Judenherc and Zollo, 2004; Battaglia et al., 2008; Sacchi et al., 2014), although alternative tectonic mechanisms have been proposed for the depression (e.g. Bellucci et al., 2006a; Milia et al., 2006). Prior to caldera formation, eruptions occurred across an area that extended into Naples, sometimes referred to as the Palaeoflegrei but post-collapse volcanism has been confined to the caldera margins (Barberi et al., 1991; Cole et al., 1994; Orsi et al., 1996; Perrotta et al., 2006; Scarpati et al., 2012).

The timing and extent of caldera formation is controversial. It is debated as to whether it was formed by a single collapse during the ignimbrite eruption of the Neapolitan Yellow Tuff (NYT, 14.9 ± 0.4 ka, Deino et al., 2004), or whether subsidence associated with the NYT occurred along faults related to an earlier collapse during the eruption of the Campanian Ignimbrite (CI, 39.85 ± 0.14 ka, Giaccio et al., 2017) that were reactivated. The CI eruption is the largest to have occurred in Europe in the last 200 ka and is variably estimated to have erupted between 54-300 km^3 (Dense Rock

Equivalent, DRE) of trachytic–phonolitic trachyte magma (Fig. 2.2, Scarpati et al., 2014; Smith et al., 2016). The preferred source for the eruption in the literature is the Campi Flegrei caldera (Barberi et al., 1978; Rosi et al., 1983; Rosi and Sbrana, 1987; Barberi et al., 1991; Fisher et al., 1993; Orsi et al., 1996; Fedele et al., 2008; Perrotta et al., 2006; Acocella, 2008; Scarpati and Perrotta, 2012) and a vent location has been inferred near the Quarto Plain from fall deposits (Scarpati et al., 2015). However, gravimetric and magnetic surveys of the region do not support the presence of a caldera with the proposed dimensions (Fig. 2.3a and b, De Vivo et al., 2001; Rolandi et al. 2003). Recent analysis of core from a pilot hole drilled at Bagnoli as part of the Campi Flegrei Deep Drilling Project (CFDDP) further confirms that caldera collapse does not propagate east of the Posillipo Hill (De Natale et al., 2016). Alternative sources proposed for the CI include eruption from a vent in the Acerra depression (a tectonic feature c. 15 km NE of Naples, Scandone et al., 1991) or from fissure eruptions along existing faults (Di Girolamo et al., 1984; Lirer et al., 1987; De Vivo et al., 2001; Rolandi et al., 2003). A source from fissures controlled by neo-tectonic Apennine faults, distributed throughout the Campanian Plain is supported by the distribution of proximal lithic deposits of the CI, the lack of thick CI deposits on the proposed caldera rim and the distribution of deposits along a belt in the western Apennines between Naples and Monte Massico (De Vivo et al., 2001; Rolandi et al., 2003).

The NYT was produced in a phreatoplinian eruption that conservative estimates suggest erupted 30-50 km^3 (DRE) of trachytic magma and has a confirmed source location at the Campi Flegrei caldera (Barberi et al., 1991; Orsi et al., 1992; Scarpati et al., 1993). Deposits are found over an area of more than 1000 km^2, with thick sequences located on the caldera margins (Wohletz et al., 1995; Scandone et al., 1991). The volcano-tectonic collapse attributed to the NYT correlates strongly with a well-defined Bouguer gravity low (8-12 μgal) related to the low-density caldera fill (Fig. 2.3b), as well as subsidence identified in seismic tomography and reflection studies (Scandone et al., 1991; Barberi et al., 1991; Florio et al., 1999; Zollo et al., 2008; Della Iacono et al., 2009; Sacchi et al., 2014). Surrounding the gravity low is a ring of positive gravity anomalies and high P-wave velocities (V_p), inferred to have been caused by hydrothermalised lava and magma intrusions, which broadly correspond to the location of vents at the surface and define a zone of ring faults c.1-2 km in width (Zollo et al., 2003; Chiabbra and Moretti, 2006; Battaglia et al., 2008; Dello Iacono et al., 2009; Capuano et al., 2013; Sacchi et al., 2014). The collapse was piecemeal and strongly controlled by regional tectonics producing the quasi-circular depression, which is elongated WNW–ESE and has a diameter of c. 7 km (Scandone et al., 1991; Bruno et al., 2004; Capuano et al., 2013; Steinmann et al., 2018). Despite the clear geophysical evidence, there are variations in the definitions of the caldera margins resulting from different interpretations of the locations of ring faults, which vary due to modification of the caldera rim by subsequent eruptions. Throughout this work, reference to the Campi Flegrei caldera is to the NYT collapse and the ring fault zone (RFZ) is taken as the approximate area of positive gravity anomalies surrounding the main collapse highlighted in Fig. 2.3b.

Following caldera formation until c. 2 ka, uplift attributed to caldera resurgence occurred in the centre of the caldera, leading to the formation of a dome like structure that is centred offshore and bound by the inner ring faults (c. 180 m), as well as the uplift of the La Starza Marine Terrace (c. 60-80 m, Di Vito et al., 1999; Bellucci et al., 2006b; Isaia et al., 2009; Sacchi et al., 2014; Steinmann et al., 2018; Marturano et al., 2018). La Starza is a NW-SE trending volcano-tectonic structure that is delimited by a fossiliferous marine cliff overlain by subaerial pyroclastics. It shows clear evidence of brittle faulting and has been interpreted as being the boundary of a resurgent block that has been uplifted by a simple shear-mechanism (Rosi and Sbrana, 1987; Orsi et al., 1996; Isaia et al., 2009; Marturano et al., 2018). After uplift ended the deformation regime reversed and the caldera began subsiding at an estimated mean rate of 12-17 mm yr^{-1}, significantly greater than may be attributed to the regional extensional tectonic regime (c. 2 mm yr^{-1}, Dvorak and Mastrolorenzo, 1991; Bellucci et al., 2006b; Sacchi et al., 2014; Steinmann et al., 2018; Marturano et al., 2018). The mechanism controlling this ground movement is unknown but has been attributed to combinations of compaction, pore-pressure changes or volume reduction of magma bodies (Rosi and Sbrana, 1987; Dvorak and Mastrolorenzo, 1991; Bellucci et al., 2006b; Todesco et al. 2014).

Hydrothermal systems have been active since before caldera formation, as recognized from the presence of the hydrothermally altered clasts in the Breccia Museo, a CI deposit (Rosi et al., 1983; Barberi et al., 1991; Rosi et al., 1996). Today the caldera hosts a high temperature hydrothermal system (mean fluid temperature >250 °C) the age of which is uncertain. It is thought that it likely to have developed soon after the NYT eruption (Rosi and Sbrana, 1987), which may have erupted through an existing area of hydrothermal activity. Altered lithic clasts in deposits from an explosive eruption at Solfatara, close to the centre of the caldera, confirm the presence of an active system at this location by at least c. 4.2 ka (Isaia et al., 2009).

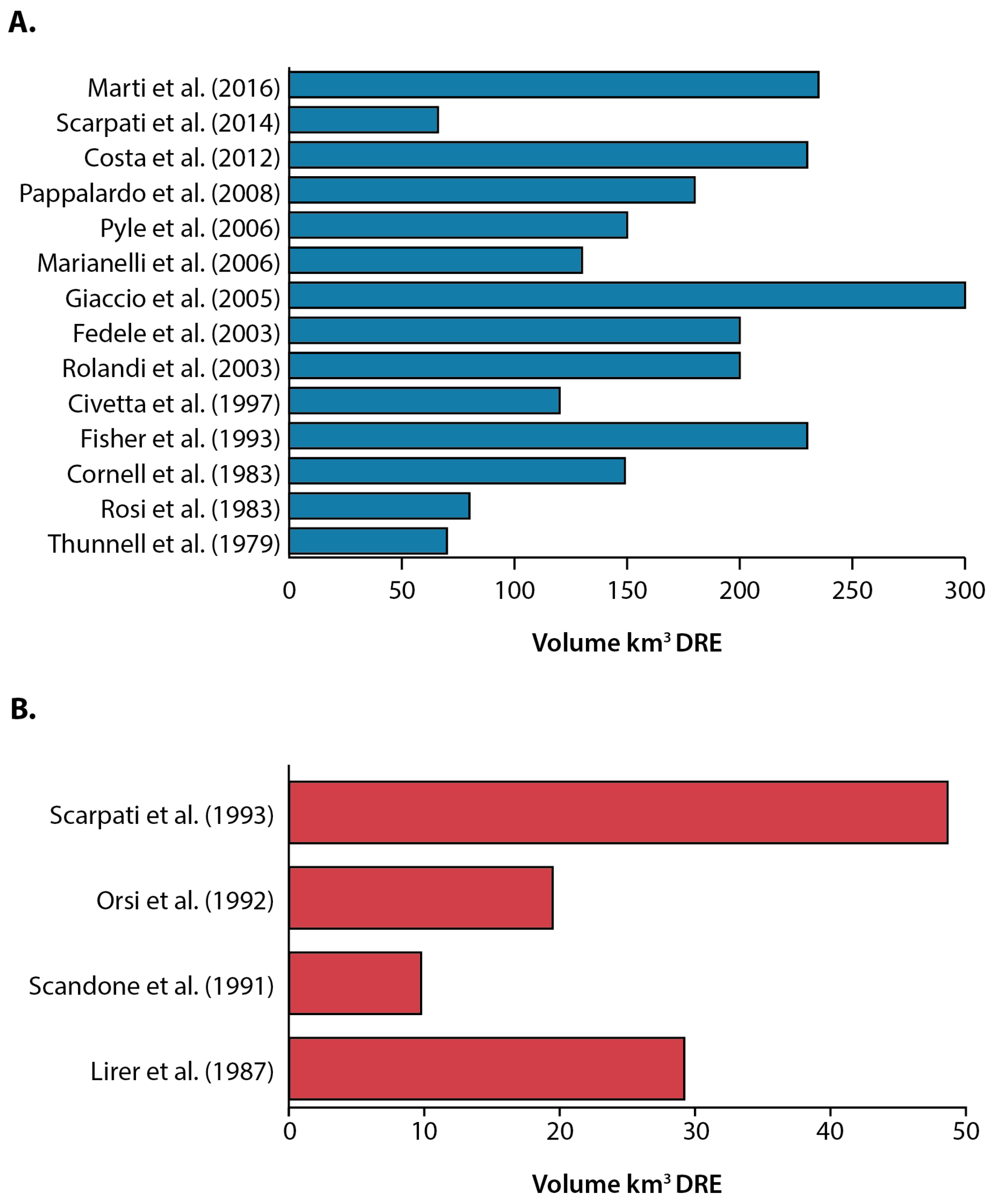


Figure 2.2: Estimates of ignimbrite volumes erupted from the Campi Flegrei Volcanic District as the Dense Rock Equivalent (DRE). A is for the Campanian Ignimbrite (modified after Scarpati et al., 2014) and B is for the Neapolitan Yellow Tuff. References for ignimbrite volumes are given in the figure.

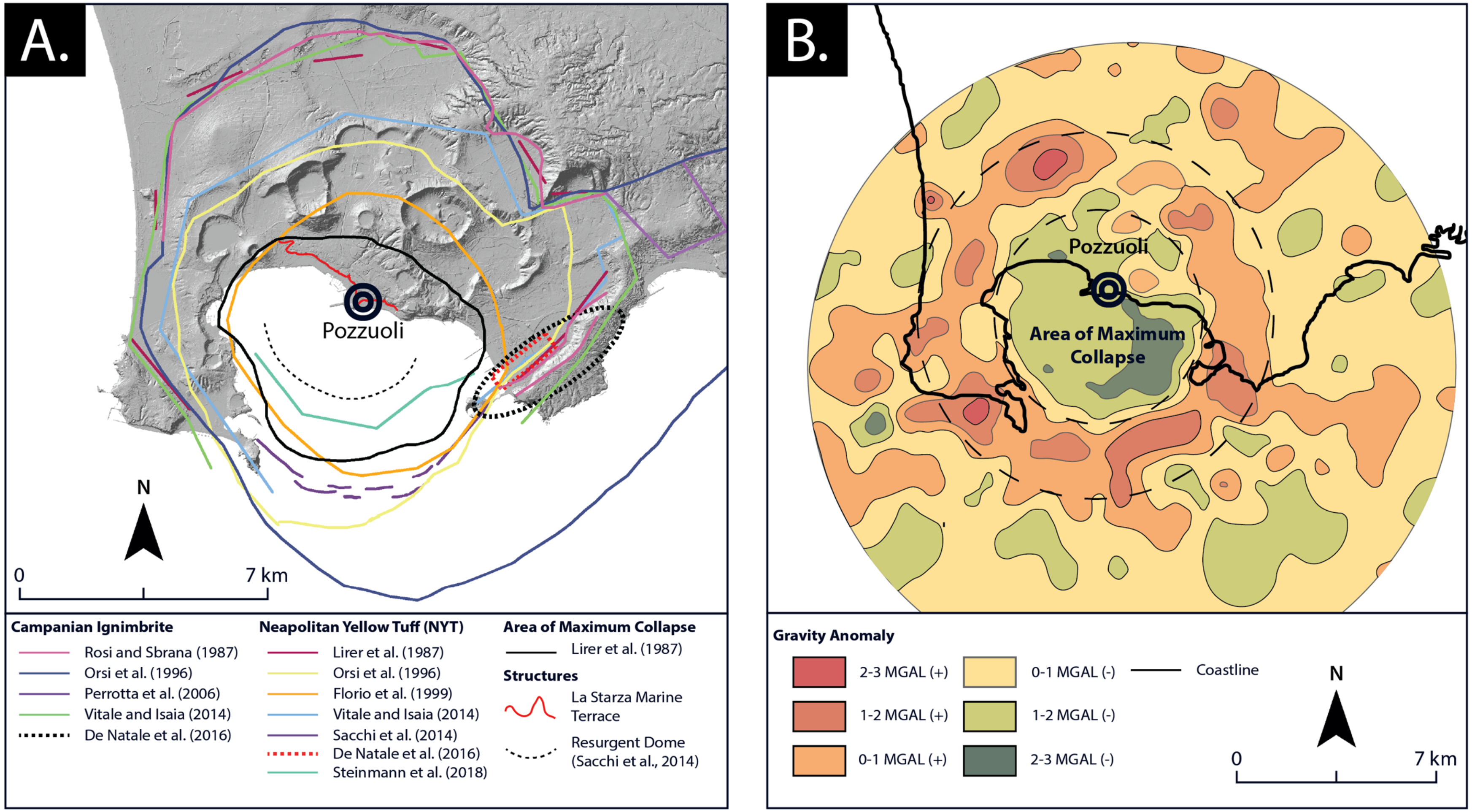


Figure 2.3: A. Interpretations of the structural boundaries of the proposed Campanian Ignimbrite (CI) and Neapolitan (NYT) calderas, as defined by the references in the figure legend. B. Bouguer gravity map for Campi Flegrei (modified from Barberi et al., 1991). A clear negative anomaly is associated with the area of maximum collapse, which is surrounded by positive gravity highs, highlighted by the dashed circles, corresponding to known vent locations and magmatic intrusions. This area corresponds with the Neapolitan Yellow Tuff caldera ring fault zone.

2.2.2 Post-NYT Eruptive Activity

Throughout the post-NYT period, eruptive activity has been confined to the onshore sector of the RFZ (Fig. 2.4, Di Vito et al., 1999; Orsi et al., 2004; Charlton, 2018). No vents along the offshore portion have been located, but shallow (<200 m depth) magma intrusions younger than 3.7 ka have been imaged within the ring faults using seismic reflection surveys (Steinmann et al., 2018). The spatial correlation between the ring faults and the magmatic activity strongly implies that these structures act as preferential pathways for magma (Charlton, 2018). At least 67 intra-caldera eruptions, from c. 40 centres have been recognised from the stratigraphic record following the NYT collapse, each of which erupted 0.1-1 km^3 (DRE) of potassic magma (k-trachyte to k-trachyphonolite, Di Vito et al., 1999; Smith et al., 2011). More mafic compositions such as shoshonite and latite are rare (Rosi and Sbrana, 1987; D'Antonio et al., 1999a). The activity was predominantly explosive and phreatomagmatic (80% of known eruptions) with occasional magmatic phases that were strombolian to plinian in style (Di Vito et al., 1999; Smith et al. 2011). Most eruptions took place at monogenetic vents, occasionally forming scoria cones (e.g. Monte Nuovo) and tuff rings (e.g. Averno). Minor dome forming activity is restricted to the central SE sector of the caldera in the Solfatara area (Orsi et al., 2004; Smith et al., 2011).

Eruptions appear to have been clustered into three distinct epochs in time (Fig. 2.4); Epoch I (15-10.6 ka), Epoch II (9.6-9.1 ka) and Epoch III (5.3-3.5 ka, Di Vito et al., 1999; Smith et al., 2011). Each is separated by a prolonged period of quiescence recognised by the appearance of palaeosols in the stratigraphic record (Di Vito et al., 1999; Orsi et al., 2004). During Epoch I between 30-37 explosive eruptions from c. 21 centres occurred from vents that were distributed throughout the RFZ (Di Vito et al., 1999; Smith et al., 2011). They were generally phreatomagmatic in character, forming tuff rings and cones, the largest of which is Monte Gauro at 331 m (Orsi et al., 1996, 2004; Di Vito et al., 1999). Epoch II followed after an apparent quiescence of c. 1000 years. Eruptive activity occurred from 7 centres and 8 low-magnitude, phreatomagmatic eruptions have been recognised (Orsi et al., 2004; Smith et al., 2011). The events are clustered in two locations; in the western sector between Baia and Capo Miseno, and in the NE. It was during this period that uplift of the La Starza marine terrace began (Di Vito et al., 1999).

At least 27 eruptions from 17 centres occurred during Epoch III (Smith et al., 2011). The majority had volumes of 0.01-0.1 km^3 and were phreatomagmatic to strombolian in character. The vents are distributed around the caldera margins but a cluster of eruptions occurred at the polygenetic centres of Astroni and Agnano (Di Vito et al., 1999; Isaia et al., 2004; Isaia et al., 2009). This period also produced the plinian eruption of Agnano Monte Spina (AMS) at 4.4 ka (0.85 km^3 DRE, Smith et al., 2011). It formed a minor caldera collapse of c. 35 m along NE-SW and NW-SE trending faults, creating a depression with a diameter of c. 3.5 km that corresponds with the Agnano Plain (De Vita et al. 1999; Arienzo et al., 2010). A repose of 150-200 years followed, during which there was a subsidence and uplift of the Campi Flegrei caldera of up to 30 m that has been attributed to the shallow (< 3 km) inflation of magma bodies (Isaia et al., 2009). Epoch III is subdivided into the periods IIIa (pre-AMS) and IIIb (post-AMS, Isaia et al., 2009). Activity during the latter produced the only known effusive eruptions, including those that produced the lava domes of Accademia, Monte Olibano and Solfatara, which are located on the ring faults of the AMS collapse on the SW flank. Solfatara is now the primary site of hydrothermal surface activity. The most recent – and only historic - eruption occurred in 1538 A.D. It produced the cone of Monte Nuovo (0.03 km^3 DRE), which buried the village of Tripergole in the first 48 hours of a week-long eruption (Di Vito et al., 1987). It was initially phreatomagmatic, originating offshore, then migrated onshore along a fissure where activity became strombolian in style. This eruption is considered to be the type event for Campi Flegrei and the most likely style of activity in the event of a future renewal of volcanism (Campi Flegrei Working Group, 2012; Di Vito et al., 2016 and references therein).

Phreatic eruptions are known to have occurred however, their products are difficult to recognise in the stratigraphic record, which is dominated by magmatic eruption products. The best-known phreatic deposits are from Solfatara and date from the period of intense, localised volcanism (4.1-4.4 ka) at the SW margin of the AMS caldera (Smith et al., 2011; Isaia et al., 2015). The Solfatara tephra sequence indicates that activity was initially phreatic, then evolved into magmatic vulcanian activity (Isaia et al., 2015; Pistolesi et al., 2016). No confirmed phreatic deposits have been located in other parts of the caldera. A historical phreatic eruption at Solfatara in 1198 A.D. has been identified from 16th and 17th century texts (Rosi and Santocroce, 1984; Scandone et al., 2010) but no associated deposits have been located and it has since been attributed to a transcription error (Guidoboni and Ciuccarelli, 2010). The most energetic confirmed historical explosive activity at Solfatara is a mud fountaining event that was dynamically triggered by the 23rd July 1930 M 6.4 Irpinia earthquake (Signore, 1930; Newhall and Dzurisin, 1988). A community living on the rim of the crater reported to the Vesuvius Observatory that, during ground shaking, mud pools in the centre of the crater stopped bubbling and that once the movement ceased, small explosions expelled mud 10-15 m in the air. These were followed on 3rd August 1930 by further explosions that caused mud fountains that reached heights of 25-30 m above the crater floor (Signore, 1930). These events are consistent with the sudden release of pressurised fluids due to fracturing during the ground motion and indicate that the stress regime within the upper crust was sensitive to external perturbations. This type of activity has not been observed since despite the occurrence of an earthquake with a similar magnitude and source location in 1980 (Mw 6.9, Irpinia, U.S. Geological Survey, Earthquake Hazards Program, 2019).

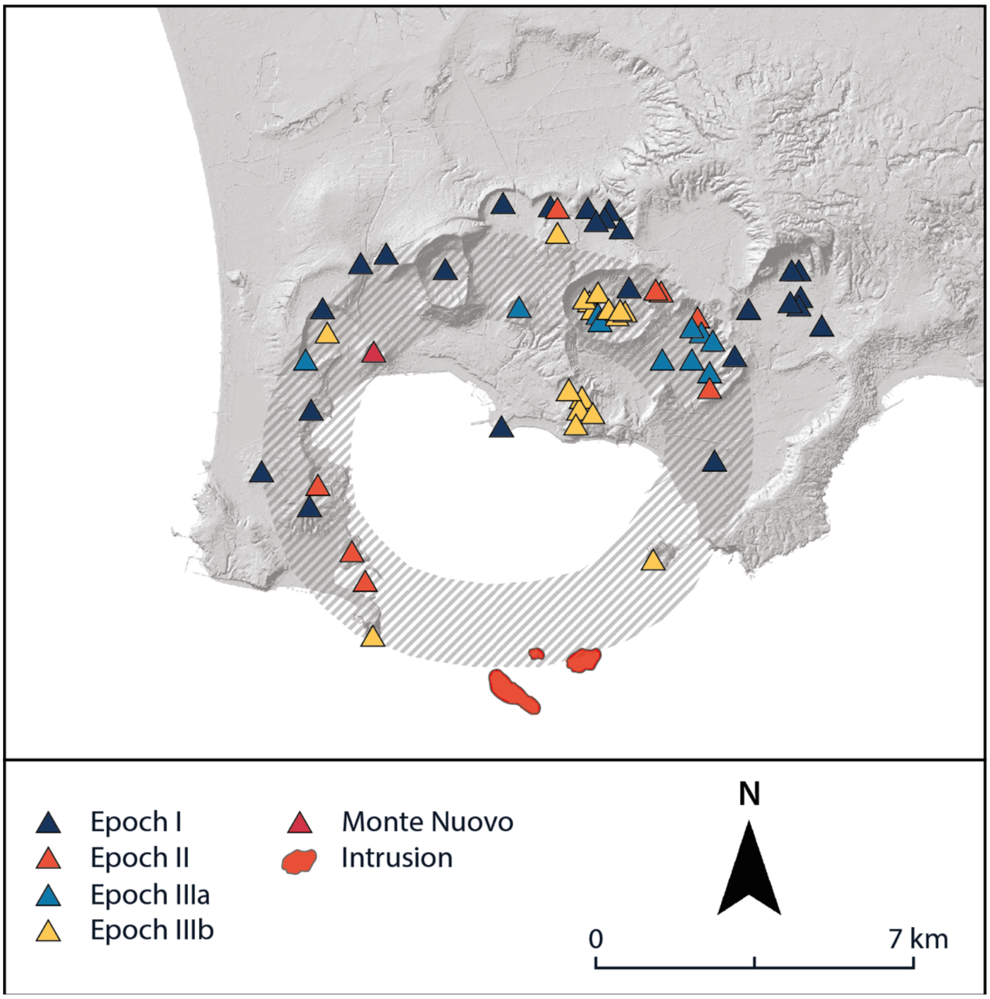


Figure 2.4: Vent locations identified from eruption deposits and the approximate Ring Fault Zone (grey hashed area). Three intrusions identified by Steinmann et al. (2018) located offshore in the southern sector have been included to highlight that whilst no vents have been located in this area, the ring faults have still been exploited as a preferential pathway for magma. Vent locations from Smith et al. (2011). Intrusion locations from Steinmann et al. (2018).

2.3 The Magmatic-Hydrothermal System

Following classic magmatic-hydrothermal system models derived from observations of porphyry-epithermal mineralisation and the results of geothermal exploration drilling (e.g. Burnham, 1979; Fournier, 1999), the Campi Flegrei system can be divided into two pore fluid pressure regimes; (i) the magmatic regime, which consists of a mid-crustal magma reservoir and an overlying storage zone of magmatic gas and brine where pore pressures can approach lithostatic; and (ii) the hydrothermal regime, where dominantly meteoric fluids flow at near hydrostatic pressures.

2.3.1 The Magmatic Regime

The top of the magmatic regime is expected at c. 3 km depth. This is the location in the crust at which high V_p/V_s anomalies interpreted as fractured reservoirs of hydrothermal fluids moving downwards from the surface terminate (Vanorio et al., 2005; Battaglia et al., 2008; De Siena et al., 2010). Meteoric circulation is limited to this depth by the presence of a low-permeability hard rock layer identified from a seismic discontinuity (c. 2.7 to 3 km depth) where there is a significant (>35%) increase in P (V_p) and S-wave (V_s) velocities compared to the overlying crust (Fig. 2.5, Battaglia et al., 2008; Zollo et al., 2008). The top of this layer was penetrated by geothermal exploration boreholes at San Vito and Mofete in the centre and west of the caldera respectively. Maximum temperatures of c. 420 °C were recorded at 3 km depth at San Vito, and at in both areas there was a decrease in permeability at these depths to $\leq 10^{-17}$ m^{-2} from mean overlying values of 10^{-15} m^{-2} (Rosi and Sbrana, 1987; Carlino et al., 2012; 2016) and conduction-dominated heat transfer conditions due to extensive recrystallisation, amorphous silica deposition and mineral precipitation (Bruni et al., 1981; Chelini and Sbrana 1987). Permeability is expected to decrease further with depth due to the onset of ductile deformation at the Brittle-Ductile Transition (BDT), which is located at c. 3.5-4 km depth (Castaldo et al., 2019). However, it must remain non-negligible to permit the persistent discharge of magmatic gases (e.g. CO_2, ^3He) from fumaroles at the surface (Tedesco et al., 1989; Tedesco and Scarsi et al., 1999; Caliro et al., 2007).

The hard rock layer is underlain by a laterally extensive seismic anomaly with a low V_p/V_s ratio of 1.3-1.4 at 3-4 km depth (Ferruci et al., 1992; Judenherc and Zollo, 2004; Vanorio et al., 2005; Zollo et al., 2008; Battaglia et al., 2008). The anomaly overlies a basement of ambiguous lithology that is generally assumed to be carbonate but there is a lack of such lithics in eruptive deposits. Its density is also consistent with crystalline igneous rock (2600-2650 kg m^{-3}, D'Antonio et al. 2011b). The low V_p/V_s anomaly at 3-4 km is widely inferred to be a highly fractured reservoir of overpressured gas and brines, which have been attributed to magmatic degassing and to decarbonation reactions of a carbonate basement (Vanorio et al., 2005; Chiarabba and Moretti, 2006; Battaglia et al., 2008; Zollo et al., 2008; De Siena et al., 2010). The $\delta^{13}C$ signature ($\delta^{13}C$ -1.4 ± 0.4‰) of fumarole gases is too negative for decarbonation (0 ± 2‰, Allard et al., 1991; Panichi and Volpi, 1999), and is consistent with degassing from a CO_2-rich magma derived from the local mantle source (Allard et al., 1991; Martelli et al., 2004; Caliro et al., 2007). Fluid pressures above hydrostatic at these depths are supported by the occurrence of vein networks in deep boreholes drilled at Mofete and San Vito that indicate flow at near lithostatic pressures (De Vivo et al., 1989). Super-hydrostatic pressures have also been measured at the base of hydrothermal circulation at other hydrothermal systems (Fournier, 1999 and references therein; Zencher, 2006). Such magmatic-gas brine layers are typical of large caldera magmatic-hydrothermal systems and similar bodies have been imaged using geoelectrical methods elsewhere (e.g. Yellowstone and Long Valley Caldera, USA, Uturuncu, Bolivia, Hurwitz and Lowernstern, 2014; Blundy et al., 2015; Hildreth, 2017; Afanasayev et al., 2018 and references therein). The reservoir at Campi Flegrei is considered to supply magmatic fluids to the overlying hydrothermal system (Caliro et al., 2007; Caliro et al., 2014). As a result, the flux of energy and mass into the hydrothermal system must be controlled by the permeability of the transition between the magmatic and hydrothermal regimes.

Traditional models of the magmatic system assumed the presence of a single large (~10^2 km^3) magma body located at 4-5 km that has been contracting in volume since the eruption of the CI (e.g. Armienti et al., 1983; Barberi et al., 1991). This was based on extrapolation of the geotherm from exploration wells at San Vito (Armienti et al., 1983; Rosi and Sbrana, 1987), bottom depths of earthquake hypocentres (De Natale and Zollo, 1986), the conversion of seismic P-waves to SV-waves at 4 km depth (Ferrucci et al. 1992), and the interpretation of temperature data from seismic Q_p models (De Lorenzo et al., 2001). However, no magma bodies with volumes greater than 1 km^3 have been imaged at depths shallower than 6 km (Zollo et al., 2008). The preferred model for the magmatic system is now one of multi-level magma storage where small volume magma batches (<1 km^3) fed by a primary magma reservoir are periodically emplaced at shallower depths. Such a conceptual model is consistent with the chemical and isotopic characteristics of eruption products, which are indicative of the polybaric evolution of magmas (Fig. 2.6, Pappalardo et al., 2008; Mangiacapra et al., 2008; Di Renzo et al., 2011; Piochi et al., 2014), as well as the evolution of general conceptual models of caldera magma systems (e.g. Cashman and Giordano, 2014; Pritchard et al., 2019). The primary magma reservoir, which is the principal source of heat and magmatic fluids throughout the magmatic-hydrothermal system, is inferred from a seismic velocity low c.1.5 km thick to be located at c. 7 km depth (Zollo et al., 2008; De Siena et al., 2010). Petrological models indicate that it is most likely shoshonitic to trachytic in composition and saturated in CO_2 (Mangiacapra et al., 2008; Arienzo et al., 2010; Mormone et al., 2011a). De Siena et al. (2010) identified a small high attenuation Vp/Vs anomaly at 3.2 km depth below Pozzuoli using a passive seismic dataset from 1983-1984, which may be attributed to either the presence of liquid magma or to a highly fractured rock volume. If it was caused by a magma body, then such a small intrusion would be expected to have solidified in less than 10 years (Woo and Kilburn, 2010; Moretti et al., 2018). As such, there is no conclusive evidence for the presence of magma at depths of less than 7 km at present.

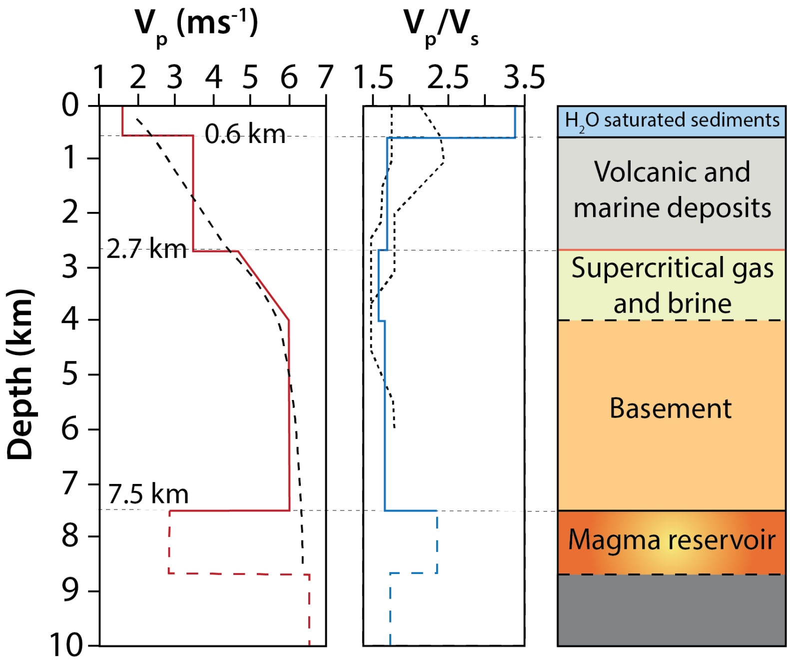


Figure 2.5: Geophysical and structural model of the subsurface structure of Campi Flegrei, as inferred from seismic tomography by Zollo et al. (2008). The left and centre panels show the variation in P-wave velocity and the ratio of P-wave to S-wave velocities through the crust respectively. The right panel is a generalised schematic of the structure of the crust at Campi Flegrei, as inferred from seismic wave velocities. Reproduced from Zollo et al. (2008).

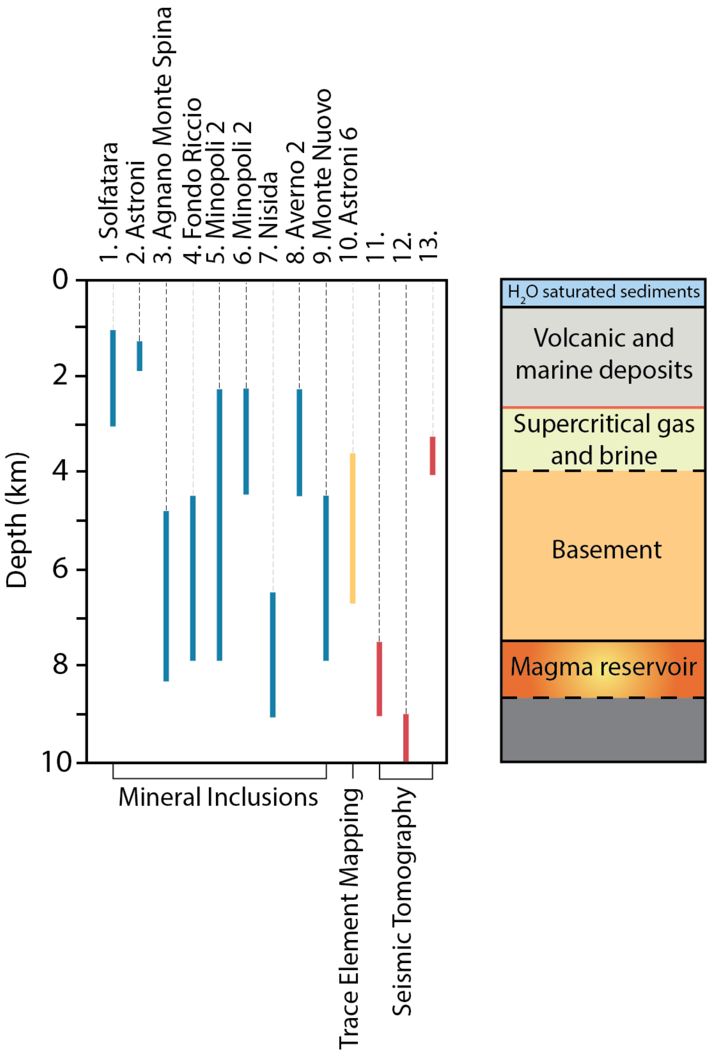


Figure 2.6: Magma storage at Campi Flegrei. The left panel is modified after Stock et al. (2018) and gives depths of magma storage as estimated from petrological (blue bars) and geochemical (yellow bar) analysis of products from the named eruptions, as well as from geophysical imaging (pink bars). Depth ranges for the blue and yellow bars are from; 1. Cipriani et al. (2008), 2. Cecchetti et al. (2003), 3. Arienzo et al. (2010), 4. Mangiacapra et al. (2008), 5. Mangiacapra et al. (2008), 6. Vetere et al. (2011), 7. Arienzo et al., (2016), 8. Fourmentraux et al. (2012), 9. Piochi et al. (2005), 10. Astbury et al. (2018). Depths have been calculated from pressures assuming an average density of 2.3 kg m^{-3} in the upper 4 km of crust (following Stock et al., 2018 and references therein) and 2.6 kg m^{-3} below 4 km (Battaglia et al., 2008). The pink bars represent the depths of magma bodies inferred from seismic tomography studies; 11. Zollo et al. (2008), 12. Nunziata, 2010, 13. De Siena et al. (2017). The right panel is the structure of Campi Flegrei as inferred from seismic tomography by Zollo et al. (2008).

2.2.2 The Hydrothermal Regime

The upper 3 km of the caldera hosts a high-temperature, low-sulphidation hydrothermal system that is driven by a local heat flow anomaly of up to 160 MW m^{-2} (Rosi and Sbrana, 1987; Corrado et al., 1998; Piochi et al., 2015). It is capped by a low-permeability alteration zone that extends across the central caldera, which has been intersected by deep boreholes (up to 3047 m T.D.) at Mofete, San Vito and Agnano (Chelini and Sbrana, 1987; Rosi and Sbrana, 1987; De Vivo et al., 1989). The base of the cap rock is located at depths between 1 and 2 km can be traced by a P-S wave velocity inversion caused by the presence of hydrothermal fluid reservoirs and is observed to bend upwards towards Pozzuoli (Vanorio et al., 2005; Vanorio and Kanitpanyacharoen, 2015). Where the permeability of the cap rock permits, gas and liquid dominated hydrothermal discharges occur, which are the surface expression of the deep system (Fig. 2.7). Active features are distributed within the ring fault zone and along NW-SE trending faults in the Gulf of Pozzuoli, where volcano-tectonic structures act as preferential fluid pathways through the crust (Fig. 2.8a, Antrodicchia et al., 1986; Carlino et al., 2012; Steinmann et al., 2016). The temperature and chemistry of these features are determined by the relative contribution of deep fluids, as well as dilution by meteoric fluids and water-rock interactions with the host rock (Ghiara and Stanzione., 1988; Valentino et al., 2003 and 2004).

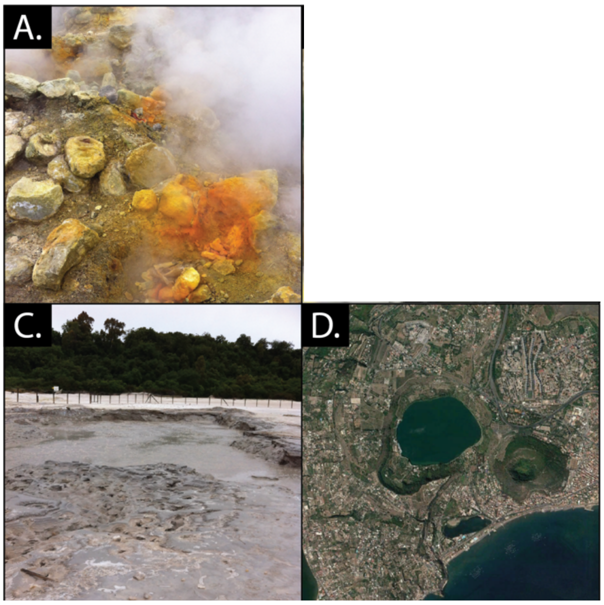


Figure 2.7: Examples of hydrothermal features at Campi Flegrei. A. The Bocca Grande fumarole (L. Smale, 2016). B. Gas emission associated with a thermal water well at Terme Agnano (Googas Catalogue, 2019). C. The Fangaia mudpool at the Solfatara crater (L. Smale, 2016). D. The Lago d'Averno crater lake (GoogleEarth, 2019).

Two principal areas of hydrothermal circulation (Fig. 2.8) can be identified from clustering of surface activity and associated local heat flux anomalies in the Mofete district (160 MW m^{-2}) and at Solfatara-Agnano (120 MW m^{-2}, Corrado et al., 1998; Wohletz et al., 1999). Both areas overlie high seismic attenuation anomalies (V_p/V_s 1.8-2.2) associated with enhanced heat flow that extend down to the hard rock layer at c. 3 km. These anomalies are consistent with the presence of fractured rock volumes saturated in hydrothermal fluids (Aster and Meyer, 1988; Aster, 1992; De Lorenzo et al., 2001; Vanorio et al., 2005; Battaglia et al., 2008; De Siena et al., 2010). The restriction of hydrothermal circulation to these two areas is confirmed by conduction dominated borehole temperature profiles from non-producing geothermal exploration wells drilled at Licola and San Vito (Rosi and Sbrana, 1987; Carlino et al, 2012), as well as the distribution of surface waters with a hydrothermal component (Fig. 2.8b and c, Fig. 2.9, Valentino and Stanzione, 2003 and 2004; Aiuppa et al., 2006; Jasim et al., 2015). The two areas are considered to be hydrologically distinct (Celico et al., 1992a) but the narrow range in $^3He/^4He$ (R) to atmospheric He (Ra) ratios (R/Ra 2.2-3.11) measured from fumaroles across the central caldera, indicates that all surface activity is fed by a common source of magmatic fluid (Tedesco et al., 1990; Caliro et al., 2007; Vaselli et al., 2011). This strongly supports the presence of a laterally extensive and persistent reservoir of magmatic fluids. The locations of present-day hydrothermal discharges are similar to those of Roman-Medieval thermal baths (Giacomelli and Scandone, 2012). The two principal areas of hydrothermal circulation have thus been stable over timescales of 10^2-10^3 years (Fig. 2.10). This is supported by the lack of argillic alteration reported at the surface elsewhere in the caldera and is consistent with a structural control on the locations of hydrothermal activity.

Investigation of hydrothermal circulation in the offshore portion of the caldera is extremely limited and largely restricted to the location of active features. Fluid vents are concentrated in the RFZ and along a major NW-SE striking fault that transects the Gulf of Pozzuoli. Shallow subsurface accumulations of gas and fluids have also recently been identified at depths of less than 200 m from seismic reflection surveys in the southern sector of the RFZ (Steinmann et al., 2016). The strong correlation between the distribution of activity in both onshore and offshore sectors of Campi Flegrei with faulting confirms the structural control on permeability in the caldera and indicates that the location of hydrothermal basins is likely controlled by structures related to volcano-tectonic collapse.

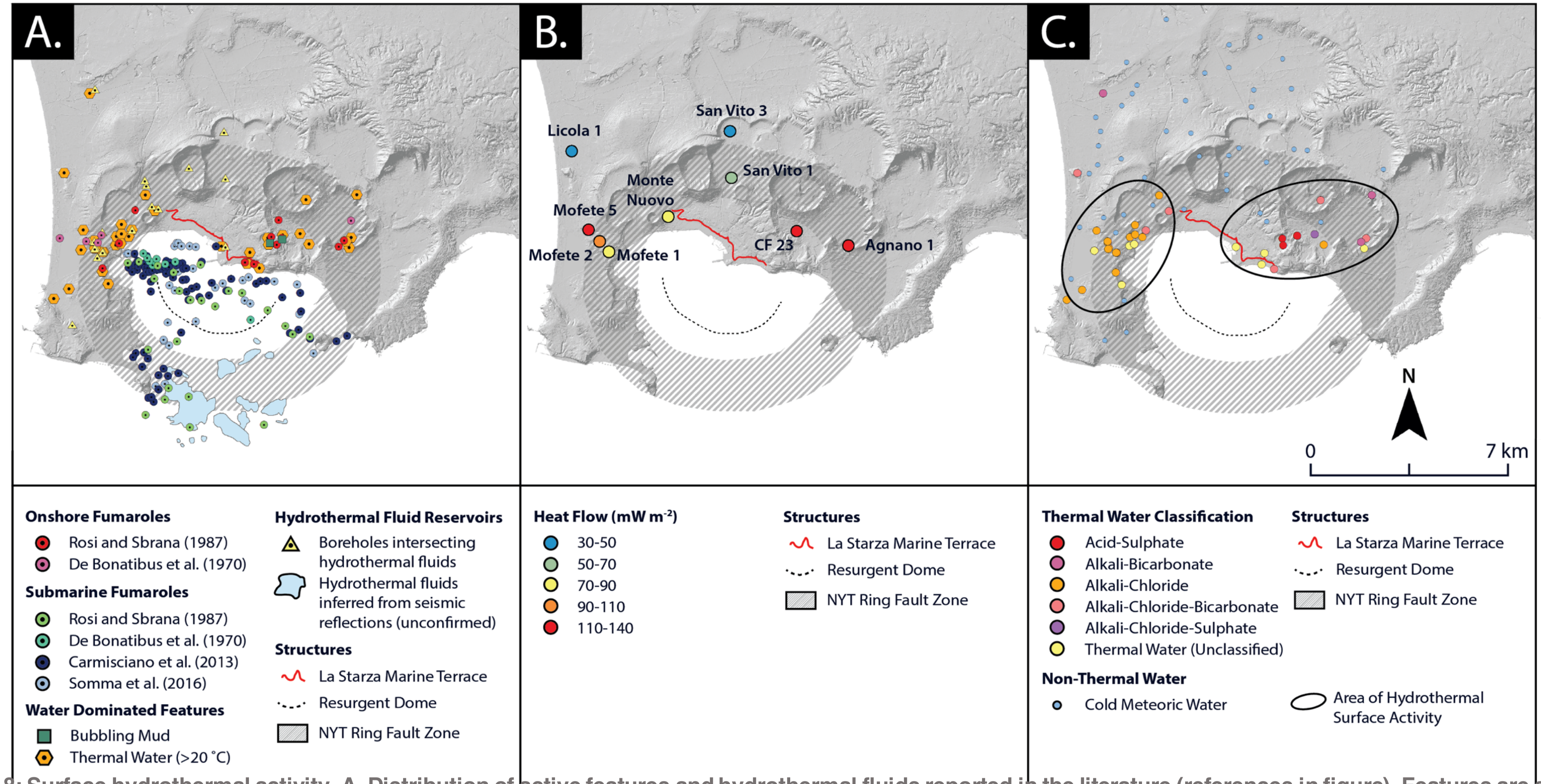


Figure 2.8: Surface hydrothermal activity. A. Distribution of active features and hydrothermal fluids reported in the literature (references in figure). Features are strongly confined to the main collapse structure and are aligned along faults. There is a large degree of uncertainty as to the geolocation of offshore fumaroles and it is likely that there is duplication of some vents. Thermal water locations are from Valentino and Stanzione, (2004) and Aiuppa et al. (2006). Borehole data is from geoThopica, the Italian National Geothermal Database, inferred fluids are from Steinmann et al. (2016). B. Heat flow measured from boreholes. The highest heat flows are associated with the two main areas of activity. Data from Corrado et al. (1998) and Costaldo et al. (2018). C. Distribution and classification of thermal waters. Thermal waters cluster at Mofete (western black ellipse) and Solfatara-Agnano (central-eastern black ellipse). Acid-Sulphate waters are steam heated and contain the highest concentrations of magmatic fluids. Alkali-Chloride waters are dominated by a Na-Cl rich deep hydrothermal component and the Alkali-Bicarbonate thermal waters are fed by a persistent source of magmatic CO2. Other thermal waters are hybrids of these end-member types (Data from Valentino and Stanzione., 2003; 2004 and Aiuppa et al., 2006).

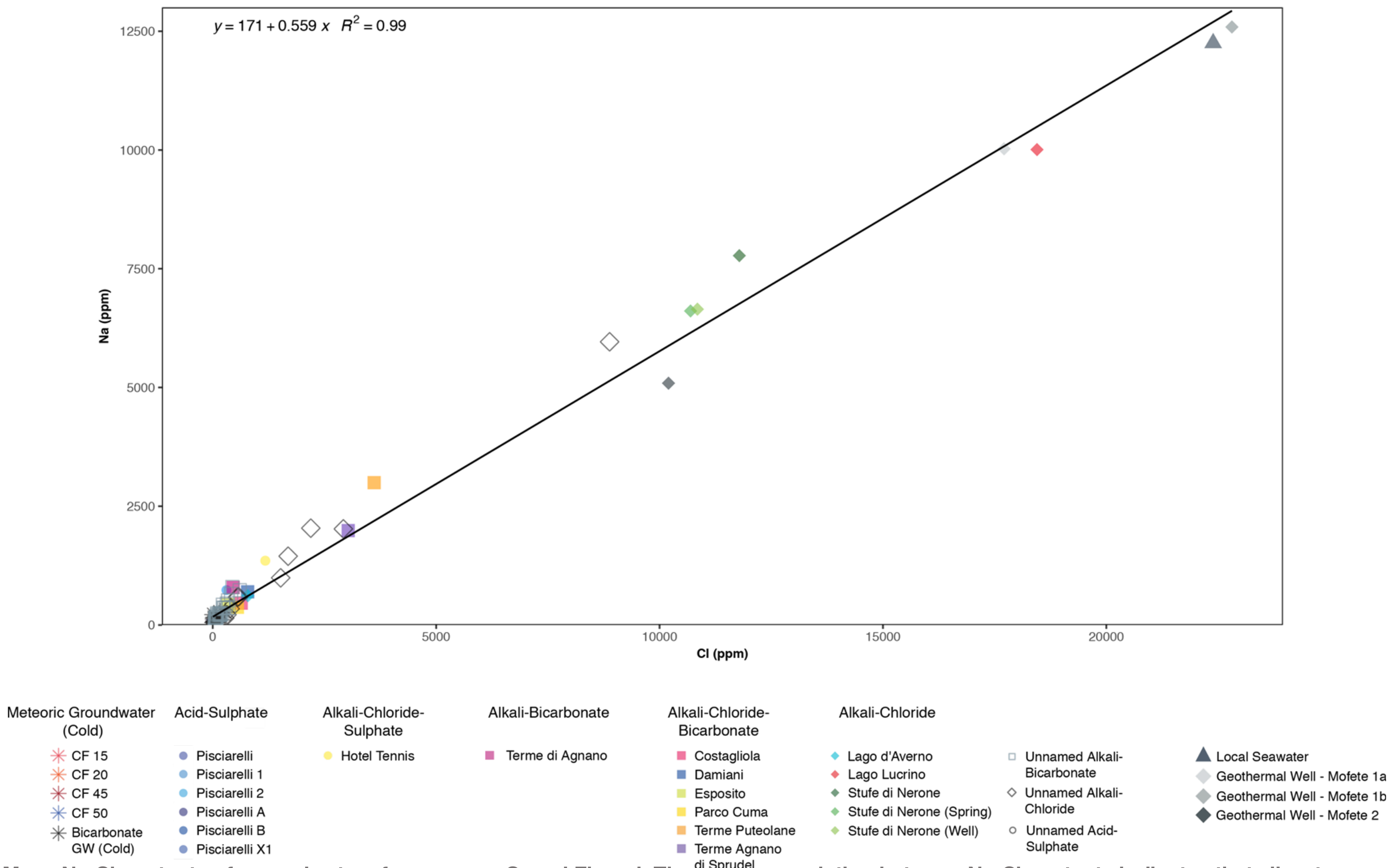


Figure 2.9: Mean Na-Cl contents of groundwaters from across Campi Flegrei. The strong correlation between Na-Cl contents indicates that all waters are variable mixtures between dilute meteoric waters and a Na-Cl hydrothermal brine that has a salinity close to that of hydrothermal brines intersected at < 2 km depth in a geothermal exploration borehole drilled at Mofete in the western sector of the caldera (MF 1). Produced using data from Valentino and Stanzione (2003) and (2004) and Aiuppa et al. (2006).

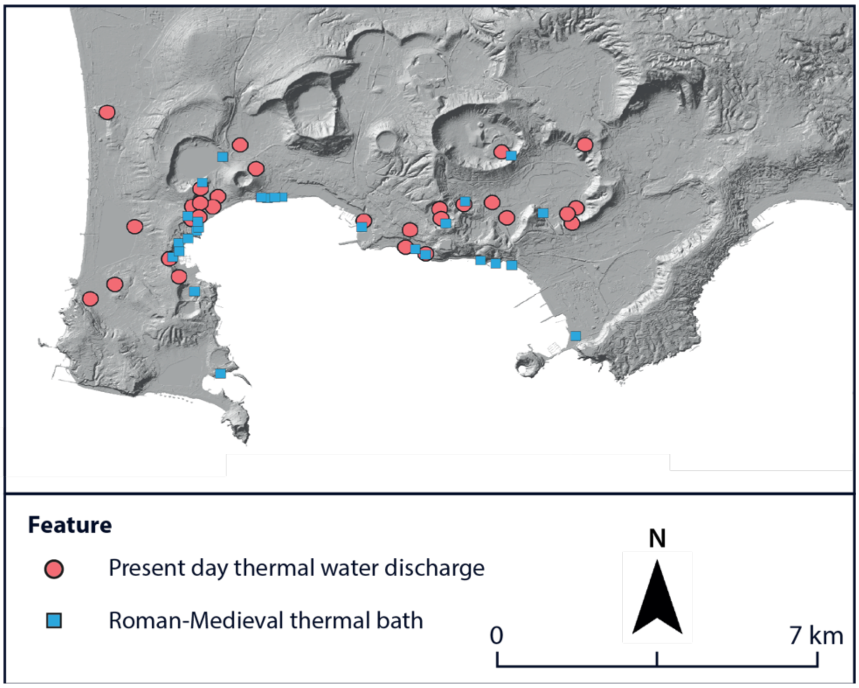


Figure 2.10: Comparison of the locations of Present-day thermal water discharges and Roman-Medieval thermo-mineral baths. The positions of the baths correspond well with the current distribution of thermal water discharges. It may therefore be concluded that the first order locations of hydrothermal features are essentially stable through time and that there has been no major caldera-wide hydrological reorganisation in at least the last c. 2000 years. Roman-Medieval bath locations digitised from Giacomelli and Scandone (2012).

The locations of hydrothermal activity in the Western Sector are controlled by N-S and NW-SE trending faults and fractures (Rosi and Sbrana, 1987; Tarchini et al., 2018). The hottest features (c. 80-100 °C) are fumaroles at the Monte Nuovo cone, and at the Cavone dell'Inferno (Cave of Hell) in the Mofete District. The latter is a c. 10 m deep, c. 60 m long fracture surrounded by a halo of argillic alteration (Rosi and Sbrana, 1987). Other gas dominated features include mofetta, which are cold (< 30 °C) CO_2 seeps, after which the Mofete District was named, and areas of diffuse soil degassing. They have historically not been considered a priority for monitoring and it is only since 2018 that regular temperature measurements have been made at a fumarole at Mofete and at Monte Nuovo (Vesuvius Observatory Bollettino di Sorveglianza Campi Flegrei, Gennaio 2019). Few geochemical analyses of the gases have been published in the literature, but it has been confirmed that CO_2 is the dominant gas in the dry fraction from the Monte Nuovo fumarole (Vaselli et al., 2011). Quantification of the flux is restricted to a single soil diffuse degassing survey of the Fondi di Baia crater that confirmed a low to moderate flux of magmatic-hydrothermal

CO_2 (10.06 ± 1.1 tons of CO_2 per day over c. 200,000 m^2, Tarchini et al., 2018). The lack of knowledge of these features presents a critical challenge both for identification of future changes in the magmatic-hydrothermal system and for assessment of potential non-magmatic hazards.

The thermal waters at Mofete-Baia have been exploited for thermo-mineral bathing since at least Roman times (Fig. 2.11, Yegül, 1996; Giacomelli and Scandone, 2012). They are near-neutral, with maximum temperatures of c. 70 °C and Total Dissolved Solids (TDS) contents of up to 33 000 ppm (Ghiara and Stanzione, 1988; Valentino et al., 2003; Aiuppa et al., 2006). The waters lie on a mixing trend between a cold meteoric source and a Na-Cl rich hydrothermal brine (Baldi et al., 1975; Ghiara et al., 1988; Valentino et al., 2004). A seawater source for the Na-Cl endmember has been discounted due to the elevated temperatures and TDS contents of these waters, as well as their enrichment in B, Li and As, which are transported in the vapour phase of hydrothermal fluids (Valentino et al., 2003; Aiuppa et al., 2006). Analysed surface waters have been classified based on their dominant anion and cation contents as alkali-chlorides or alkali-chloride-bicarbonates (Fig. 2.12; Ghiara and Stanzione, 1988; Valentino et al., 2003 and 2004; Aiuppa et al., 2006). The latter are more strongly diluted by meteoric waters, which accounts for their lower temperatures and TDS contents, as well as the similarity of HCO_3 concentrations with the meteoric endmember. The HCO_3 is generally thought to be derived from meteoric circulation through carbonates but a component may also be due to weak CO_2 degassing (Valentino and Stanzione, 2003). The least dilute waters with the strongest hydrothermal component are those discharged at Stufe di Nerone, which are hot (c. 65-70 °C), saline (c. 20 000 ppm), mature (i.e. in equilibrium with the minerals in the host lithology), alkali-chloride fluids. They are enriched in As and plot close to hydrothermal reservoir fluids intersected by geothermal exploration boreholes at Mofete on B-Cl, Li-Cl (Fig. 2.13) and δD vs. $\delta^{18}O$ plots. As such they have been interpreted as the surface outflow of a hydrothermal reservoir (located at < 2 km depth) composed of seawater boiled at 320 °C that is fed by hot hydrothermal and magmatic gases from greater depth (Valentino and Stanzione, 2003; Aiuppa et al., 2006).

Sub-lacustrine hydrothermal discharges feed the Lago d'Averno crater lake, Lago Lucrino and Lago Fusaro (Valentino and Stanzione, 2003; Aiuppa et al., 2006). They act as a sink for magmatic-hydrothermal CO_2, although the overall dissolved gas chemistry of the lakes is controlled by biogenic processes (Caliro et al., 2008; Tassi et al, 2018). Each has been the site of sudden gas release triggered by limnic overturns, accompanied by transient changes in the water colour and fish mortalities. Since Roman times, Averno has been associated with gas releases that are thought to have contributed to its name, which is derived from *Aomon*, the Greek for without birds (Tassi et al, 2018). The most recent overturns at this lake occurred in 2002, 2003, 2005 and 2017. It has been concluded that they were the result of a cooling of near-surface waters (the epilimnion) to below 7 °C in winter, leading to density stratification and a sudden sinking of the cold-water mass, displacing gas-rich bottom waters upwards (Caliro et al., 2008; Rouwet, 2017). Historical overturns at Lucrino in August 1922 and at Fusaro in August-September 1927 (Signore, 1930) occurred during the summer and as such are unlikely to be caused by the same trigger mechanism. The Fusaro events were investigated by the Vesuvius Observatory who reported that, between 8-10th August 1927, the lake became turbid, turned white and dead fish appeared at the surface, then the water turned red. The lake rapidly recovered before a second overturn on 10-12th September 1927 that followed the same sequence (Signore, 1935). The white water was probably produced by gas bubbles, or by the precipitation of sulphur and carbonates as bottom waters rich in H_2S and CO_2 arrived at the surface, whilst the red water was produced by oxidation of Fe^{2+} (Caliro et al., 2008). In the absence of a thermal anomaly and elevated HCO_3 contents in the water, the overturn was attributed to an injection of hydrothermal fluids and that the trigger was volcanic, despite occurring during a period of quiescence (Signore, 1935). An alternative explanation is that the fluid injection was related to pulses of thermal water known to have occurred along the coast during the mid-19th to mid-20th Century that resulted in "*sudden and rapid*" increases in temperature sufficient to make these areas inaccessible to people (Signore, 1935; Scandone et al., 2010). The pulses occurred during a period of general caldera-wide heating of groundwaters, which has been attributed to a relative shallowing of the water table as a result of subsidence following the Monte Nuovo eruption (Scandone et al., 2010). Possible processes that regulated such pulses include sealing-hydrofracture cycles in the shallow subsurface or tidal forcing (e.g. Berrocoso et al., 2018).

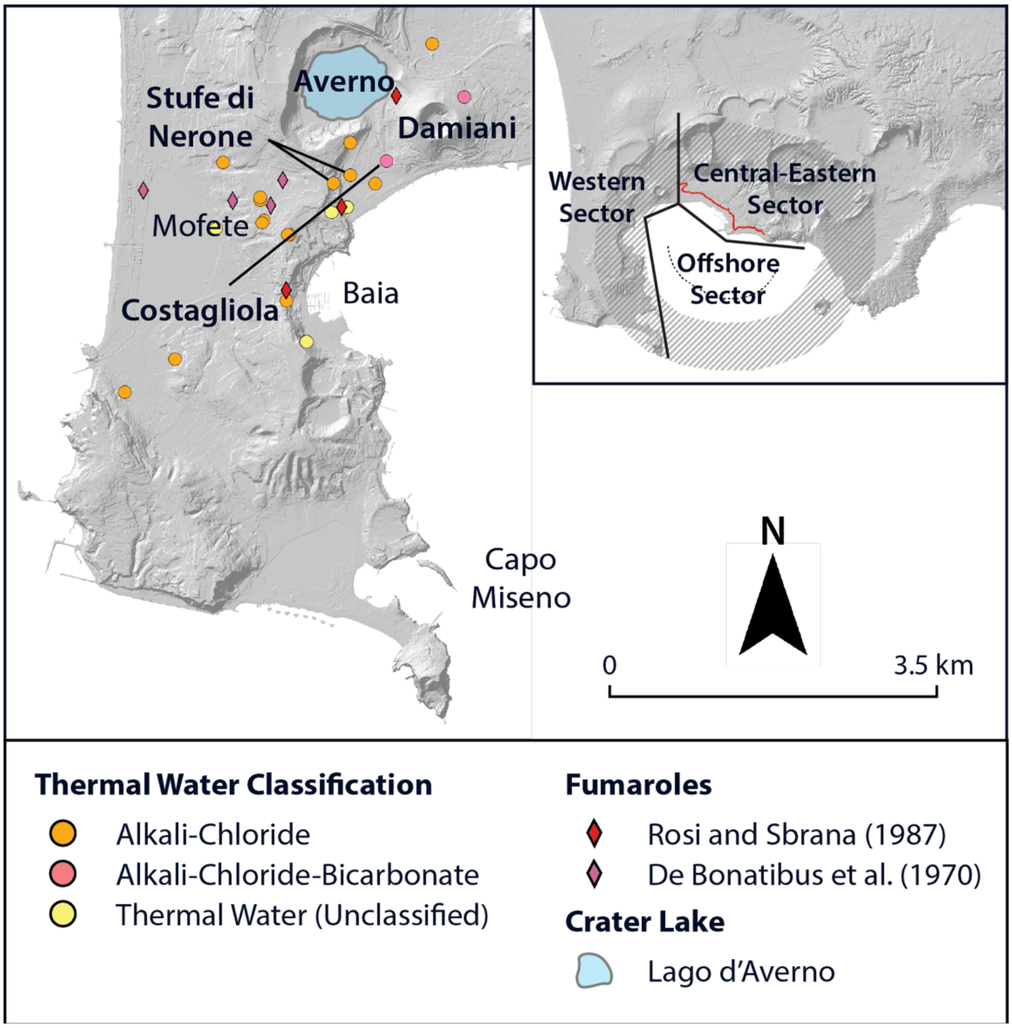


Figure 2.11: Hydrothermal surface activity in the Western Sector of Campi Flegrei. Thermal water locations digitised from Valentino and Stanzione 2003 and 2004; Aiuppa et al., 2006. Fumarole locations digitised from De Bonatibus et al. (1970) and Rosi and Sbrana (1987).

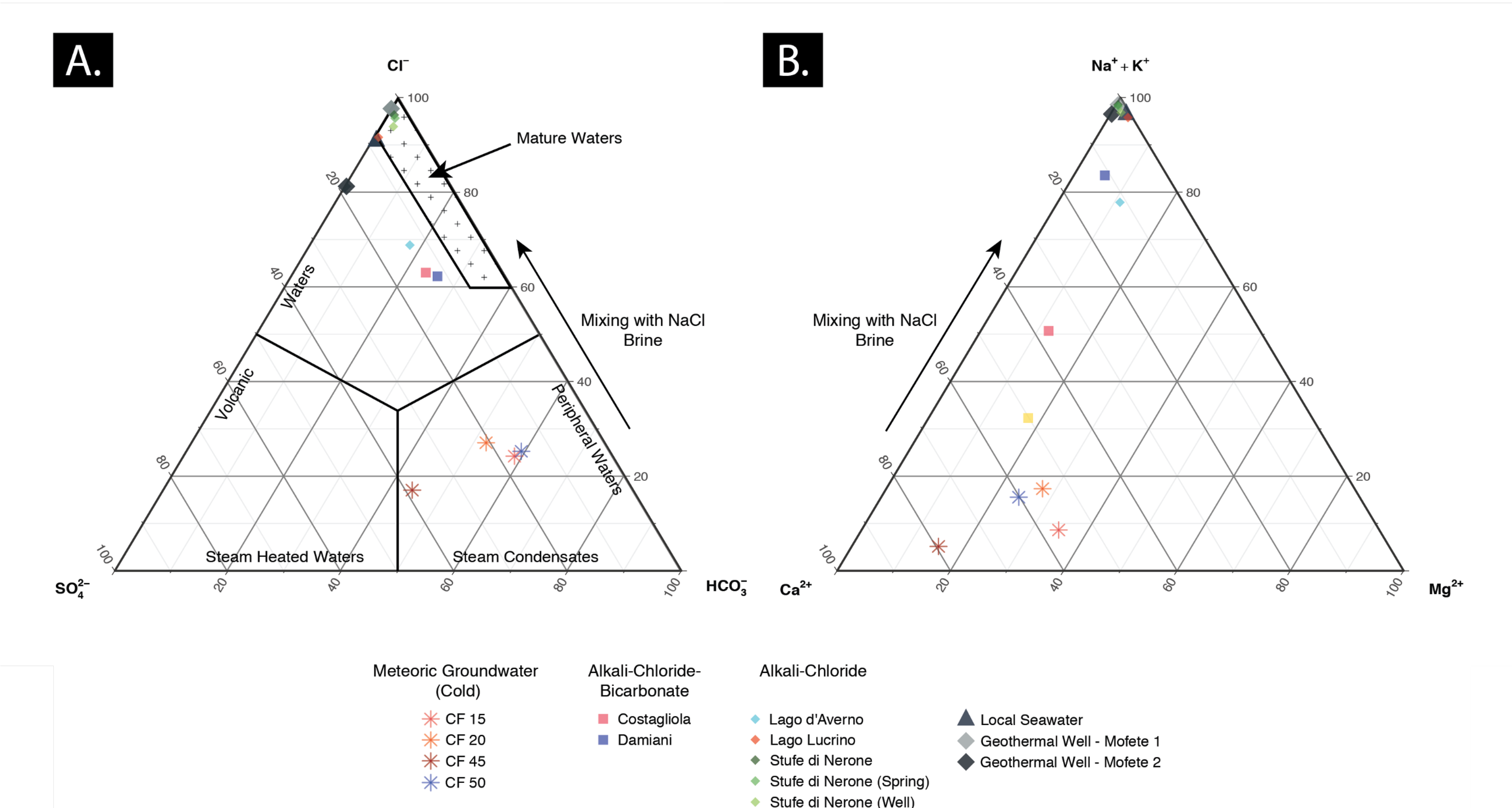


Figure 2.12: Ternary diagrams for classification of water types based on the mean proportions of dissolved major anions (A.) and cations (B.) for Campi Flegrei meteoric waters and thermal waters in the Western Sector. The arrows indicate an increasing component of a Na-Cl brine endmember, which is the hydrothermal reservoir fluid. The thermal waters are alkali-chloride to alkali-chloride-bicarbonate in composition. The Stufe di Nerone waters plot within the field for mature waters, which supports the interpretation that they represent a direct outflow from hydrothermal reservoirs. Data from Valentino and Stanzione (2003) and (2004) and Aiuppa et al. (2006)

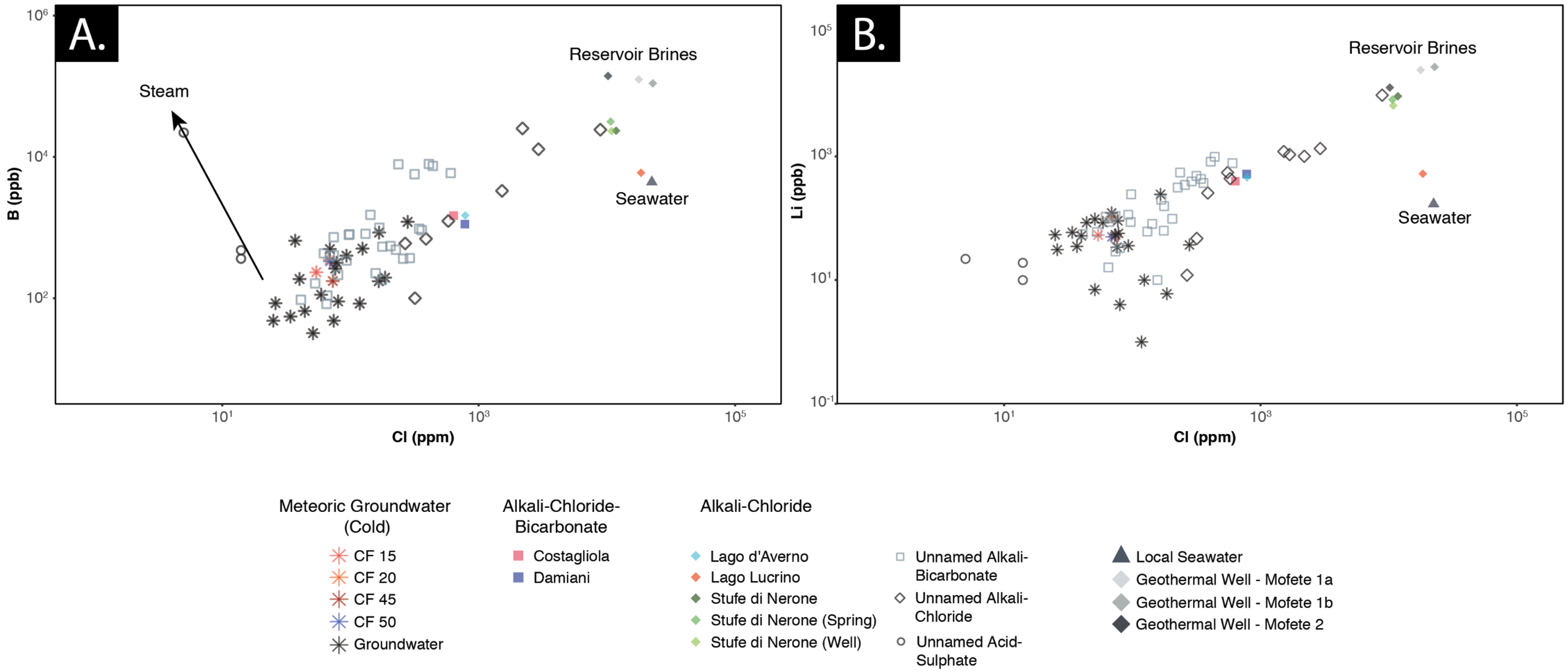


Figure 2.13: B-Cl and Li-Cl contents of Campi Flegrei surface waters. Classified thermal waters are named for their location and those in the Western Sector are highlighted in colour and named in the legend. The groundwaters plot along a compositional trend between a meteoric component and a hydrothermal reservoir brine. Stufe di Nerone waters are the hottest and have a composition close to that of known hydrothermal reservoirs intersected by boreholes at Mofete. As such they are considered to be a direct outflow from hydrothermal reservoirs at depths <2 km. Reproduced from Aiuppa et al. (2006) with additional data from Valentino and Stanzione (2003) and (2004).

Geothermal exploration in 1939-1943 by the Società Anonima Forze Endogene Napoletane (SAFEN) and in 1979-1985 by an Azienda Geologica Italiana Petroli (AGIP) - National Electric Agency (ENEL) joint venture confirmed that the Mofete-Monte Nuovo circulation is a liquid dominated system of near-neutral Na-Cl fluids (Fig. 2.14, Penta, 1951; Carella et al., 1986; Carlino et al., 2012). During the AGIP-ENEL drilling programme, seven deep boreholes were drilled at Mofete (three vertical and four deviated) with a maximum Total Depth (T.D.) of 2.7 km that were part of a wider campaign that also included the deep boreholes at San Vito in the centre of the caldera and at Licola, outside the RFZ (Guglielmineti, 1986). Extensive downhole testing, together with core and fluid analysis from these wells, form the basis of the current geological understanding of the uppermost 3 km of the caldera.

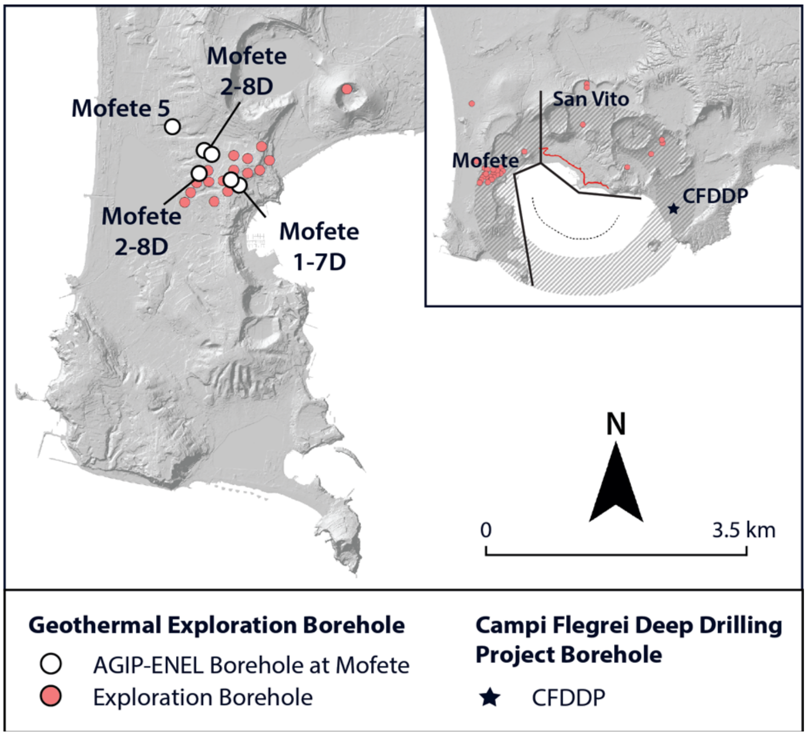


Figure 2.14: Location of geothermal exploration boreholes across Campi Flegrei drilled between 1939-1985. The Mofete boreholes drilled as part of the AGIP-ENEL joint venture deep drilling project are highlighted by the white circles. The Campi Flegrei Deep Drilling Project borehole, which was drilled as part of the International Continental Scientific Drilling Program in 2012 is included and marked by the black star. Well locations from Carlino et al. (2016).

The stratigraphy intersected at Mofete is dominated by tuffs intercalated with subordinate lavas and marine sediments (Rosi and Sbrana., 1987). The lithology has been pervasively altered by fluid-rock interactions, causing an increase in density with a corresponding decrease in porosity with depth (De Vivo et al., 1989; Mormone et al., 2011b). In-situ borehole measurements found that the permeability of these formations was generally less than 10^{-16} m^2 but was as high as 10^{-13}-10^{-2} m^2 in fracture zones, either original or induced by drilling (Piochi et al., 2014). Extensive core analysis has identified an alteration assemblage that is typical of high temperature hydrothermal systems that consists of four mineralogical zones: the argillic, chlorite-illite, calc-aluminium silicate and thermometamorphic zones (Fig. 2.15, Rosi and Sbrana., 1987; Chelini and Sbrana, 1987; De Vivo et al., 1989; Mormone et al., 2011b).

The uppermost argillic and chlorite-illite zones are representative of temperature conditions of less than 250 °C and are characterised by zeolitisation and neogenic clay minerals. Alteration has led to a permeability reduction and an increase in both the shear and tensile strength of the host lithology, forming the cap rock that confines underlying hydrothermal fluid reservoirs (Rosi and Sbrana, 1987; De Vivo et al., 1987; Vanorio and Kanitpanyacharoen, 2015). The upper part of the reservoir region below the cap rock corresponds to the calc-aluminium alteration zone where the lithology becomes brittle due to the precipitation of the hydrothermal minerals K-feldspar, adularia, albite and silica (mainly quartz), from high temperature fluids in pore spaces and open fractures (Rosi and Sbrana, 1987; Mormone et al., 2011b). These minerals indicate reducing conditions and temperatures between c. 220-350 °C. The reservoir extends into the thermometamorphic zone, the top of which coincides with the 325 °C isotherm (Chelini and Sbrana, 1987; Piochi et al., 2014). Porosity increases are observed within formations with a carbonate matrix as a result of decarbonation reactions. Scapolite is also present, which is representative of high CO_2 activity (Rosi and Sbrana, 1987). Throughout the reservoir region, two-phase (liquid + vapour) liquid dominated and two-phase vapour dominated fluid inclusions have been found within the same hydrothermal mineral assemblages (De Vivo et al., 1989; Lima et al., 2017). Their co-existence indicates boiling caused by decompression due to fracturing, which is supported by vein textures that record changes in permeability, as well as the presence of calcite and adularia that forms during boiling (De Vivo et al., 1989). These observations are consistent with non-steady state fluid flux controlled by episodic permeability generation and loss by fracturing and mineral precipitation respectively. It has thus been proposed that the system is a modern analogue of a low-sulphidation epithermal system (e.g. De Vivo et al., 1989; Bodnar et al., 2006; Lima et al., 2009; Lima et al., 2017), where fracturing episodes can be triggered by the transport of superhydrostatically pressured magmatic fluids into the hydrothermal system (e.g. Fournier, 1999; Sillitoe et al., 2003; Sillitoe, 2010).

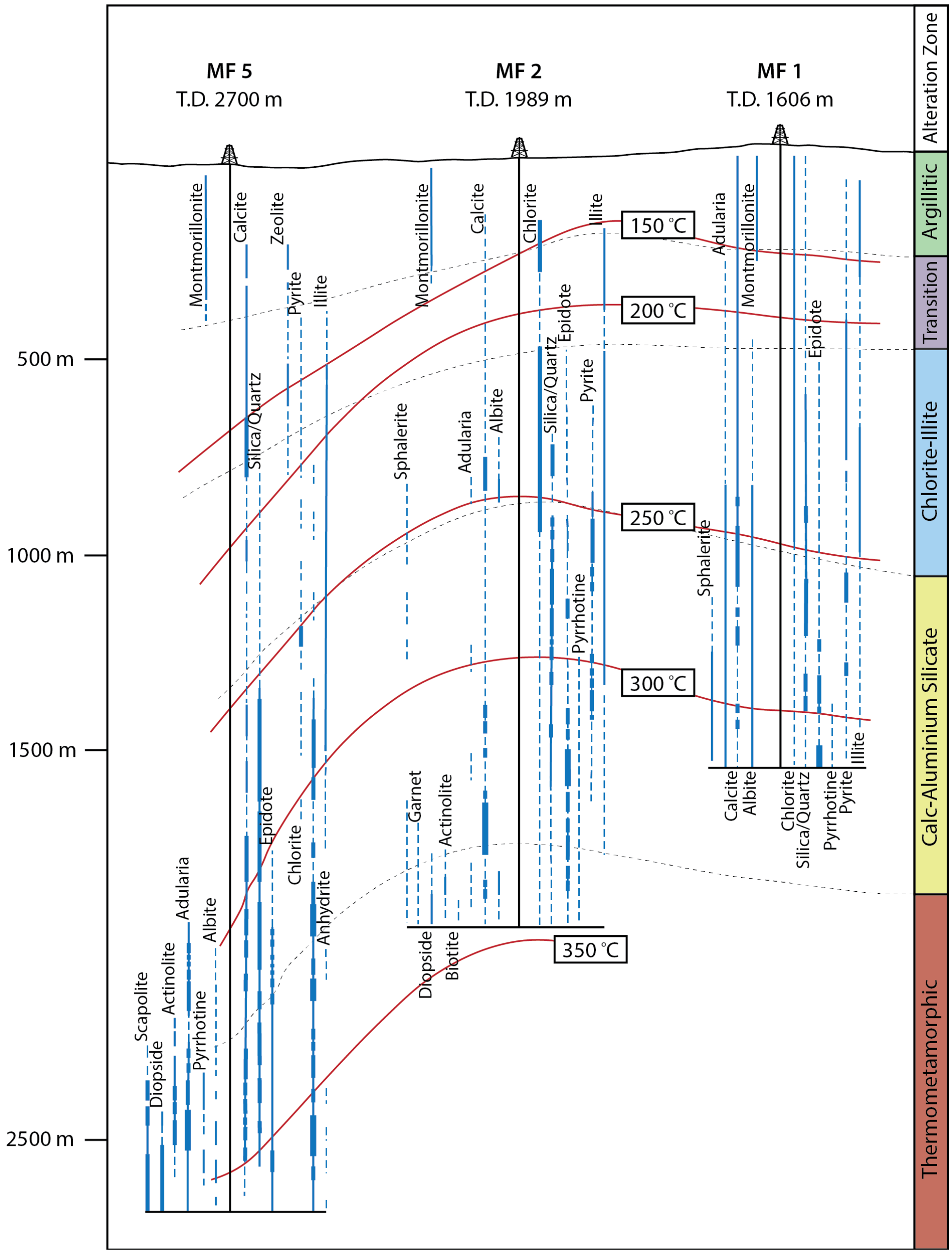


Figure 2.15: The hydrothermal alteration zones intersected in three vertical wells (MF 5, MF2 and MF 1) drilled at Mofete by the AGIP-ENEL joint venture from Rosi and Sbrana (1987), p. 96. The zones are defined based on the alteration assemblages present and the boundaries between them correspond with isotherms. Hydrothermal reservoirs are located in the calc-aluminium and thermometamorphic zones below a clay-mineral dominated cap rock which is represented by the argillic and illite-chlorite zones. The thickness of the blue vertical lines is representative of the mineral abundance.

At Mofete the AGIP-ENEL boreholes intersected three high-temperature (>250 °C), stacked hydrothermal fluid reservoirs localised in fractured formations of tuffs and tuffites at 500-900 m, 1800-2000 m and 2500-2700 m (Table 2.1, Carella and Guglielmineti, 1983). The original model of the origin of the fluids assumed that the hypersaline deep reservoir resulted from the concentration of seawater by boiling and evaporation in a zone of limited recharge. The resultant vapour phase was then thought to migrate towards the surface, interacting with, and heating fluids in the intermediate and shallow reservoirs on ascent (Carella and Guglielmineti,1983). Analysis by Caprarelli et al. (1997), however, found that the deep reservoir is geochemically and isotopically distinct from those at shallower depths and proposed the existence of two hydrothermal fluid reservoirs; one at depths greater than 2 km where the isotopic composition of the brines ($\delta^{18}O$ +5.7-8.3‰, δD -46‰) is indicative of mixing between magmatic fluids and meteoric water, and the second at depths less than 2 km ($\delta^{18}O$ +1.2–3.6‰, δD +1‰) where seawater mixes with steam-heated groundwater and is modified by fluid-rock interaction processes. The meteoric water source is most likely to be the Apennine mountains (Celico et al., 1992a). This model is in agreement with the geochemistry of surface thermal waters in the Western sector (Fig. 2.16, Valentino and Stanzione, 2003 and 2004) and the lack of mixing between the two reservoirs is readily attributable to the density contrast between the fluids as a result of the differences in salinity (Caprarelli et al., 1997). A generalized schematic summarizing the structure of the hydrothermal system at Mofete is given in Fig. 2.17.

Table 2.1: Depths and characteristics of hydrothermal reservoirs intersected by boreholes at Mofete from Carlino et al. (2012). TDS stands for Total Dissolved Solids.

Reservoir	Depth (m)	Temperature (°C)	TDS (ppm at Reservoir Conditions)	Vapour Content
Shallow	500-900	247-308	30000	20%
Intermediate	1300-1900	337	18200	40%
Deep	2500-2700	347	150000	Vapour Dominated

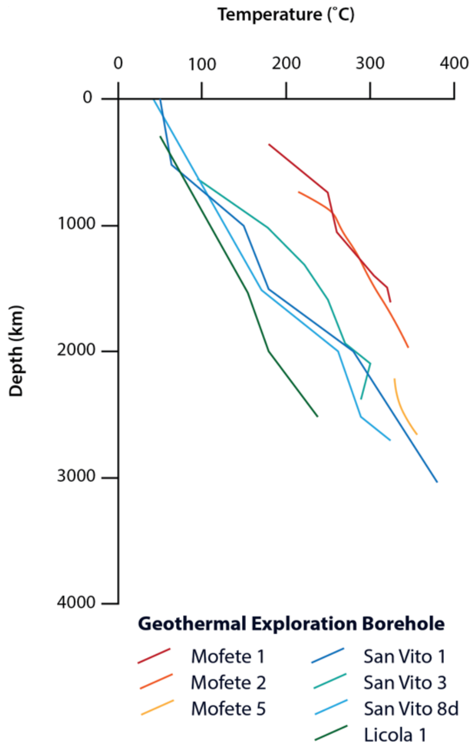


Figure 2.16: Geothermal gradients measured from AGIP-ENEL deep boreholes from Rosi and Sbrana (1987). Very high average geothermal gradients of c. 150 °C km-1 (maximum c. 250 °C km-1) were measured at the Mofete wells where hydrothermal fluids transport mass and heat towards the surface through faults and fractures. The shallower geothermal gradients at the San Vito and Licola boreholes are consistent with conduction dominated heat transport and no significant hydrothermal circulation.

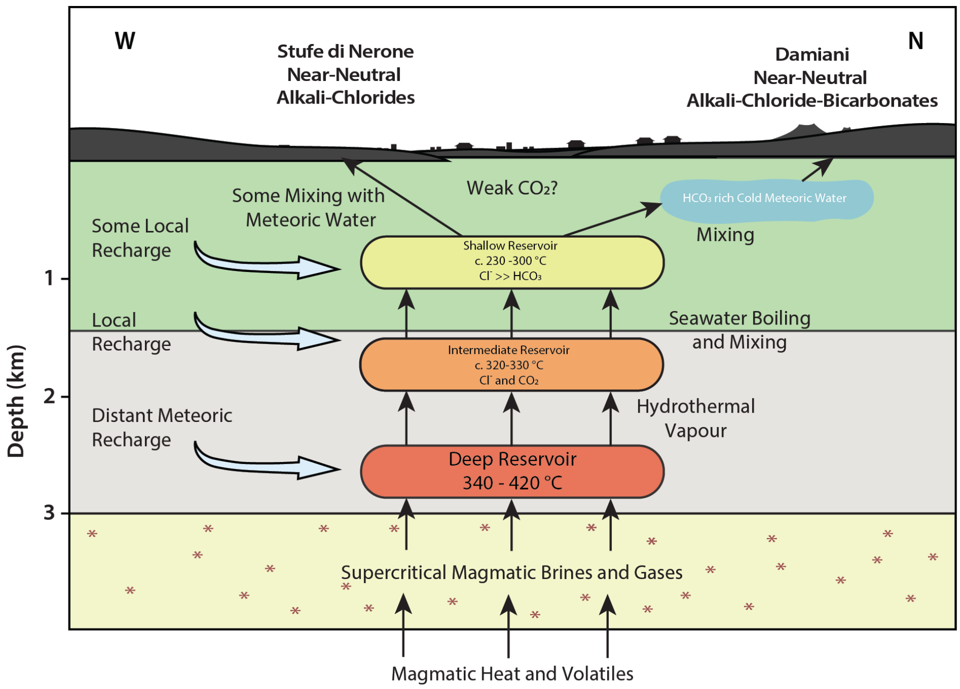


Figure 2.17: Generalised schematic summary of the hydrothermal system in the Western Sector as summarised from the literature.

Surface activity in the Central-Eastern sector is distributed along the coast, aligned with the La Starza Marine Terrace fault system, and around the ring fault zone of the Agnano Monte Spina (AMS) caldera (Fig. 2.18). The principal area of activity in the sector, and the primary site of energy release in the caldera, is the Solfatara-Pisciarelli Diffuse Degassing Structure (DDS), as defined by Chiodini et al. (2001), which is centred on the Solfatara crater and the adjacent Pisciarelli fault system. The crater was formed c. 4.2 ka ago during a period of intense localised volcanism on the SW flank of the AMS caldera that included phreatic to phreatomagmatic eruptions and lava dome emplacement (Isaia et al., 2009, 2015). It is bound by NW-SE, SW-NE and N-S trending faults related to the crater formation. It is considered to be a maar-diatreme that intersected an existing high temperature hydrothermal system, although seismic imaging of the crater has not located a central chimney underlying the maar, as is typical of such structures (Isaia et al., 2009, 2015; Bruno et al., 2017). The Pisciarelli faults are two sub-parallel ring faults related to the AMS collapse that strike NW-SE and dip 60-70° to the NE (Chiodini et al., 2010, 2011).

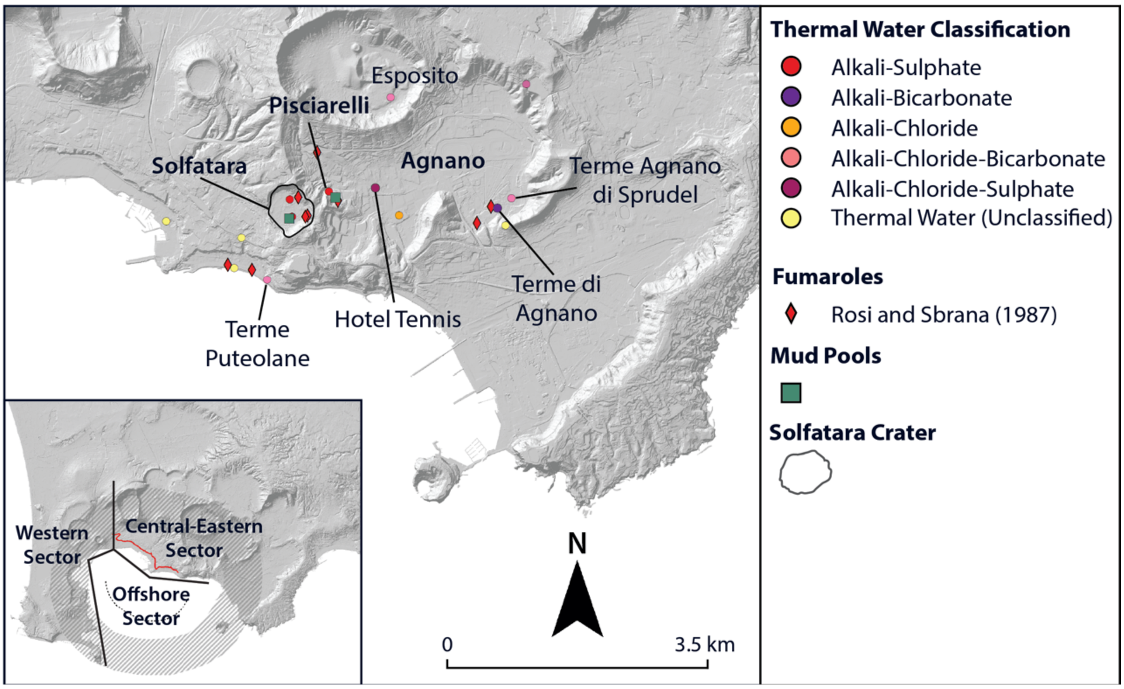


Figure 2.18: Hydrothermal surface activity in the Western Sector of Campi Flegrei. The acid-sulphate pools are and mud pools are characteristic of magmatic-steam heated environments and are the surface expression of a vapour-dominated hydrothermal plume. Surrounding thermal waters are alkali-chloride waters that have been affected by the outflow from the Solfatara-Pisciarelli plume, except at Terme di Agnano where the TDS contents are controlled by HCO_3, due to a persistent source of magmatic CO_2. Thermal water data from Valentino and Stanzione 2003 and 2004; Aiuppa et al., 2006. Fumarole and mud pool locations from Rosi and Sbrana (1987).

The DDS is defined by an area of diffuse soil degassing that broadly corresponds to an area devoid of vegetation. The daily flux of CO_2 has ranged between 750 and 2800 t d^{-1} since measurements began in 1998, with a mean of c. 1390 t d^{-1} (Chiodini et al., 2001; Cardellini et al., 2017). Direct degassing occurs from low-moderate temperature fumarole vents (c. 95-165 °C) that are concentrated along the SE wall of the Solfatara and the Pisciarelli faults where the generally low-sulphidation setting of Campi Flegrei locally grades into a high-sulphidation system (Piochi et al., 2015). The principal vents are the Bocca Grande (c. 165 °C) and the Bocca Nuova (155 °C), which are located at the intersection of NW-SE and NE-SW trending faults in the Solfatara crater and are fed by a single gas reservoir located at 60 m depth (Bruno et al., 2007; Gresse et al., 2018). They are the hottest discharges at Campi Flegrei and the gases have the highest measured R/Ra ratios (2.9-3.1) across the caldera. This indicates that the flux of deep fluid to the surface is greater at this location than anywhere elsewhere in Campi Flegrei (Valentino et al., 2004; Aiuppa et al., 2006). After H_2O, the discharged gases are dominated by CO_2, followed by H_2S, N_2, H_2, CH_4, He, Ar and CO (Chiodini et al., 2001). Magmatic-acidic species such as SO_2, HCl and HF are largely removed by scrubbing (e.g. Symmonds et al., 2001) in the hydrothermal system before reaching the surface (Caliro et al., 2007; Moretti et al., 2013; Aiuppa et al., 2013), although Vaselli et al. (2011) reported moderate amounts of HCl and HF for samples collected in 2004 from Bocca Grande and Bocca Nuova. The $\delta^{13}C$ (CO_2, CH_4) signatures of the gases suggest that the CO_2 originates from degassing of a large, stable magma source, (Allard et al., 1991; Caliro et al., 2007), compatible with the magma reservoir located at c. 7.5 km (Zollo et al., 2008), although a minor contribution from decarbonation of hydrothermal calcite cannot be excluded (Cardellini et al., 2017 and references therein). Helium has a mixed mantle (^3He) and crustal (^4He) origin, whilst N_2 and Ar present in the discharges can have mantle, crustal and atmospheric sources. H_2S, H_2, CH_4 and CO are reactive species produced in the hydrothermal system (Giggenbach, 1980; Symmonds et al., 2001). The H_2O is thought to be derived from a local meteoric component based on δD and $\delta^{18}O$ systematics, which recharges the deep hydrothermal reservoir, as at Mofete (Baldi et al., 1975; Panichi and Volpi, 1999; Caliro et al., 2007). The combined CO_2 flux from the Bocca Grande and Bocca Nuova vents, the main fumarolic area at Pisciarelli and diffuse soil degassing exceeds 2000 t d^{-1}, which is comparable to fluxes measured at erupting volcanoes (Chiodini et al., 2001; Granieri et al., 2010; Cardellini et al., 2017).

Liquid dominated features within the DDS (Fig. 2.19) include the Fangaia mud pool located on an E-W fault close to the centre of the Solfatara crater, and acid-sulphate bubbling pools in the crater and along the Pisciarelli faults (pH 1.4-2.4, 47-96 °C, TDS <10 000 ppm, Valentino and Stanzione, 2003, 2004; Aiuppa et al., 2006; Gresse et al., 2017). This thermal water type is enriched in SO_4 and NH_4 due to a persistent supply of magmatic $H_2S_{(g)}$ and $NH_{3(g)}$ (Fig. 2.20, Chiodini et al. 2001 and 2003, Aiuppa et al., 2006), as well as elements mobilised during biphasic convection of hydrothermal fluids such as B, Li and As (Valentino and Stanzione, 2003; Aiuppa et al., 2006). Such thermal waters are only found at this location and their presence is significant, because they are typical of magmatic steam heated environments (Giggenbach, 1988; Nicholson 1993). Their restriction to the DDS is an indicator that the activity here is the surface expression of an anomalous volume of rock with a permeability high enough for the transport of large volumes (10^3 t d^{-1}) of deep fluids to the surface that boil on decompression, forming a localised vapour-dominated system. A vapour-dominated plume feeding the DDS activity would account for the comparable $\delta^{34}S$ signatures of SO_4 from the pools (–1.3 ± 0.3‰) and H_2S from fumarole gases (–0.3 ± 0.3‰), as well as the similarity in the respective SO_4/NH_4 and H_2S/N_2 ratios, which implies that they are supplied by a common fluid source (Chiodini et al. 2001, 2003; Valentino and Stanzione, 2003; Aiuppa et al., 2006). It would also account for the heavy isotope enrichment of $\delta^{18}O$ in surface discharges (Aiuppa et al., 2006), the high CO_2 flux (Chiodini et al. 2001, 2015a), the large volumes of condensates (Byrdina et al., 2014; Di Giuseppe et al., 2015; Gresse et al., 2017) and associated high thermal energy flux (130 MW, Chiodini et al., 2001), as well as the upwelling of the water table by 80-90 m relative to the surrounding area (Bruno et al., 2007; Petrillo et al., 2013).

At Hotel Tennis, c. 1 km NW of the Bocca Grande and Bocca Nuova vents, thermal waters (74-88 °C) are classified as alkali-chloride-sulphates. They have low Cl/B, Cl/As ratios, similar to those for the acid-sulphate group, and are fed by a common source of H_2S but are near-neutral in pH because of water-rock interactions. Their enrichment in Cl relative to sulphate indicates a return to liquid dominated conditions and that the permeability of the cap rock is no longer sufficient for boiling. This group is representative of the mixing between condensate outflow from the DDS, hydrothermal reservoir brines and meteoric groundwaters on the periphery of the hydrothermal plume (Valentino et al., 1999; Valentino and Stanzione, 2004). The return to liquid dominated conditions is confirmed by a borehole (CF 23) drilled to 1850 m T.D. with a bottom hole temperature of 300 °C at the SW margin of the AMS collapse between the Pisciarelli faults and the Hotel Tennis thermal waters that intersected a liquid dominated reservoir at 1250-1600 m depth (Carlino et al., 2012).

Hydrothermal activity elsewhere in the Central Eastern sector includes low-temperature fumaroles, CO_2 seeps and thermal waters. Most thermal waters are classified as alkali-chloride-bicarbonates, which are moderate temperature (c. 40-60 °C), near-neutral, dilute Na-Cl hydrothermal brines that have been enriched in HCO_3 by interaction with CO_2 (Ghiara and Stanzione, 1988; Valentino and Stanzione, 2003, 2004). These types of waters are typical of those found at the outflows of high-temperature hydrothermal plumes (Giggenbach, 1988; Nicholson 1993). The hottest (c. 90 °C) measured fumaroles are located in the SE sector of the margins of the AMS close to Terme di Agnano, where the only known thermal alkali-bicarbonate waters discharge at c. 45 °C and are associated with CO_2 emission (Vesuvius Observatory Bollettino di Sorveglianza Campi Flegrei, Gennaio 2019; Googas Catalogue). These waters are enriched in HCO_3, NH_4, B and As and have low Cl/B ratios indicating absorption of magmatic gases (Valentino and Stanzione, 2003; Aiuppa et al., 2006). Measurements of gas fluxes from Agnano are absent in the literature but the AMS caldera is known to be a site of significant accumulation of CO_2, for example at the Grotta del Cane, an anthropogenic cavern where temperatures are elevated at c. 50 °C and CO_2 levels are close to 10 vol% (Halliday and Cigna, 2006).

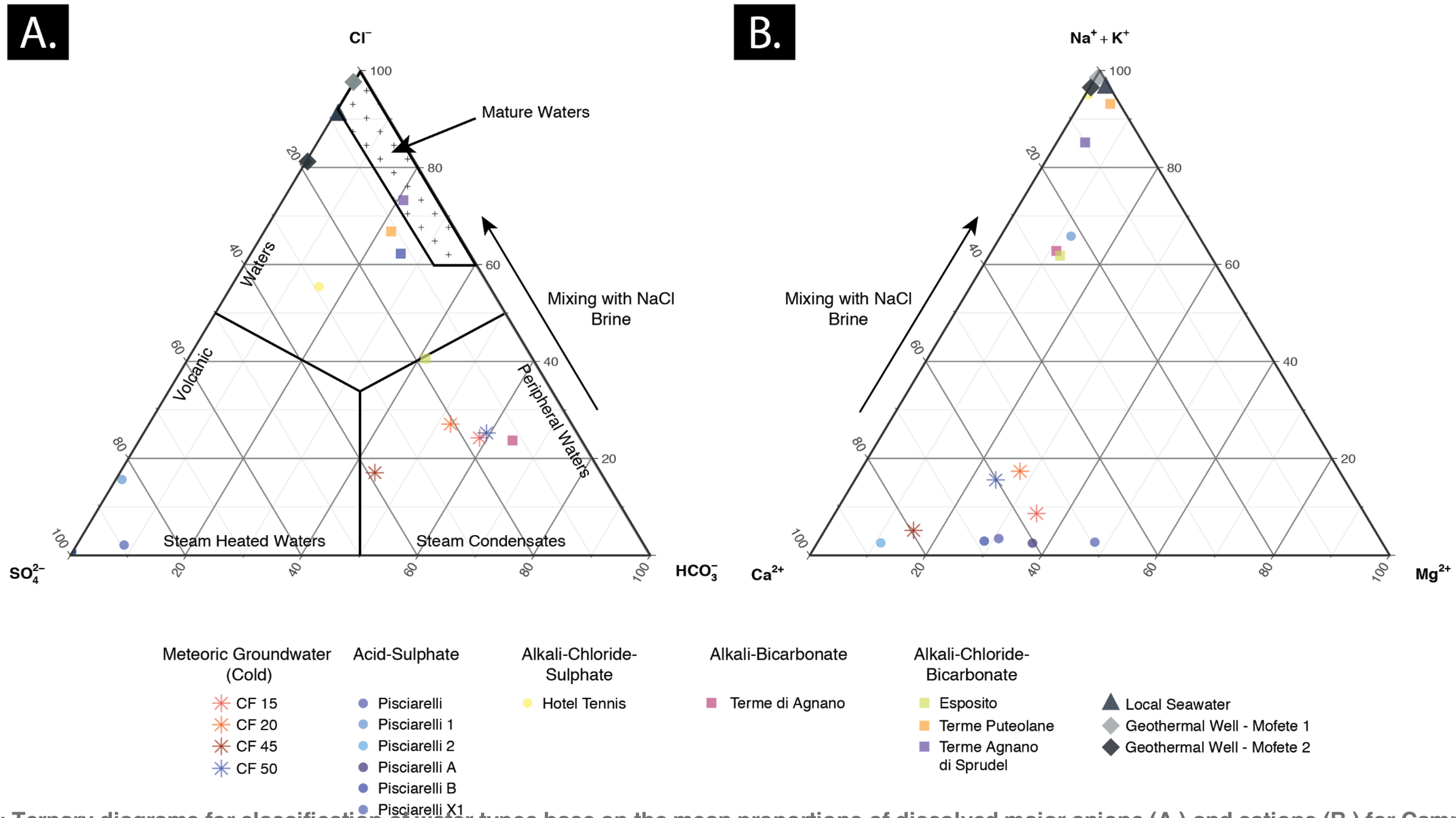


Figure 2.19: Ternary diagrams for classification of water types base on the mean proportions of dissolved major anions (A.) and cations (B.) for Campi Flegrei meteoric waters and thermal waters in the Central-Eastern Sector. The water classification groups consistent with a vapour dominated plume centred on the Solfatara-Pisciarelli DDS, the outflow from which affects surrounding waters. Data from Valentino and Stanzione (2003) and (2004) and Aiuppa et al. (2006).

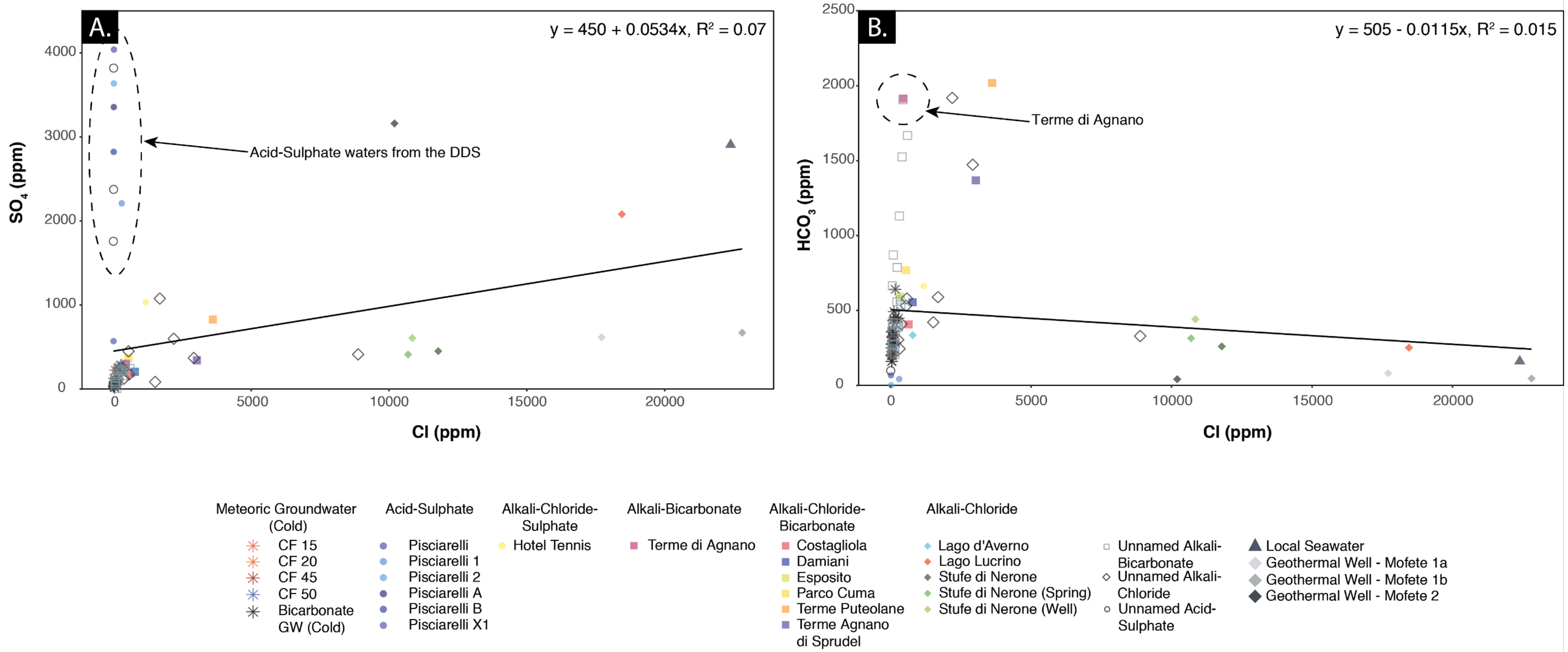


Figure 2.20: SO_4 (A.) and HCO_3 (B.) against Cl contents. The acid-sulphate pools from the Solfatara-Pisciarelli DDS are strongly enriched in SO_4, characteristic of a magmatic-steam heated environment. They are formed by the separation of a high-enthalpy H_2S rich vapour from boiling Na-Cl hydrothermal brine that condenses near the surface, heating shallow meteoric waters. As the fluid condenses, the near surface waters become enriched in SO_4 by the supergenic oxidation of H_2S in the reaction: $H_2S_{(g)} + 2O_{2\,(aq)} = 2H^+_{(aq)} + SO_{4\,(aq)}$. Panel B highlights the strong enrichment in HCO_3 at Terme Puteolane and waters from Agnano. This is due to mixing with a persistent source of CO_2, most likely related to the outflow of the hydrothermal plume at the DDS. Data from Valentino and Stanzione (2003) and (2004) and Aiuppa et al. (2006).

Hydrothermal basins have been identified at two locations in the Central-Eastern sector; below Agnano and the La Starza Marine Terrace at Pozzuoli. The Agnano basin corresponds to a high attenuation (high Qp^{-1}) anomaly, located between 1 and 3 km depth below the AMS collapse (De Siena et al., 2010). It has been inferred to be liquid dominated based on seismic tomography, which is consistent with the findings of the CF 23 borehole (Carlino et al., 2012) and the presence of alkali-chloride-bicarbonate and alkali-bicarbonate surface waters. The Pozzuoli basin is elongated W-E, parallel to the La Starza Marine Terrace and is only known from geophysical imaging (Fig. 2.21, Aster and Meyer, 1992; De Siena et al., 2010). It is defined by an extensive low S-wave velocity structure, with a high V_p/V_s ratio (1.9-2.6) and low Q_p and Q_s values, that coincides with a gravity low (Aster and Meyer, 1988; De Lorenzo et al., 2001; Vanorio et al., 2005; Battaglia et al., 2008; De Siena et al., 2010; Capuano et al., 2013; Caló and Tramelli, 2018). These observations are consistent with the presence of a strongly fractured volume of low-density rock saturated with liquid and gas (De Siena et al., 2010; Caló and Tramelli, 2018). The higher vapour content inferred at Pozzuoli from geophysical imaging, relative to that at the Mofete-Monte Nuovo hydrothermal circulation and below Agnano, suggests a reduction in the confining pressure as a result of a higher permeability. Given the correspondence between the Pozzuoli basin and an actively deforming area that undergoes episodic uplift and subsidence (see Chapter 3, section 3.2.1), a plausible mechanism for permeability generation and maintenance at this location is the accommodation of these ground movements by brittle deformation, forming fractures and thus increasing permeability.

Figure 2.21: Schematic of the magmatic-hydrothermal system based on the comparison of the results from attenuation tomography and seismic velocities from De Siena et al. (2010).

Connecting the Pozzuoli basin to the surface is a near-vertical, low density, high S-wave velocity (Vs) anomaly, which is highly resistive (50-100 Ω m) and a unique feature at Campi Flegrei (Fig. 2.22, Battaglia et al., 2008; Zollo et al., 2008; De Siena et al., 2010; 2018). It is caused by a pervasively fractured structure that intersects the cap rock and transports deep fluids from a gas-rich zone at c. 2.25 km to the DDS at the surface (De Siena et al. 2010; Troiano et al., 2014; Chiodini et al., 2015a). The high resistivity of the structure and corresponding low Vs anomaly is consistent with the channelization of vapour-dominated fluids and the interpretation from surface features of the presence a hydrothermal plume supplying activity at the DDS. The feeding structure is considered to be a magma plumbing system that fed volcanic vents during the intense localised activity at the SW sector of the AMS ring faults between 4.1-4.4 ka BP (Isaia et al., 2015; De Siena et al., 2018), implying that the DDS did not previously exist. Isaia et al. (2015) have directly related it to the proposed Solfatara maar-diatreme; however the surface expression of the DDS is not confined to the crater and the high Vs velocity structure that defines the feeding system is offset to the SE. Similar high velocity structures have also been located offshore in the Western sector and under the volcanic centres of Monte Gauro and Astroni, although none are associated with significant transport of hydrothermal fluids to the surface (De Siena et al., 2018). The anomalous permeability at the DDS feeding structure that allows for the high fluid flux is most likely related to brittle deformation related to the aforementioned ground deformation and its location on the active Pisciarelli fault system (Aster et al., 1992).

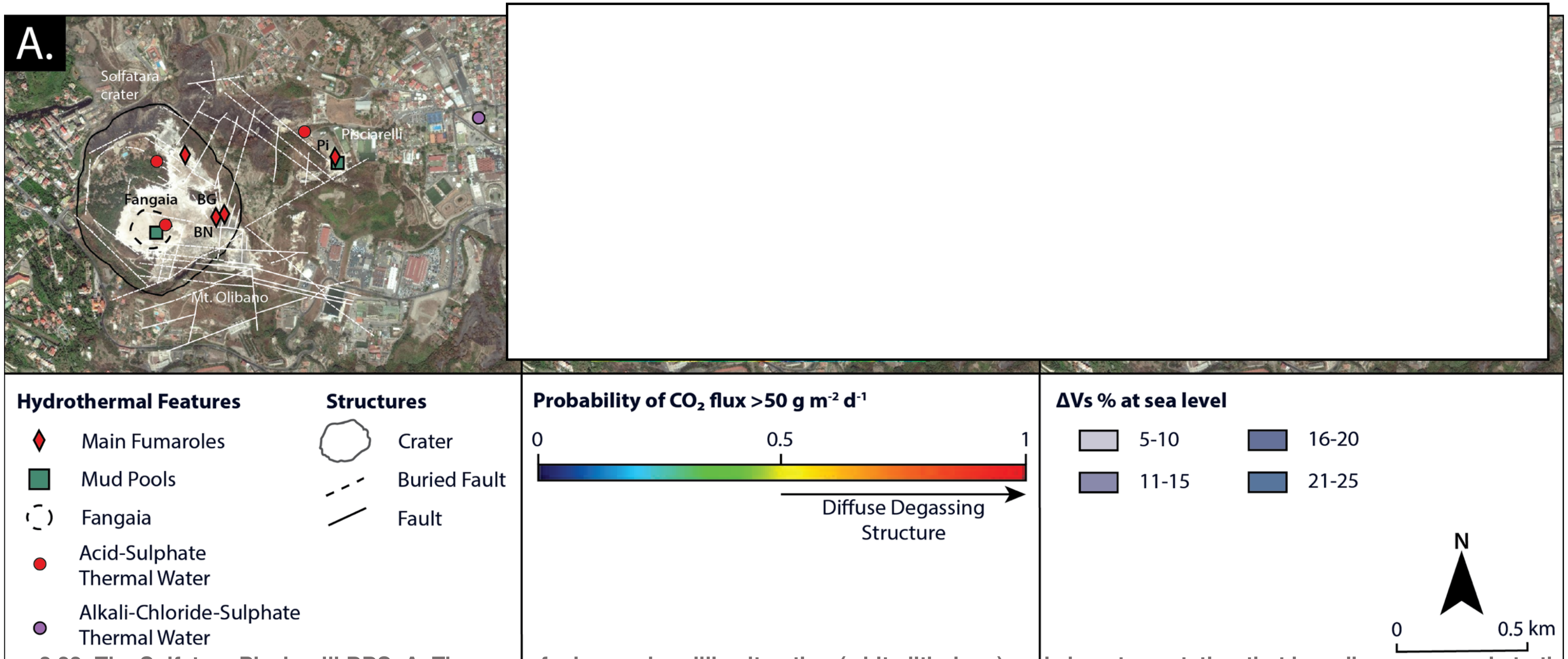


Figure 2.22: The Solfatara-Pisciarelli DDS. A. The area of advanced argillic alteration (white lithology) and absent vegetation that broadly corresponds to the DDS structure. The alteration is characteristic of a magmatic-steam heated environment and is caused by acidic fluids reacting with the host rock. The neogenic mineral assemblage is dominated by supergenic sulphate, native sulphur, amorphous silica (mainly Opal-A) and alunite, which give the distinct white colouring (Piochi et al., 2015; Mayer et al., 2016; Montanaro et al., 2017). B. The area of soil diffuse degassing of magmatic-hydrothermal CO_2. CO_2 fluxes of less than 50 g m^{-2} d^{-1} are assumed to have a biological source (Chiodini et al., 2001; Cardellini et al., 2017). C. The anomalies in Vs velocities at sea level that define the feeding structure of the DDS (Chiodini et al., 2015a; De Siena et al., 2017). The anomalies refer to the percent variations with respect to the mean value at each depth. Fault locations and CO_2 flux map are from Cardellini et al. (2017), Vs anomalies are digitised from Chiodini et al. (2015a). Base map: Esri, HERE, Garmin, © OpenStreetMap contributors, and the GIS user community, Source: Esri, DigitalGlobe, GeoEye, Earthstar Geographic, cNESAirbus DS, USDA, USGS, AeroGRID, IGN, and GIS user community.

The prevailing geochemical model of the hydrothermal plume feeding the DDS was developed by Caliro et al. (2007) on the basis of equilibrium temperature-pressure estimates of gases discharged at the surface, H_2/Ar geothermometry, gas oxygen isotopic compositions and thermodynamic modelling using the TOUGH 2 simulator. It is a refinement of earlier models developed from 1984 onwards based on equilibrium temperatures of gases from Bocca Grande and Bocca Nuova (e.g. Cioni et al., 1984, 1989; Chiodini and Marini, 1998; Chiodini et al., 2001). The essential features of the model (Fig. 2.23) are (i) a source of magmatic fluids located at 3-4 km that corresponds to the supercritical gas and brine reservoir, (ii) a mixing zone at 2-3 km where magmatic and meteoric fluids mix under oxidising conditions at ≥ 360 °C, 20-25 MPa forming a superheated vapour, (iii) a vapour dominated plume where reducing hydrothermal conditions dominate, and (iv) a single phase gas zone (SPGZ) at c. 200 °C located between 100-300 m depth that feeds the fumarole vents. The SPGZ was hypothesised from the results of TOUGH 2 modelling of the Solfatara plume (e.g. Todesco et al., 2003), and its presence has subsequently been questioned based on the results of geoelectrical studies of the upper 400 m of the system that have located H_2O in both the gas and liquid phase at these depths (Byrdina et al., 2014). The gases that are discharged from the Solfatara fumaroles are mixtures that are on molar average 26% magmatic fluids and 74% vapourised hydrothermal fluids (Caliro et al., 2007). A generalised schematic summarising the structure of the hydrothermal system below the central caldera is given in Fig. 2.24.

Figure 2.23: Conceptual geochemical model of the Solfatara-Pisciarelli hydrothermal plume (from Caliro et al., 2014). Magmatic fluids enter the deep hydrothermal reservoir at approximately 3 km depth where they mix with meteoric H_2O. These deep fluids ascend towards the surface in a vapour-dominated plume. The reduction in confining pressure in the shallow subsurface leads to the formation of the hypothesised Single-Phase Gas Zone (SPGZ).

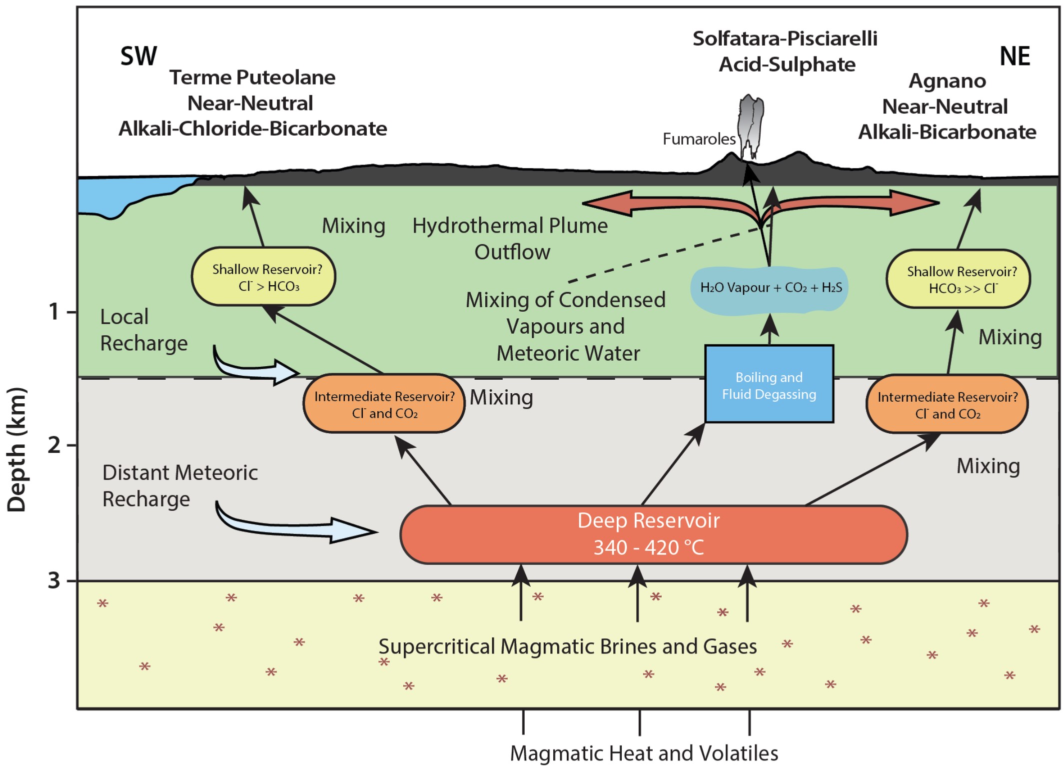


Figure 2.24: Generalised schematic summary of the hydrothermal system in the Central Eastern Sector as summarised from the literature.

Surveys of the offshore sector conducted since the 1970s have identified a large number of active hydrothermal fluid vents (Fig. 2.25, Versino, 1972; Pescatore et al., 1984; Bruno et al., 2004; Passaro et al. 2016; Steinmann et al. 2016; Somma et al., 2016). However, analysis of the fluids and the feeding system of these features is extremely limited in the literature. Sampled vents have discharge temperatures between 18-100 °C and the isotopic compositions of C and He are similar to those from subaerial fumaroles, consistent with a common origin for the magmatic component in hydrothermal fluids across Campi Flegrei (Tedesco et al., 1989; Vaselli et al., 2011; Di Napoli et al., 2016). The hottest known fumarole is Le Fumose (c. 100 °C), which is one of a series of vents located along the Secca delle Fumose submarine relief at the uplifted eastern margin of a N-S graben-like structure that formed during the Monte Nuovo eruption. This feature is associated with a wide pH and CO2 anomaly and is the largest known offshore degassing area. The associated CO2 flux is 50 t d-1, similar to recently active volcanoes such as Poas (Costa Rica), Soufriere Hills (Montserrat) and Hekla (Iceland, Di Napoli et al., 2016 and references therein). The heat flux is also significant at 80 MW and comparable to the margins of the main onshore degassing areas at Mofete and Solfatara-Pisciarelli (Di Napoli et al., 2016).

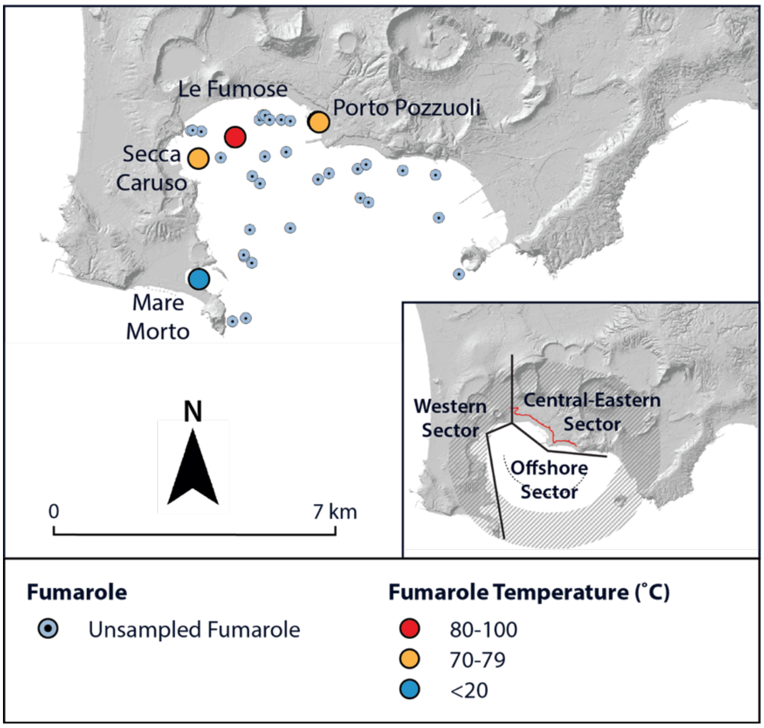


Figure 2.25: Offshore fluid vents. Unsampled vent locations from Somma et al. (2016). Only vents with the least uncertainty as to location are included. Temperature data for the labelled fumaroles are from Tedesco et al. (1990) and Vaselli et al. (2011).

2.4 Summary

The Campi Flegrei magmatic-hydrothermal system is typical of that of large calderas. It can be subdivided into the magmatic regime, where fluids are dominantly magmatic and pore pressures approach lithostatic, and the overlying hydrothermal regime where principally meteoric fluids circulate at close to hydrostatic pressures. The primary source of heat and magmatic fluids through the system is a magma reservoir located at c.7.5 km depth. The transport and storage of magma at shallower depths is known from the presence of intrusions in cores from exploration boreholes that penetrate the upper 3 km of the crust and petrological analysis of erupted material. However, there is no evidence for the presence of a magma body of >1km^3 existing at depths shallower than that of the main reservoir at Present. The uppermost part of the magmatic regime contains a laterally extensive reservoir of overpressured magmatic fluids (brines and gases) that corresponds to the Brittle-Ductile transition where fractures create fluid storage space, and continuously supplies the overlying hydrothermal system with magmatic gases (e.g. CO_2) at a background rate. Veining in the deep part of the hydrothermal system evident in borehole cores suggests the occurrence of episodic enhanced fluxes of magmatic fluids into the shallow crust that is modulated by the opening and sealing of fractures.

In the hydrothermal regime a characteristic alteration assemblage has developed. The hydrothermal circulation is confined to a brittle zone at c.1.5-3 km depth where permeability is fracture controlled, between a cap rock and the low permeability transition zone that defines the boundary between the magmatic and hydrothermal regimes. Two main upflows have been distinguished; at Mofete in the west of the caldera, and below Pozzuoli-Agnano in the central-eastern sector. The primary control on the locations of these upflows are the ring faults and volcano-tectonic structures related to the collapse of the Neapolitan Yellow Tuff (NYT) caldera. Heat flow at the surface is greater at Pozzuoli than at Mofete, indicating more efficient heat transport by fluids and therefore a higher permeability in this region. This is consistent with the density distribution of faults at the surface. Connecting this reservoir to the surface is the Solfatara-Pisciarelli Diffuse Degassing Structure (DDS), the location of which is controlled by a localised, pervasively fractured and vertically extensive rock volume. This provides the most direct pathway for deep fluid transport to the surface in the caldera, as evidenced by the high temperature and magmatic fluid contents of fumaroles at this location relative to elsewhere at Campi Flegrei. Surface activity at Solfatara-Pisciarelli would therefore be expected to show the greatest, and most rapid, physico-chemical variations in response to changes in the deep hydrothermal circulation. The hydrothermal reservoirs below Pozzuoli also coincide with the area of maximum deformation in the caldera where phases of uplift and subsidence have been concentrated. It is suggested that active ground deformation, is the primary mechanism for the creation and maintenance of fractures in this part of the caldera and as such, must exert a critical influence on fluid flow. A schematic model summarising the essential features of the magmatic-hydrothermal system is presented in Fig. 2.27.

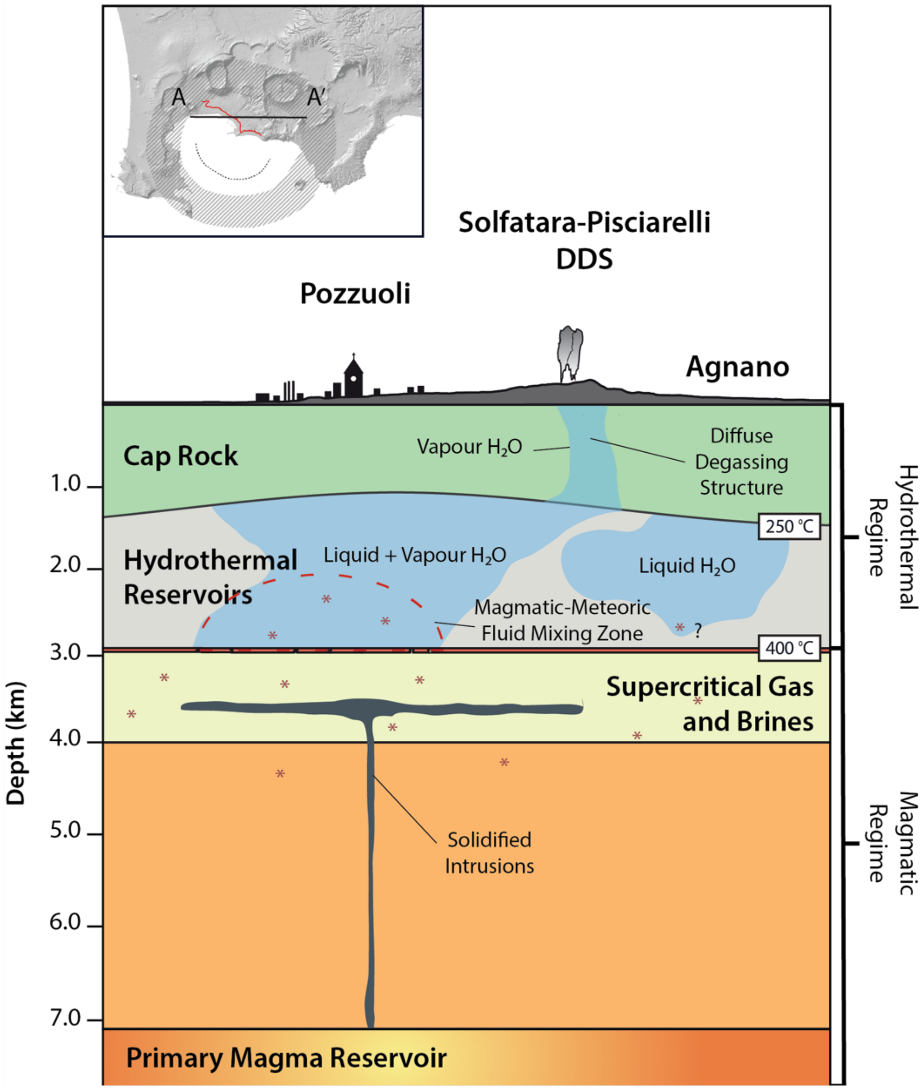


Figure 2.2.26: Generalised schematic summarising the structure of the Campi Flegrei magmatic-hydrothermal system. Approximate locations of the hydrothermal basins from De Siena et al. (2010).

Chapter 3

Models of the Role of the Hydrothermal System in Ground Movements from 1982

Conventional models of uplift in 1982-1984 have variably attributed it to the pressurisation of a magma body (Berrino et al., 1984; Bianchi et al., 1987; Dvorak and Berrino, 1991; Bellucci et al., 2006b; D'Auria et al., 2015; Amoruso et al., 2017), hydrothermal or magmatic fluids (De Natale et al., 1991; Todesco et al., 2003; De Vivo and Lima, 2006; Bodnar et al., 2007; Troiano et al., 2011; Chiodini et al., 2015a; Moretti et al., 2017) or some combination thereof (Gaeta et al., 1998; Orsi et al., 1999a; De Natale et al., 2006; Battaglia et al., 2006; Gottsman et al., 2006). A general consensus has emerged, however, that the aseismic subsidence that followed the uplift most likely represents a reduction in pore pressure in the upper 3 km in the crust.

Two categories of model that consider post-1984 subsidence to result from a loss of pore pressure are defined here based on the causative process. The first assumes the depressurisation of the hydrothermal system by lateral outflow, and the second attributes it to the escape of magmatic fluids from below a hydrological barrier following an episode of fracturing. Both groups are constrained by the geodetic signal during the uplift-subsidence sequence and consider subsidence to be caused by an increased coupling between the magmatic-hydrothermal systems during the uplift. A third set of models has emerged since the resumption of uplift in 2004, which are constrained by compositional changes in Solfatara fumarole gases. This group considers the deformation sequence between 1984 to Present (2019) to reflect a depressurisation and re-pressurisation of the hydrothermal system.

This chapter evaluates the success of these models in accounting for the observed temporal trends in monitoring parameters since 1982 and the emergence of the slow ground oscillation after 1984.

3.1 Models of the 1982-1984 Uplift and Following Subsidence

3.1.1 Pressurisation and Depressurisation of the Hydrothermal System

Models assigned to this category attribute the 1982-1984 uplift to the pressurisation of pore fluids in the hydrothermal system triggered by either a discrete pressure source at its base or an injection of magmatic fluids. Pressurised hydrothermal fluids propagate outwards from the source during the uplift, then subsidence follows as overpressure in the hydrothermal system is dissipated by lateral outflow. Within this group models can be categorised according to the trigger of unrest; (i) a pressurisation of a stationary magma body located at 4 km depth (Gaeta et al., 1998; Orsi et al., 1999a; Castagnolo et al., 2001; De Natale et al., 2001; Troise et al., 2001); (ii) magma degassing (Todesco et al., 2003; Chiodini et al., 2003; De Natale et al., 2006; Battaglia et al., 2006; Troiano et al., 2011); and (iii) regional tectonism (Barberi et al., 1984; Martini et al., 1984; Lupi et al., 2017).

(i) Magma pressurisation-trigger models initiate ground movement in two ways. In the first, a three-step sequence of ground movement is initiated by a volume change in a magma body (Fig. 3.1a; Gaeta et al., 1998; Castagnolo et al., 2001; De Natale et al., 2001; Troise et al., 2001; Troise et al., 2001). The first step corresponds to the initial aseismic uplift in 1982 (c. 0.1 m), during which the inflation of the magma body induces a change in the vertical stress field, pressurising the base of the hydrothermal system. During the second step, the main phase of uplift proceeds as hydrothermal fluid flow transports the pressure disturbance outwards as a pore pressure wave and the effective source depth shallows. Once it reaches minimum depths the third step begins in which elevated pressures are dissipated by lateral outflow resulting in subsidence. Alternatively, Orsi et al. (1999a) proposed that the principal effect of an inflation of magma is seismogenic fracturing that establishes hydraulic connectivity between the magmatic and hydrothermal systems (Fig. 3.1b). High temperature and pressure magmatic fluids then flow into the hydrothermal system under the pressure gradient established by the pressurisation of the magma body, increasing pore pressures and generating uplift. Once the pressurisation of magma ends, the input of magmatic fluids into the hydrothermal system ceases. Subsidence then proceeds at a rate determined by the permeability contrast between the hydrothermal system and the surrounding caldera. In contrast to the previous three-step sequence described, pressurisation of the hydrothermal system occurs throughout the uplift period.

(ii) The magma degassing-trigger model considers the 1982-1984 uplift to result from the transport of a batch of exsolved magmatic fluids to 2.5-3.5 km depth from a deeper magma reservoir (Fig. 3.1c; De Natale et al., 2006; Battaglia et al., 2006). The fluids accumulate in a horizontal lens throughout the initial aseismic uplift in 1982 then, once a critical overpressure is exceeded, seismogenic brittle failure of the overlying crust occurs forming fractures. The accumulated fluids flow into the hydrothermal system, increasing pore pressures and generating uplift. Subsidence begins once the supply of magmatic fluids is exhausted and results from lateral outflow through fractures. In a variation of this mechanism suggested by Troiano et al. (2011) based on earlier work modelling fumarole gas chemistry at Solfatara by Todesco et al. (2003) and Chiodini et al. (2003), a permanent connection between the hydrothermal system and a reservoir of magmatic fluids at 3-4 km depth is assumed (Fig. 3.1d). The ground level is then modulated by pore pressures in the hydrothermal system that vary depending on the flux of magmatic fluids from the underlying reservoir that is in turn controlled by the episodic delivery of fluids from greater depths.

(iii) The regional tectonic earthquake-trigger model also attributes pressurisation of the hydrothermal system during the 1982-1984 uplift to an influx of lithostatically pressured magmatic fluids, but in this case the transport of these fluids to shallower depths is dynamically triggered by the 1980 M 6.9 Irpinia earthquake (e.g. Barberi et al., 1984; Martini et al., 1984; Lupi et al., 2017). According to Lupi et al. (2017) the process is initiated by a high strain rate of c. 10-5 s-1 at the ductile crystallised margin of the primary magma reservoir imposed by the passage of body waves, which triggers brittle failure. Exsolved fluids then propagate through hydraulic fracture into the hydrothermal system where volumetric expansion on decompression results in a pore pressure increase and uplift (Fig. 3.1e). Hydrofracturing occurs aseismically until fractures propagate into the brittle shallow crust. As in the magma degassing models the duration of uplift is limited by the supply of magmatic fluids and subsidence begins once this is exhausted.

Implicit in each of these models is that without a further episode of pressurisation, subsidence will continue until overpressure in the hydrothermal system has completely dissipated through fluid flow into the surrounding caldera. An incomplete return to the starting ground level in 1982 is attributed to either; a residual overpressure in either a magma body or reservoir of magmatic fluids, a residual overpressure in the hydrothermal system, or a thermal expansion of the rock matrix.

Alternatively, Orsi et al. (1999a) suggested that a permanent component of uplift could result from slip along fractures during pressurisation of the magma body that initiates the uplift and subsidence sequence.

The magma pressurisation-trigger models (i) were developed assuming a model of the crust that included a large-magma reservoir at 4 km depth left over from the Neapolitan Yellow Tuff caldera forming eruption. Such a magma body is now known from seismic tomography not to exist (e.g. Zollo et al., 2008). Nevertheless, the same sequences can be applied assuming a different magmatic pressure source, such as the primary magma reservoir located at c. 7 km depth, or a shallow intrusion (see Chapter 2). In the case of the magma degassing-trigger models (ii), magmatic fluid accumulation below the main hydrothermal circulation is compatible with the zone of supercritical gas and brine at 3-4 km depth inferred from seismic tomography (Zollo et al., 2008).

The regional tectonism-trigger model (iii) is more difficult to reconcile. In this case it is assumed that all uplifts at Campi Flegrei, irrespective of magnitude, result from the transport of a batch of magmatic fluids into the hydrothermal system initiated by an earthquake rather than endogenic processes. Lupi et al. (2017) justified this on the basis that 8 out of 12 uplifts since 1945 were preceded by a regional earthquake (i.e. originating less than 300 km from Campi Flegrei) with a magnitude greater than Mw 4.5 1-3 years before uplift began. The occurrence of uplift, the variable lag times between an earthquake and a change in the ground level, and magnitudes of deformation were concluded to be a function of the availability of magmatic fluids, and therefore the rate of crystallisation of the parent magma, in addition to the degree of permeability generation at the margin of the magma. Given that tectonic activity occurs more frequently than major uplifts, the implication of the model is that they represent periods of enhanced crystallisation of the magma. An immediate response of the shallow hydrothermal system to regional seismicity has been documented at Solfatara since at least the early 20th Century. Observations include; mud fountaining to heights up to c. 30 m following the M 6.6 Irpinia earthquake in 1930, increased gas flux at fumaroles after the 1980 M 6.9 Irpinia earthquake, and changes in the pattern of thermal energy release (related to changes in degassing) after the M 5.8 L’Aquila event in 2009 (Signore et al., 1935; Martini et al., 1986; Lupi et al. and references therein). These variations in activity result from the opening and closing of fractures in the upper part of the system. An immediate response of the deep system is not apparent in monitoring data.

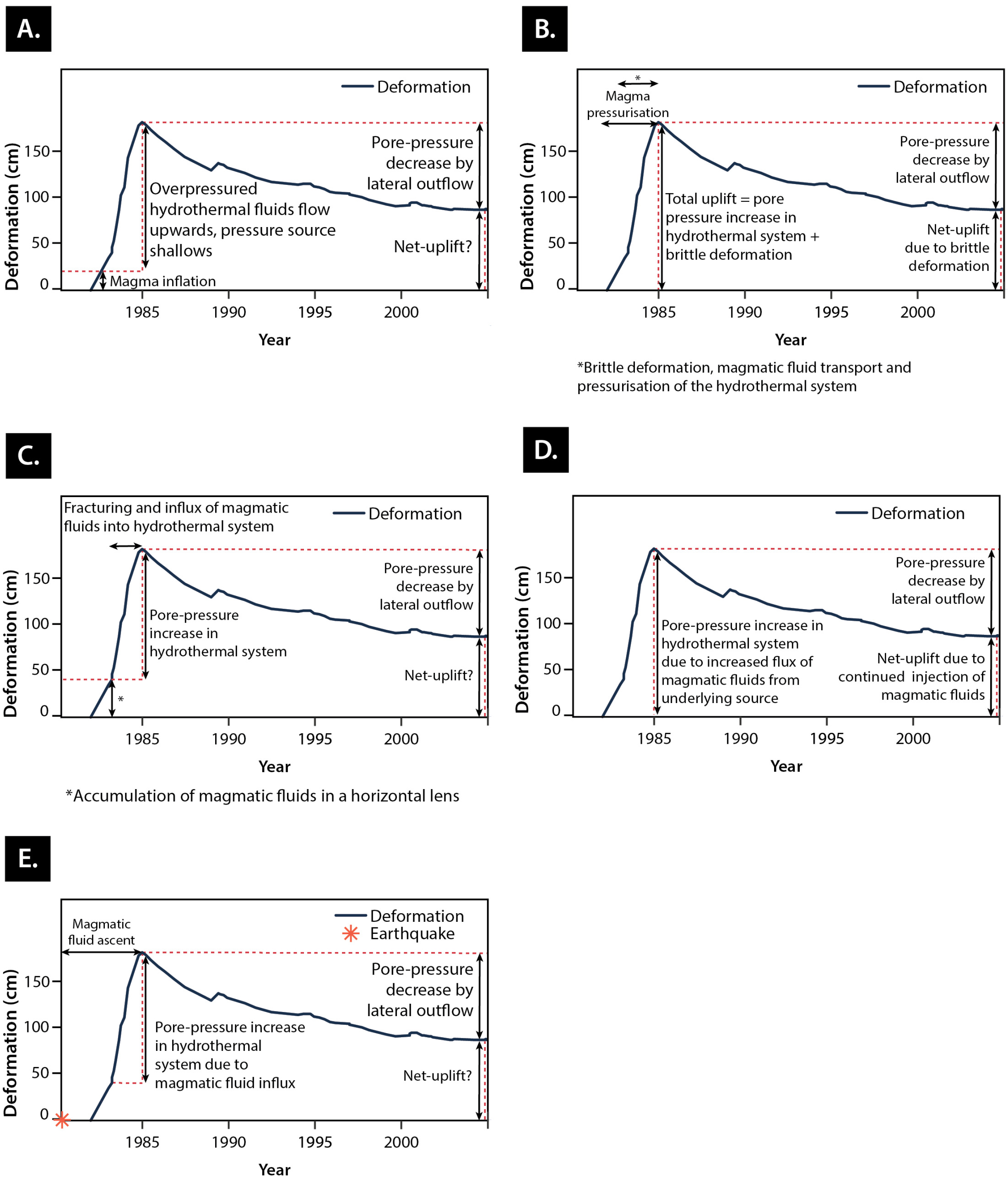


Figure 3.1: The processes of deformation during the 1982-1984 uplift and following subsidence according to models that assume subsidence represents a depressurisation of the hydrothermal system. A. and B. describe the magma pressurisation-trigger models. A. is the deformation sequence according to Gaeta et al. (1998); Castagnolo et al. (2001); De Natale et al. (2001); Troise et al. (2001); Troise et al. (2001). B. is the sequence described by Orsi et al. (1999a). C. and D. are the magma degassing-trigger models. C. is the sequence according to De Natale et al. (2006) and Battaglia et al. (2006) and D. is that described by Troiano et al. (2011). E. is the earthquake-trigger model (Barberi et al., 1984; Martini et al., 1984; Lupi et al., 2017).

Deformation in response to a change in fluid dynamics in the hydrothermal system is comprised of a thermo-elastic component caused by variations in heat transport by circulating fluids, and a poro-elastic response of the crust to the re-distribution of pore pressure (Hurwitz et al., 2007; Hutnak et al., 2009; Fournier and Chardot, 2012). Bonafede (1991) was the first to test the feasibility of a hydrothermal control for the 1982-1984 uplift and found that an increase in temperature of 100 □C in a 1 km thick porous volume saturated in H2O could reproduce the observed magnitude of the deformation. However, it was assumed that deformation principally resulted from a thermal expansion of the host rock (c. 85%) rather than a poro-elastic inflation. It has since been found that over the timescales of caldera unrest the pressure and temperature fields are effectively decoupled, so that the immediate elastic response of the crust is due to changes in pore pressure. The latter only becomes significant over centennial or longer timescales (Hurwitz et al., 2007; Fournier and Chardot, 2012; Coco et al., 2016). This is consistent with observations from a range of geological contexts and applications where mechanical deformation of rocks has been observed to occur immediately following changes in pore pressure (Todesco et al., 2008).

Two approaches have been taken to numerical modelling of the 1982-1984 uplift and following subsidence as a response to changes in pore pressure (Table 3.1). Orsi et al. (1999a) quantified the deformation of an elastic, porous volume saturated in H_2O to a pressurisation and depressurisation of pore fluids, without considering thermal effects. The more common approach, however, has been to simulate the changes in the pressure and temperature profile of a saturated porous volume following an increase in pressure at its base by solution of governing equations describing the conservation of momentum, mass and thermal energy (Ingebritsen et al., 2010). The results are then used to predict the response of the crust using mechanical models. The earliest simulations of hydrothermal fluid flow (e.g. Gaeta et al., 1998) simulated pore fluids as a single-phase, non-compressible, pure H_2O. Subsequent models that can account for the condensation and vaporisation of H_2O have utilised the TOUGH2 code (Pruess et al., 2012), which simulates flow using more realistic H_2O + CO_2 mixtures (e.g. Todesco et al., 2004; Chiodini et al., 2003; Rinaldi et al., 2010; Troiano et al., 2011). TOUGH2 has further advantages in that it can account for the reduction in fluid velocity caused by the preferential occupation of large void spaces by gas where liquid and gas phases co-exist (Elder, 1981; Todesco, 2008), mass loss at the surface and the presence of CO_2, which can enhance ground deformation relative to a pure H_2O system (Hutnak et al., 2009).

Table 3.1: Summary of quantitative models of pore-pressure induced deformation for Campi Flegrei.

Reference(s)	Approach	Modelled Deformation	Fluid	Pressure Source	Permeability (m^2)	Porosity	Key Results
Gaeta et al. (1998)	Used 1-D fluid dynamical modelling to predict changes in the volumetric flux and associated temperature and pore pressure distribution across a cylindrical saturated porous volume to an applied pressure at its base. Uplift was estimated by summing stacked Mogi (1958) point sources to represent a migrating pressure front.	1982-1984 Uplift	Single-Phase H_2O	10 MPa	10^{-12}-10^{-11}	0.4	- Propagation of pressurised fluids through a porous volume can induce and amplify uplift. - Travel time for fluids to reach surface from 3 km depth ~ 1-10 years. - Maximum amplitude of 1.3-2.6 m achieved for a simulation time of 1 year depending on the radius of the uplifting volume (1.5-3 km).
Orsi et al. (1999b)	Modelled uplift and subsidence by increasing pore pressure to a maximum in a cylindrical saturated porous volume representing the hydrothermal system (2 km radius, 4 km depth) and allowing pore pressure to decrease through lateral diffusion of fluids into an enclosing ring of lower permeability (4 km across, 4 km depth) representing the surrounding caldera.	1982-1984 Uplift + Subsidence	Single-Phase H_2O	100 MPa	10^{-13}-10^{-10}	0.2	- Maximum uplift attained for maximum pore pressures. - Once pressurisation of the volume ends subsidence occurs by lateral diffusion at a rate determined by the permeability contrast between the inner and outer model domains. - A pore pressure increase and decrease can reproduce the observed evolution of the 1982-1984 uplift.
Todesco et al. (2003; 2004)	Used the TOUGH2 code to model flow, pressure and temperature in the Solfatara plume by injecting hot (350 ?C) fluids into the base of a saturated porous volume 1.5 km deep with a radius of 2.5 km. Pore pressure changes were simulated by varying the injection rate. The mechanical response of the host rock was calculated from the TOUGH2 results using the FLAC3D code for rock mechanics.	Uplift + Subsidence (non-specific)	Two-phase $H_2O + CO_2$ mixture	Injection rates 5x and 10 x from 1500 t d^{-1} of CO_2 and 3000 t d^{-1} of H_2O then reduced to initial conditions.	10^{-14}	0.2	- Confirmed maximum uplift attained at the end of pressurisation period and that rate of subsidence is slower than uplift. - Elastic response of rock to pore pressure changes is immediate. - Deforming volume becomes larger overtime as pore pressure propagates. - Reproduced evolution of 1982-1984 uplift. - Change in fluid density leads to subsidence beyond initial ground level. - Deformation centred on plume. - Achieved displacements ~0.1 m.

Table 3.1 continued.

Reference(s)	Approach	Modelled Deformation	Fluid	Pressure Source	Permeability (m^2)	Porosity	Key Results
Rinaldi et al. (2010)	Used TOUGH2 to simulate the effects of changes in hydrothermal flow to variations in the flux of fluids injected from a point source into the base of saturated porous volume 1.5 km deep with a radius of 10 km to simulate the shallow portion of the caldera. A model of linear poro-elasticity was then used to simulate the mechanical response of the crust to these changes.	1982-1984 Uplift + Subsidence	Two-phase $H_2O + CO_2$ mixture	Increase in injection rate from 1000 t d^{-1} CO_2 + 2400 t d^{-1} H_2O to 6000 t d^{-1} CO_2 + 6100 t d^{-1} H_2O	10^{-14}	0.2	- Displacements controlled by injection rate, permeability and elastic parameters of the crust. - Confirmed immediate elastic response. - Confirmed reduction in fluid density over time leads to subsidence beyond initial ground level. - Achieved displacements ~ 0.1 m.
Troiano et al. (2011)	Modelled the hydrothermal system as a saturated porous cylinder 3 km in depth with a radius of 1.5 km, enclosed by a lower permeability ring with a diameter of 10 km. Uplift and subsidence was simulated by varying the rate of injection of fluids from a point source in the centre of the model. Pressure and temperature conditions were simulated with TOUGH2 and a mechanical model was used to calculate the elastic response of the crust.	1982-1984 Uplift + Subsidence	Two-phase H_2O, two-phase $H_2O + CO_2$ mixture	Increased injection rate to 8000 t d^{-1}	10^{-16}-10^{-15}	0.2	- Reproduced the magnitudes of the 1982-1984 uplift and the evolution of the following subsidence by varying the rate of fluid injection into the base of the modelling domain. - Ground level can be modulated by pore pressure.

Numerical models have confirmed that the evolution of the 1982-1984 uplift can be reproduced by an increase in pore pressure at the base of the hydrothermal system that propagates outwards from the source during uplift and is dissipated by lateral outflow during the following subsidence. They have also established the following constraints: the elastic response of the crust is immediate; maximum uplift is attained at the end of the period of pressurisation at the base of the system; and subsidence proceeds at a rate that is controlled by the decreasing pore pressure gradient over time, as well as the permeability of the volume through which lateral outflow occurs.

TOUGH2 models simulating the trends in the CO_2/H_2O ratio of fumarole gases and gravity measurements at Solfatara during uplift and subsidence have found that variations in these parameters are compatible with a pressurisation and depressurisation of the Solfatara plume, and thus the main hydrothermal reservoir (e.g. Todesco et al., 2003; Chiodini et al., 2003; Todesco and Berrino, 2005). The use of numerical models to quantify the degree to which the hydrothermal system contributes to deformation, however, is critically limited by the lack of constraints as to the magnitude of the pressure disturbance at the base of the system, and the permeability profile of the crust below Pozzuoli. These are process limiting parameters that control the pore pressure distribution in the hydrothermal system, and even minor variations in these values have been shown to significantly change monitored parameters (e.g. the characteristics of deformation) at the surface (Todesco et al., 2010; Rinaldi et al., 2010). Attempts have been made to constrain these values from observed geochemical and geophysical signals. For example, Rinaldi et al. (2010) simulated unrest by injecting an H_2O + CO_2 mixture into the base of a saturated porous volume (Table 3.1) at a rate constrained by the CO_2/H_2O ratio of fumarole gases during uplift. The computed maximum uplift was 0.1 m, an order of magnitude less than observed during the 1982-1984 unrest. This was in part attributed to the fact that only the upper 1.5 km of the hydrothermal system was modelled to keep within the P-T limits imposed by the TOUGH 2 code (350 °C, 100 MPa). Troiano et al. (2010) simulated the mechanical response of the entire 3 km depth of the hydrothermal system to pore pressure changes modulated by variations in the rate of fluid injection at the base of the model. In this case the permeability of the modelling domain and the rate of fluid injection were constrained by the observed deformation profile. The model was therefore entirely self-consistent.

3.1.2 Accumulation and Release of Magmatic Fluids from below a Hydrological Barrier

An alternative model to those discussed in the previous section proposes that uplift occurs when magmatic fluids accumulate at lithostatic pore pressures below a hydrological barrier in the shallow crust, and that subsidence represents the escape of these fluids into the overlying hydrothermal circulation (e.g. De Vivo et al., 2006; Bodnar et al., 2007). In contrast to the models discussed in the previous section, the hydrothermal system does not contribute to ground movements. Similar scenarios have been applied to ground deformation at other calderas such as Yellowstone, United States (Fournier, 1989), Long Valley Caldera, United States (Hildreth, 2017) and Nisyros, Greece (Gottsman et al., 2007). For Campi Flegrei, the model assumes that the magmatic-hydrothermal system is analogous to porphyry-epithermal ore forming environments, where batches of lithostatically pressured magmatic fluids are episodically transported into an overlying hydrothermal system and undergo decompression boiling (e.g. Burnham, 1979, Fournier, 1999). Such a process has been inferred to operate at Campi Flegrei from fluid-crystal inclusions found in cores from the boreholes drilled by AGIP at Mofete and San Vito. In particular, the presence of liquid-vapour-crystal inclusions associated with hypersaline fluids are attributed to the phase separation of magmatic gas, and the coexistence of liquid and vapour-dominated inclusions to boiling (De Vivo et al., 1989; De Vivo and Lima, 2006).

The model considers the 1982-1984 uplift and the following subsidence in three distinct stages (Figs. 3.2 and 3.3; De Vivo et al., 2006; Bodnar et al., 2007; Lima et al., 2009). In the first stage (Fig. 3.2a), CO_2-H_2O rich fluids exsolve from a parent magma located at 6 km depth and accumulate at lithostatic pore pressures below a ductile crystallised margin. The surrounding crust stretches in response to the increasing fluid pressures and, once a critical strain is exceeded tensile failure occurs. During the second stage (Fig. 3.2b), the exsolved fluids propagate through hydraulic fracture to c. 2 km depth where their ascent is limited by an impermeable claystone-siltstone layer. Uplift then proceeds during a second phase of fluid accumulation below this hydrological barrier. The final stage of the model (Fig. 3.2c) begins once the accumulation of overpressured fluids is sufficient to trigger fracturing of this barrier, increasing permeability and allowing the lithostatically pressured magmatic fluids to escape into the overlying hydrothermal system where they undergo decompression boiling. Subsidence results from the loss of pore pressure beneath the claystone-siltstone layer.

Figure 3.2: The three stages of the model (from Lima et al., 2009). A. In the initial phase the lithostatic and hydrostatic pressure regions of the crust are isolated. B. Fracturing of the impermeable margin allows accumulated magmatic fluids to propagate into the overlying crust where their ascent is limited by an impermeable sedimentary layer and uplift begins. C. Fluids accumulate below this shallow hydrological barrier until overpressures result in a second phase of fracturing. Fluids dissipate in the overlying hydrothermal system, resulting in subsidence. Associated SiO_2 precipitation during boiling seals fractures, returning the system to initial conditions.

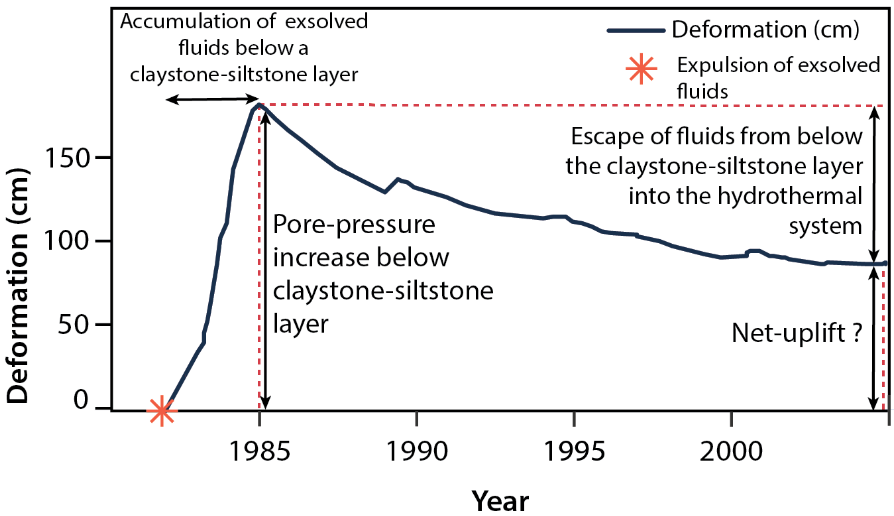


Figure 3.3: The evolution of the 1982-1984 unrest according to De Vivo and Lima (2006); Bodnar et al. (2007) and Lima et al. (2009).

Bodnar et al. (2007) demonstrated that the crystallisation of 0.83 km^3 of hydrous basaltic magma (3 wt% H_2O) saturated in CO_2 (358 ppm) at 6 km depth could generate a mechanical energy release of 7 x 10^{15} J. This is sufficient for the 1982-1984 uplift, although Aiuppa et al. (2013) have argued that the volatile content of the assumed parent magma would be an inadequate CO_2 source for the observed rates of degassing at Solfatara-Pisciarelli. In order to generate uplift and subsidence, the model requires that magmatic fluid transport in the crust occurs at velocities in the order of km yr^{-1}, and that these fluids enter the hydrothermal system. Such velocities are compatible with the results of numerical simulations of porphyry ore genesis (e.g. Weis et al., 2012), whilst an input of magmatic gas-rich fluids into the hydrothermal system during the 1982-1984 uplift is consistent with the interpretation of an enrichment in CO_2 at Solfatara fumaroles at this time (Chiodini et al., 2003). However, the consistency of the model with observations breaks down when the distribution of seismicity during the uplift is considered. The two-stages of fracturing imply a shallowing of hypocentres during the unrest, but this was not observed. The presence of a laterally extensive claystone-siltstone layer of sufficient strength to act as a hydrological barrier below which lithostatically pressured fluids can accumulate is also debatable. Such a lithological layer has only been located from a borehole at Agnano, where pore pressures are hydrostatic and it is not known if it has the required mechanical strength (Piochi et al., 2014). Alternatively, this constraint could be fulfilled by the low permeability hard rock layer that isolates the hydrothermal circulation from the underlying magmatic system at c. 2.7 km depth, below which magmatic fluids are thought to accumulate (see Chapter 2). A key assumption of the model is that the hydrothermal system does not act as a pressure source at any point during the uplift-subsidence sequence. It therefore does not consider a potential contribution to uplift from the transport of magmatic fluids into the hydrothermal system, unlike the models described in the previous section, or if a component of subsidence results from an increase in outflow triggered by the permeability change during the second stage of fracturing.

A natural consequence of decompression boiling in the hydrothermal system is the precipitation of SiO_2, which seals fractures (e.g. Fournier, 1989). In the context of the model such sealing is considered to end connectivity between the lithostatically and hydrostatically pressured regions of the crust. The system is thus returned to initial conditions so that a new cycle can begin. As a result of its cyclic nature, the model has been proposed to explain all known uplifts since the end of Epoch III volcanism, irrespective of the scales of the ground movements (Fig. 3.4; De Vivo and Lima, 2006; Bodnar et al., 2007; Lima et al., 2009). The magnitude and duration of individual episodes is modulated by the availability of magmatic fluids, the degree of fracturing during a cycle, and it is assumed that the relation between deformation and other monitoring parameters should always be the same. It is therefore difficult to account for the contrasting deformation behaviours following the 1969-1972 and 1982-1984 uplifts and why enrichment in CO_2 in fumarole gases began before the onset of the current uplift rather than lagging the deformation signal as it did for the previous episode.

Figure 3.4: Schematic representation of vertical ground displacement at Campi Flegrei and the inferred timings of magmatic fluid transport and fracturing (from Lima et al., 2009). The H_2O saturated carapace is the region above the parent magma where volatiles accumulate below a ductile crystallised margin in the first part of the three-stage model described.

3.2 Models for Uplift Since 2004

An independent category of model constrains interpretations of uplift since 2004 entirely from compositional trends in gases from Solfatara fumaroles since the 1982-1984 unrest. Two competing models have emerged: the latent heat of condensation model (Chiodini et al., 2015a; 2016) and the CO_2 induced drying model (Moretti et al., 2017 and 2018). Both consider the post-1984 subsidence to represent a decompression of the hydrothermal system following an influx of magmatic fluids during an intrusion of magma at c. 3-4 km depth in 1982-1984 (as in the models described in section 3.1.1), and the uplift since 2004 to reflect heating in the hydrothermal system. The assumption of heating is based on temperature estimates of the deep hydrothermal reservoir (c. 2-3 km depth) from CO_2-CH_4 isotopic exchange and of the upper Solfatara hydrothermal plume (<0.5 km depth) from hydrothermal gas equilibria. The estimated temperature increases since 2000 are c. 50 °C in the deep reservoir and 15-30 °C in the Solfatara plume (Chiodini et al., 2011, 2015a, 2016 and Moretti et al., 2017). The models contrast in their interpretations of where in the crust degassing bodies of magma are currently located, as well as the origin of N_2 in fumarole gases, the control on the CO_2/H_2O trend after 2000 and the reactivity of CH_4 in the Solfatara plume.

3.2.1 Latent Heat of Condensation

According to the latent heat of condensation model (Fig. 3.5), magma from the primary magma reservoir at c. 7 km depth was intruded at 3 to 4 km depth in 1982-1984. Since then, it has continued to decompress in-situ to Present. A key assumption is that the magmatic component of Solfatara fumaroles from the 1982-1984 unrest onwards has been controlled by degassing of this shallow magma. The model interprets an enrichment of CO_2 and N_2 in Solfatara fumarole gases during the uplift and their subsequent decline together with He during subsidence, as reflecting the changes in the composition of gas separating from the magma as it became depleted in the least soluble species (N_2 being the least soluble). This was based on geochemical modelling by Caliro et al. (2014) who found that the temporal trends in the ratios of these gases, in particular N_2/CO_2 and N_2/He, could be reproduced by modelling the continuous decompression of a trachybasaltic magma by 120 MPa. Peaks in these gases superimposed on the main trend are considered to represent episodic injections of magmatic fluids into the hydrothermal system. The onset of a continuous increase in CO_2 and the inverse trend in H_2O observed from 2000 was then taken to indicate that the magma had decompressed to the Critical Degassing Pressure (CDP), as defined by Chiodini et al. (2015a). At the CDP, the model predicts an order of magnitude increase in the volume of degassed fluids as H_2O starts to be released and in greater amounts through time (Chiodini et al., 2015a). An expected consequence of the enhanced rate of degassing is an increase in the frequency of injections of progressively H_2O enriched magmatic fluids into the hydrothermal system after 2000. Such an increase in frequency was inferred from peaks in the redox indicator CO_2/CH_4 in Solfatara fumarole gases, which were interpreted as periods of increase influx in CO_2-rich, oxidising magmatic gases, supressing the formation of CH_4 (Chiodini et al., 2015a; 2016). According to simulations of the upper 1.5 km of the Solfatara plume using TOUGH2, this would raise pore pressures, promoting the condensation of progressively larger amounts of H_2O. The removal of H_2O vapour in combination with the increased frequency of magmatic fluid injections, increases the relative concentration of CO_2 in the gas phase, which is compatible with the observed CO_2/H_2O trend in fumarole gases (Chiodini et al., 2015a). Uplift is then attributed to the combined effect of the pore pressure increase in the main hydrothermal reservoir caused by the increased frequency of magmatic fluid injections, and to a thermal expansion of the crust caused by the latent heat of the condensation.

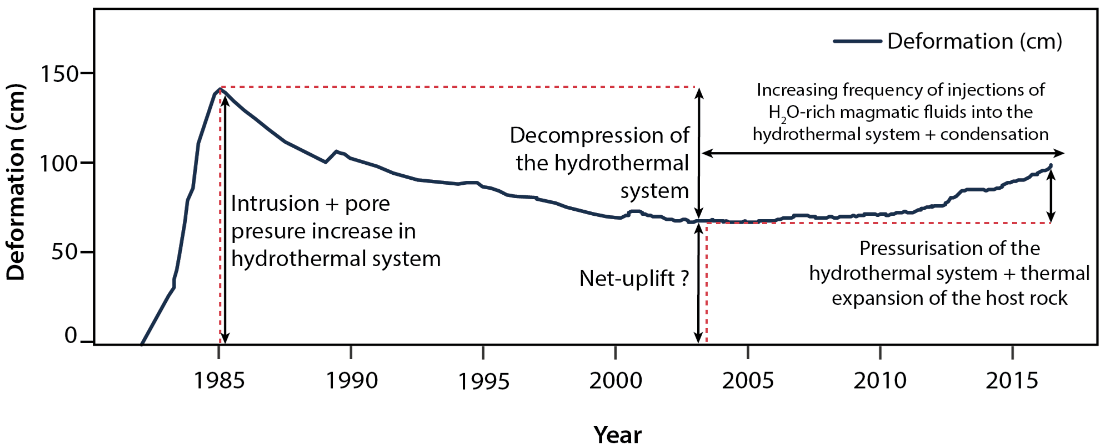


Figure 3.5: Evolution of the deformation profile since 1982 according to Chiodini et al. (2015a; 2016).

According to the model, the pore pressure increase in the main hydrothermal reservoir would propagate through the Solfatara plume. Equilibrium pressures in the plume estimated using gas equilibria in the CO_2-H_2O-H_2-CO system indicate a pressure increase of c. 4 MPa since 2000 (Chiodini et al., 2015; 2016). Such values are comparable to the maximum following the 1982-1984 uplift calculated using the same method. The differing rates and magnitudes of the two uplift periods are not addressed but the implication is that the 1982-1984 uplift represents deformation due to the magma intrusion and an additional hydrothermal component of an amount comparable to uplift since 2004. The model assumes that the post-1984 subsidence represents a decompression of the hydrothermal system, which requires that the hydrothermal component of the previous uplift represents a pore pressure increase. In the case of the current uplift, however, the model assumes that the mechanical response of the crust to the increase in pore pressure is secondary to the thermo-elastic component of deformation caused by heating. Chiodini et al. (2015) estimated that the associated energy released by condensation between 2003-2014 was $\sim 6.2 \times 10^{12}$ kJ, which is sufficient for a 5 °C temperature increase in a volume of 0.0625 km^3. Assuming a rock density of 2000 kg m^{-3} and a volumetric expansion coefficient of 30×10^{-6} m °C^{-1}, this could produce a thermo-elastic inflation of the host rock of 0.94×10^5 m^3. This is the same order of magnitude as the observed uplift, but it is not clear that such a process could generate this deformation over the required timescales, as numerical modelling suggests that the poro-elastic component would dominate (Hurwitz et al., 2007; Fournier and Chardot, 2012; Coco et al., 2016).

The principal advantage of the latent heat model is its ability to reproduce the long-term trends in Solfatara fumarole gases since the onset of monitoring at Bocca Grande in 1983, by which it is constrained. However, it is limited by the assumption that all CO_2 degassed from Solfatara since 1982 was sourced from a degassing magma at 3-4 km depth. Assuming an average rate of degassing through Solfatara of 1500 t d^{-1} (e.g. Chiodini et al., 2001), this would require the solidification of c. 6 to 32 km^3 of magma (Table 3.2). An intrusion of such a volume is incompatible with estimates of the source volume from geodetic inversions (Woo and Kilburn, 2010) and the absence of evidence for the presence of magma of more than 1 km^3 at depths shallower than c. 7 km depth (Zollo et al., 2008). This suggests that more than one degassing source has contributed to the observed geochemical trends since 1982-1984. Possible sources include the zone of supercritical magmatic fluids at 3-4 km depth and the main magma reservoir at c. 7 km.

Table 3.2: Estimates of magma volumes for the observed CO_2 flux at Solfatara since 1982 assuming a magma density of 2500 kg m^3. Parent magma compositions from Chiodini et al. (2016).

Initial Magma Composition		Annual Flux		CO_2 Flux Since 1982	
wt % H_2O	wt % CO_2	Mass of Magma (kg)	Volume of Magma (km3)	Mass of Magma (kg)	Volume of Magma (km^3)
5.39	0.025	2.2E+12	8.8E-01	8.1E+13	3.2E+01
4.89	0.046	1.2E+12	4.8E-01	4.4E+13	1.8E+01
3.89	0.079	6.9E+11	2.8E-01	2.6E+13	1.0E+01
3.38	0.093	5.9E+11	2.4E-01	2.2E+13	8.7E+00
2.86	0.105	5.2E+11	2.1E-01	1.9E+13	7.7E+00
2.02	0.123	4.5E+11	1.8E-01	1.6E+13	6.6E+00

3.2.2 CO_2-Induced Drying

The CO_2-induced drying model considers the 1982-1984 uplift to comprise of an inelastic component related to the intrusion of a thin sill (<10 m thick) and an additional inflation of the hydrothermal system that was recovered during subsidence (Fig. 3.6). As for the latent heat model, it interprets the trend in CO_2 during this period to be controlled by the degassing of a magma intrusion at 3-4 km depth, but the models differ in their interpretations of the cause of the progressive increase in CO_2 after 2000 and uplift since 2004. In the CO_2-induced drying model, the magma intrusion that initiates the 1982-1984 uplift is assumed to have essentially solidified by 2000 (c. 75-80%). This is based on numerical modelling of a degassing trachyte sill at constant temperature. The loss of this degassing source in 2000 requires that from this period the magmatic component of Solfatara fumarole gases originates from a second source elsewhere in the magmatic system (Moretti et al., 2013). Moretti et al. (2017) suggest that until 2000 the sill acted as a hydrological barrier to background degassing from the primary magma reservoir at c. 7 km depth. At this time the model expects hydraulic connectivity between the hydrothermal and magmatic systems to have been re-established by fracturing of the sill during cooling and contraction. The continuous increase in CO_2 and inverse trend in H_2O observed thereafter, is attributed to the progressive enrichment of the hydrothermal system in hot (1000 °C) CO_2-rich fluids from the magma at c. 7 km depth. Uplift results from a consequent thermal pressurisation of the main hydrothermal reservoir of 15 MPa. Unlike the model discussed in section 3.1.2.

the accumulation and release of magmatic fluids below an impermeable layer does not contribute to deformation.

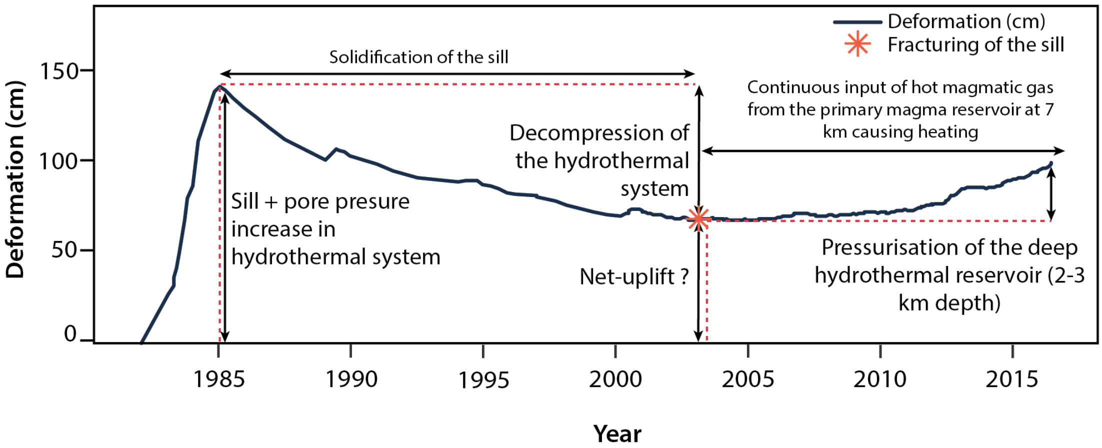


Figure 3.6: Evolution of the deformation profile since 1982 according to Moretti et al. (2017; 2018).

To explain the change in the behaviour in CO_2 and H_2O concentrations from oscillatory to continuous trends in fumarole gases from 2000, the model suggests that decompression of the hydrothermal system during post-1984 subsidence caused H_2O to boil, resulting in the progressive loss of the liquid phase from the main hydrothermal reservoir over this period. The continuous increase in CO_2 and decrease in H_2O in fumarole gases from 2000 is then explained in terms of the quasi-isenthalpic ascent of a progressively CO_2 enriched single-phase fluid from the hydrothermal reservoir through the Solfatara plume. The model suggests that a consequence of the relative increase in the gas phase since the 1982-1984 unrest would be a progressive decompression of the Solfatara plume towards atmospheric values, suggesting that since the onset of uplift in 2004 pore pressures in the shallow hydrothermal system have decreased, whilst those in the main hydrothermal reservoir, which feeds the plume, would have increased (Moretti et al., 2018). This conflicts with the conclusion of Chiodini et al. (2015) that an increase in pore pressure in the main reservoir propagates through the Solfatara plume and that pressure in the plume has continuously increased since 2000 to values comparable to those at the end of the 1982-1984 uplift (c. 4 MPa).

The conflicting interpretations as to whether pore pressures are increasing or decreasing in the shallow hydrothermal system results from differing assumptions about whether the Solfatara plume is mono- or bi-phasic, and where CH_4 equilibrates when estimating pressure conditions from chemical equilibria. According to Chiodini et al. (2015a), CH_4 equilibrates in the main hydrothermal reservoir (c. 2-3 km) at 360-436 °C, rather than in the plume, and H_2O condensation thermally buffers the deep hydrothermal system along the line of liquid-vapour coexistence. These assumptions are based on the δ ^{13}C fractionation temperatures between CO_2-CH_4 (e.g. Caliro et al., 2007; 2014) and the results of TOUGH2 simulations of the Solfatara plume respectively. Under these assumptions CH_4 equilibria cannot be used to estimate P-T conditions in the upper plume and the fugacity of water (f_{H_2O}) in equilibria calculations must be fixed by the co-existence of vapour and liquid according to the $f - T$ relation of Giggenbach (1980). Moretti et al. (2017) alternatively argue that CH_4 cannot be assumed to be unreactive in the plume because the isotopic equilibrium of δ ^{13}C between CO_2-CH_4 occurs c. 400 times more slowly than chemical equilibrium, whilst a key outcome of the model is that the decompression of the plume would prevent the condensation of H_2O. As a result, they consider estimation of equilibrium pressures in the Solfatara plume using the CO_2-CO-H_2O-H_2-CH_4 system to be appropriate and that the assumption of condensation may be relaxed. By doing so, a continual decompression trend since 1984 can be obtained (Moretti et al., 2017). In contrast to the prevailing assumption that N_2 in fumarole gases has a magmatic origin (e.g. Caliro et al., 2007), Moretti et al. (2017) suggested that an enrichment in this gas from Solfatara fumaroles during the 1982-1984 uplift could be interpreted as resulting from the exsolution of crustal N_2-rich fluids or flashing of NH_3 in the hydrothermal reservoir, rather than an influx of N_2-rich magmatic fluids. The decline during the following subsidence then represents its progressive removal through outflow. This allowed for the estimation of pressure in the plume using equilibrium constants for the N_2-NH_3 conversion that agreed with those for CO_2-CO-H_2O-H_2-CH_4. However, a non-magmatic source for N_2 is difficult to reconcile with the strong co-variance between variations in the concentration of this gas and CO_2 (assumed to have a magmatic origin) prior to 2000, which implies a common source, and the subduction-zone fluid like signature of $δ^{15}N$ signature.

The increase in CO_2 transport to the surface during uplift since 2004 assumed by the CO_2-drying model is compatible with the observed increase in degassing at Solfatara since the early 2000s. However, the assumption of a single-phase plume of hydrothermal fluids below Solfatara is neither consistent with the increase in activity at Pisciarelli, which has been associated with the arrival of increased volumes of condensates at the surface (Chiodini et al., 2011), or the results of electrical resistivity imaging of the plume (e.g. Byrdina et al., 2014). Both the increase in degassing and activity at Pisciarelli can be explained in terms of enhanced fluids transport of two-phase fluids to the surface. This favours a pressurisation of the plume and the propagation of pore pressures from the main hydrothermal reservoir to the surface through Solfatara, as in the latent heat of condensation model.

3.3 Discussion

Common to all models for the post-1984 subsidence, as described in sections 3.1.1 and 3.1.2, is the assumption that this ground movement is a consequence of an increased coupling between the magmatic-hydrothermal systems during the preceding uplift. Those that require the pressurisation of the uplift source to continue throughout uplift and the transport of magmatic fluids into the hydrothermal system can, qualitatively at least, account for the geodetic signal and the observed enrichment in CO_2 in Solfatara fumarole gases during the 1982-1984 unrest. There is, however, no agreement as to the extent to which the hydrothermal system acted as a deformation source during the 1982-1984 uplift and following subsidence.

A necessary condition of the decompression of the hydrothermal system models is that a component of the preceding uplift resulted from the pressurisation of the hydrothermal reservoir, which is most commonly attributed to an input of magmatic fluids. Implicit is that the minimum contribution of the hydrothermal system to the uplift is equivalent to the following subsidence (0.9 m). In contrast, the model described in section 3.1.2 expects the transport of magmatic fluids into the hydrothermal system during uplift but considers its involvement in deformation to be negligible. Numerical models have shown that pore pressure changes in the hydrothermal system are a viable mechanism for producing the observed geodetic signal during the uplift-subsidence sequence, but quantification of the maximum contribution of the hydrothermal system to uplift is limited by the lack of constraints for the pressure gradient during the unrest and the permeability of the crust. Simulations of the upper 1.5 km of the hydrothermal system are only able to produce c. 5% of the observed uplift in 1982-1984 (e.g. Rinaldi et al., 2010), whilst those that can produce the observed displacement require a pressure source that is either at or exceeds the tensile strength of the crust (<10 MPa, e.g. Gaeta et al.,1998; Orsi et al., 1999a), or constrains the model by the observed deformation (e.g. Troiano et al., 2011). Critically, they assume permeability (which exerts a first order control on pore pressure) to have been static throughout uplift. This cannot be the case as Volcano-Tectonic (VT) seismicity in the shallow crust during the 1982-1984 uplift is an indicator of fracturing and faulting, and thus changes in flow paths. According to general models of pore pressure induced deformation, fracturing or self-sealing processes can potentially lead to large changes in the observable parameters at the surface, so that unrest may reflect permeability changes in the shallow crust, rather than the state of the magmatic system (e.g. Todesco et al., 2010). Thus, an inflation of the hydrothermal system is not necessarily a pre-requisite for subsidence, which favours models such as that in section 3.1.2 where subsidence results from mechanical changes in the crust.

The characteristics of successive uplifts between 1950-1984 are similar, in particular the rates and magnitudes of the ground movements, which implies a common source mechanism, but the observed characteristics of deformation following the 1969-1972 and 1982-1984 uplifts differ. Rather than a prolonged aseismic subsidence, the 1969-1972 unrest was followed by an immediate minor subsidence, and swarms of VT seismicity persisted at declining rates until the onset of the subsequent uplift. Decompression of the hydrothermal system models assume that aseismic subsidence is an inevitable consequence of uplift and that the maximum magnitude of subsidence is equivalent to the hydrothermal component of the uplift. Such models therefore cannot be applied to explain the evolution of the 1969-1972 unrest and are unable to account for why the subsidence signal was not observed following uplifts prior to 1982-1984.

In the alternate model type for subsidence (section 3.1.2), ground movements are regulated by the accumulation of magmatic fluids below a hydrological barrier and the episodic generation of permeability by fracturing across this lithological layer during uplift. Volcano-Tectonic (VT) seismicity can be considered as a proxy for the amount of brittle deformation during uplift (Kilburn, 2012). As such, higher rates of VT seismicity recorded during the 1982-1984 uplift relative to that in 1969-1972 (Corrado et al., 1977; Orsi et al., 1999b; Barberi et al., 1984; D'Auria et al., 2011 and 2015) may be interpreted as reflecting a greater degree of fracturing. In the context of the model this fracturing is occurring in the crust overlying the accumulation of magmatic fluids, so that the loss of overpressured fluid from below the hydrological barrier into the hydrothermal system, and thus subsidence, would be expected to be greater following the 1982-1984 episode. The degree of fracturing, however, is dependent on the exceedance of a critical overpressure below the hydrological barrier. The model assumes that each uplift is an independent event, so that the comparable magnitudes of the 1969-1972 (1.76 m) and 1982-1984 (1.79 m) uplifts would suggest that similar overpressures had accumulated in each case. As such, the model cannot account for the greater seismic energy release during the 1982-1984 uplift.

In all models described in sections 3.1 and 3.2, subsidence after 1984 represents a return towards lithostatic equilibrium. The uplift since 2004 would therefore require an increased coupling between the magmatic and hydrothermal systems in order to move the system away from these conditions. The mechanisms proposed for uplift in 1982-1984 in section 3.1 cannot be applied as they expect variations in the concentration in CO_2 in Solfatara fumarole gases to lag the geodetic signal. The increase in CO_2 in fumarole gases from 2000 before the onset of uplift in 2004 therefore requires the operation of a different causative process. Both the latent heat of condensation and CO_2-induced drying models for the current uplift consider this ground movement to be a consequence of the evolution of a magma body intruded in 1982-1984. Each assumes that the intrusion of magma resulted in a pressurisation of the hydrothermal system caused by an input of magmatic fluids during the unrest and that the subsidence represents a loss of pore pressure in the hydrothermal system. They are also both constrained by the same temporal trends in fumarole gas compositions. Differing assumptions as to the volume of magma intruded during the 1982-1984 unrest and the rate at which it is solidifying has led to contrasting interpretations as to current conditions in the magmatic-hydrothermal system with radically opposing implications for hazard. According to the latent heat model, a degassing magma is currently located at 3-4 km depth and, as uplift progresses, the probability of a pathway opening from the magma to the surface opening increases (Chiodini et al., 2015; 2016). It also suggests that the upper Solfatara plume is becoming pressurised, increasing the likelihood of a phreatic eruption or other explosive hydrothermal phenomena. Conversely, the CO_2-induced drying model concludes that no magma is currently present at depths shallower than the primary magma reservoir (c. 7 km), and that pore pressures in the Solfatara plume have decreased over time. The implication is that the eruptive hazard, magmatic or non-magmatic, is presently at its lowest since the 1982-1984 unrest. In neither case can the models account for why a slow ground oscillation was not observed following earlier uplift episodes in 1950-1952 and 1969-1972.

3.4 Summary

Models of the evolution of the 1982-1984 uplift have shown that a loss of pore pressure in the upper 3 km of the crust is an effective mechanism for subsidence after 1984 and can reproduce the required magnitude of the deformation for relevant rates and timescales. Most assume a minimum 0.9 m of the 1982-1984 uplift was related to a pressurisation of the hydrothermal system. However, the alternate model suggests that this is not a pre-requisite and that subsidence can be triggered by mechanical changes in the shallow crust that result in a redistribution of pore pressure, without a previous inflation of the hydrothermal system.

Compositional changes in fumarole gases since 1983 have been shown to be compatible with a deformation sequence initiated by a magma intrusion in 1982-1984, where the following subsidence and uplift since 2004 represents a depressurisation and re-pressurisation of the hydrothermal system. Such a model can account for residual uplift once subsidence ceased, but the inferred processes controlling the re-pressurisation of the hydrothermal system are critically dependent on the evolution of the inferred magma intrusion. Rates of CO_2 degassing at the surface since 1982 and the change in the behaviour of geochemical trends in gases from 2000 favour a change in the degassing source at this time. In neither case do the discussed models consider the mechanical effect on the crust of a magma intrusion in 1982-1984.

Common to all models discussed is the assumption that subsidence represents a return to lithostatic equilibrium and that an increased coupling between the magmatic-hydrothermal systems is necessary for both rapid and slow uplifts to proceed. None can account for the emergence of the slow ground oscillation after 1984 and why such behaviour was not observed following the earlier rapid uplift episodes in 1950-1952 and 1969-1972. Given that current interpretations of ongoing uplift since 2004 have opposing implications for hazard and therefore unrest management, it is suggested that a new approach to interpreting the potential causative processes of slow ground movements since 1984 is required.

Chapter 4

Methodology

Chapters 2 and 3 provided the geological context of the Campi Flegrei magmatic-hydrothermal system and identified the key limitations of existing conceptual models in accounting for the change in the characteristics of deformation after 1984. In order to develop a robust model of unrest that is compatible with the full sequence of ground movements, it is necessary to establish the known behaviour of the caldera before and after the onset of unrest in 1950. In particular; the characteristics of deformation through time, the temporal relationship between deformation and other monitoring parameters, and changes in hydrothermal surface activity.

This chapter first describes the history of instrumental monitoring of the caldera in order to provide context as to the availability of monitoring data over time. It then goes on to describe the collation of long-term trends in monitoring parameters and observations of hydrothermal activity from existing catalogues, observatory reports, the scientific literature, and media sources. These were then combined with observations and interpretations in the existing literature and reviewed chronologically. Finally, the method applied to analyse the distribution of seismicity relative to the location of hydrothermal reservoirs in 1982-1984 is described. The review and seismicity analysis form the basis of the conceptual model developed in Chapter 6, which reinterprets the knowledge of the behaviour of the caldera to account for that change in characteristics of deformation after 1984. Further methodological details related to the collection and analysis of data used in determining the perceptions of scientists are presented together with the results in Chapter 7, which is a self-contained chapter.

4.1 Volcano Monitoring at Campi Flegrei

4.1.1 Geophysical Parameters

Direct measurements of vertical ground movements for volcano monitoring were first conducted by the Italian Military Geographic Institute (IGM), who carried out levelling surveys in 1905, 1919, 1922, 1953 and 1968 (Dvorak and Mastrolorenzo, 1991; Del Gaudio et al., 2010). Continuous monitoring of the ground level began in March 1970 with the installation of four permanent tide gauges (Fig. 4.1a and b) following the recognition of uplift by fishermen, who reported changes in the heights of structures along the coast and a shallowing of the harbour (Yokoyama, 1971; Scherillo, 1977; Orsi et al., 1999b). From this time until the end of 1972, repeated levelling surveys were conducted every 1-3 months by the IGM and the Ministero di Lavori Publici (Ministry of Public Works). Levelling took place along three principal lines that ran from Pozzuoli to Baia, to Quarto and to Nisida (Versino, 1972; Corrado et al., 1977; Orsi et al., 1999b). Horizontal deformation was also measured from a trilateration network of 27 stations along survey lines that ran from the Italian Air Force Academy (1.7 km east of Serapeo), to end locations in Baia, Ricettone and Nisida (Dequal, 1972; De Michelis et al., 1975; Dvorak and Berrino, 1991).

Following the end of the 1969-1972 uplift, the frequency of levelling surveys was decreased to 1-2 times per year and in 1975 responsibility for monitoring ground movements was transferred to the Vesuvius Observatory, who expanded the levelling network. The 1982-1984 uplift was recorded by repeated levelling surveys across 124 benchmarks every three months (Fig. 4.1c), whilst the trilateration network (Fig. 3d) was extended in 1983 to include a local network at Solfatara (Berrino et al., 1984; Dvorak and Berrino, 1991). After uplift ended, the frequency of surveys was again reduced and over time the network was further expanded to c. 350 benchmarks along 135 km of levelling lines, arranged in 14 loops (Fig. 4.1d, Orsi et al., 1999b; Del Gaudio et al., 2009). Since 2000 ground movements have also been recorded by continuous GPS (cGPS) stations that are part of the regional Neapolitan Volcano Continuous GPS network (NeVocGPS, De Martino et al. 2014). Aseismic movements below the resolution of cGPS are recorded by six Sacks-Everton dilatometers and two arrays of long-baseline underground water tube tiltmeters installed in 2004-2005 and 2008 (Di Lieto et al., 2017). Offshore ground movements are monitored using four instrumented buoys in the Gulf of Pozzuoli that have been progressively installed since 2008 as part of the INGVs MEDUSA project (MEDUSA, 2019).

Continuous seismic surveillance was established in March 1970 (Fig. 4.2a), with a network of three permanent three-component and ten portable radio-controlled stations. A further three stations became operational at the end of 1972 (Corrado et al. 1977). During the 1982-1984 uplift seismicity was monitored by a maximum of twenty-two analogue single component seismometers operated by the Vesuvius Observatory and AGIP. These were supplemented by an additional eighteen digital three-component stations between September to November 1983 from the Institut de

Physique du Globe de Paris, and by ten digital three-component stations between December 1983 to June 1984 from the University of Wisconsin (Fig. 4.2b, Aster et al., 1992; Orsi et al., 1999b; D'Auria et al., 2015). Currently twenty-one permanent stations are in operation, which are part of the regional Osservatorio Vesuviano Seismic Network (Fig. 3.2c, Castellano et al., 2002; Chiodini et al., 2017). This includes the four MEDUSA buoys in the Gulf of Pozzuoli, which are also equipped to record seismicity.

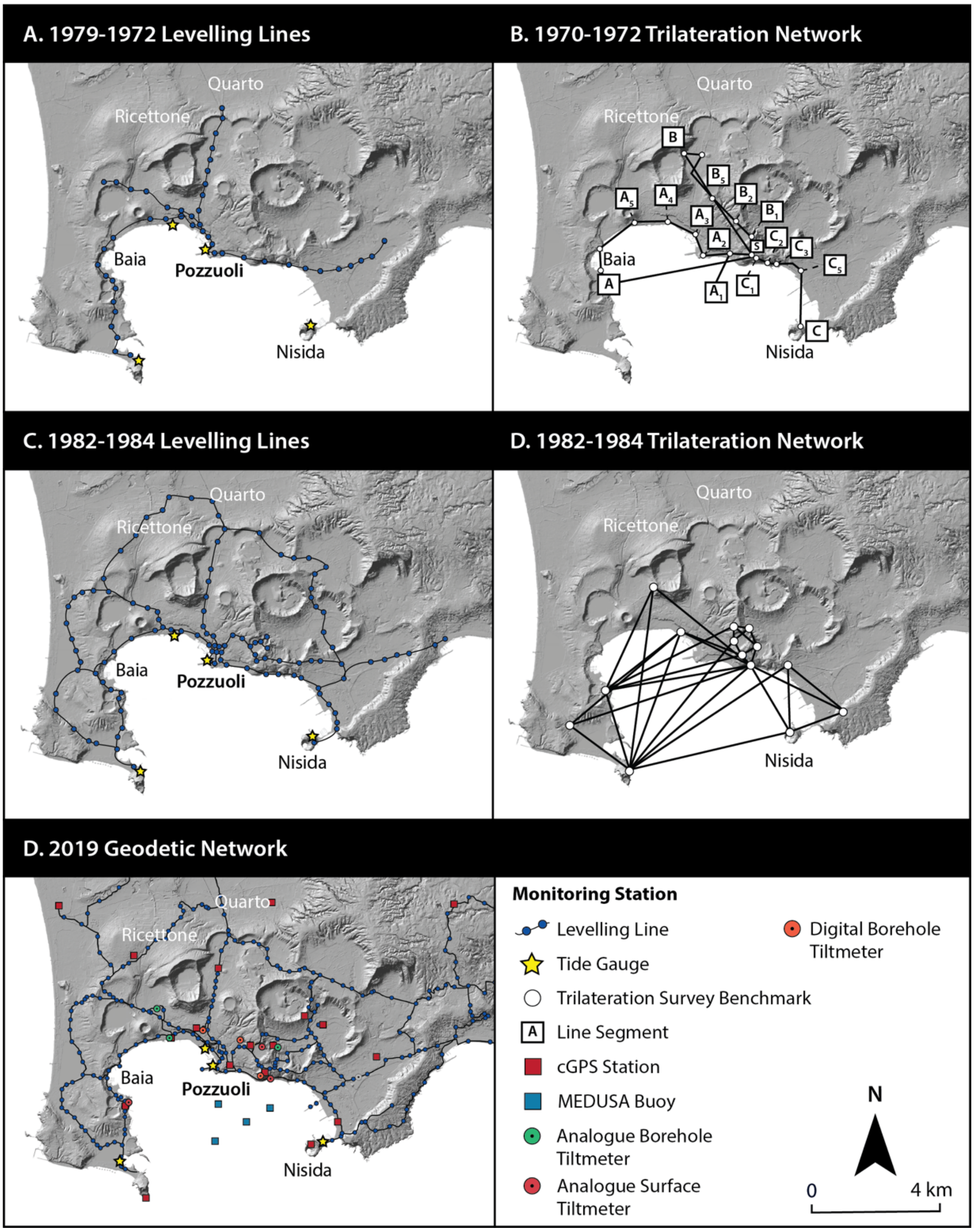


Figure 4.1: Ground deformation monitoring network during uplift periods. A and B show the levelling network and trilateration survey lines respectively between 1970-1972. The line segments refer to the locations used to measure change in horizontal distance between the Italian Air Force Academy (S) and points at Baia (A), Ricettone (B), and Nisida (C). C and D show the configuration of the vertical and horizontal deformation monitoring stations in 1982-1984, and D shows the present-day network. Station locations are digitised from Dequal et al. (1972); Corrado et al. (1977); Berrino et al. (1984); Orsi et al. (1999b) and (INGV-OV, 2019).

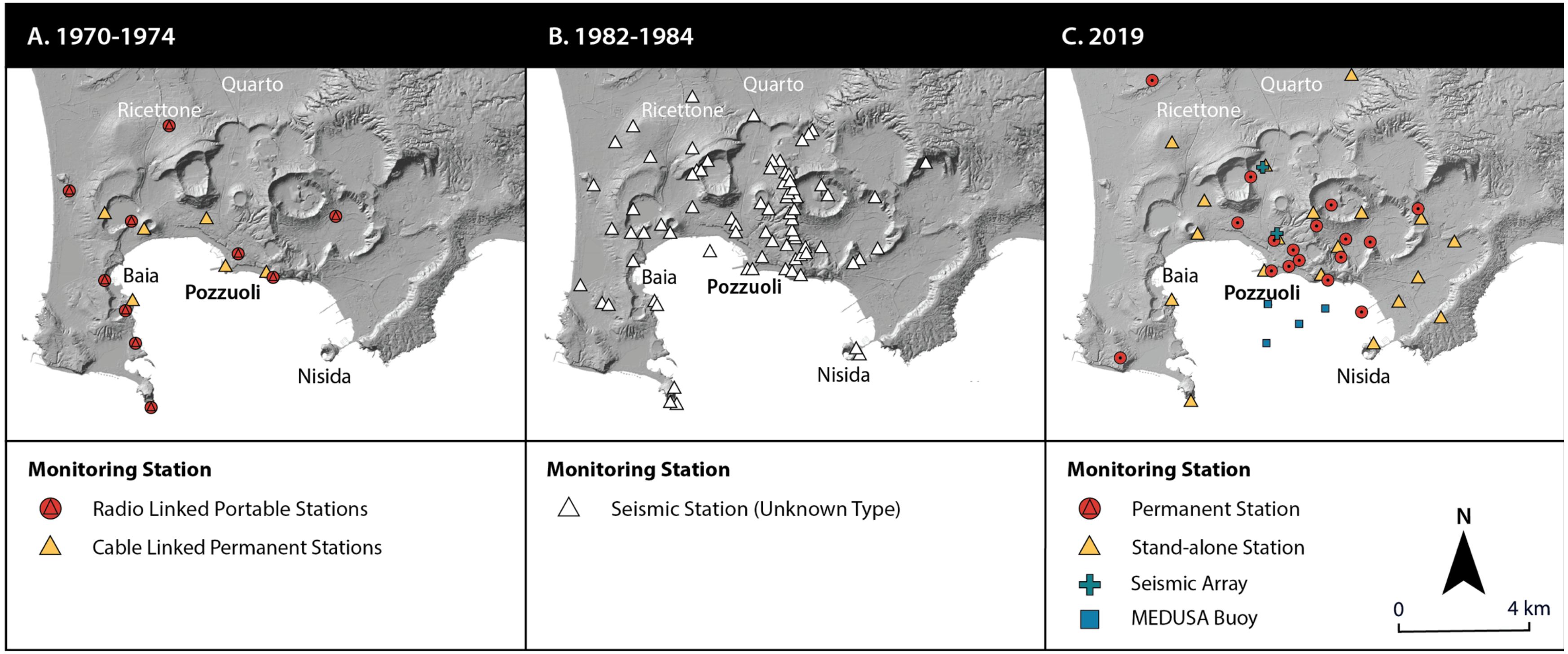


Figure 4.2: The seismic monitoring network during uplift periods. A. The seismic monitoring network from 1970-1974. B. The network during the 1982-1984 uplift. In this case the map shows the points at which seismometers were located during the uplift but not all stations operated continuously throughout the unrest (D'Auria et al., 2015). C. The current configuration of the seismic monitoring network. Station locations digitised from Corrado et al. 1977; D'Auria et al., 2015 and (INGV-OV, 2019)

4.1.2 Hydrothermal Features

Observations of hydrothermal features across the caldera since at least the Roman Times are available in the literature (e.g. Giacomelli and Scandone, 2012 and references therein), whilst measurements of the physical and chemical characteristics of fumarole gases and thermal waters are known to exist from the 19th century onwards. However, consistent sampling and reporting of the characteristics of hydrothermal features for volcano monitoring has largely been restricted to the hottest and most vigorously degassing features located at Solfatara-Pisciarelli (Chapter 2, section 2.2.2), in particular the Bocca Grande, Bocca Nuova and Pisciarelli fumaroles (Fig. 4.3). The temperature and composition of the gases from Bocca Grande have been analysed episodically since at least 1923 (Dall'Aglio et al., 1972; Martini et al., 1986 and references therein) but regular sampling did not begin until March 1970. From then until the end of the 1969-1972 uplift episode, gas analyses were conducted by the Consiglio Nazionale delle Ricerche (CNR), as part of a state sponsored investigation into the unrest, and a group from the Institut de Physique du Globe de Paris (Dall'Aglio et al., 1972; Global Volcanism Program, 2013). In 1978, continuous monitoring began at another vent located in the Solfatara crater called the Fumarola Circolare. Monitoring was then expanded to include Bocca Grande in 1983 following the onset of seismicity during the 1982-1984 unrest. Gases from both vents were regularly sampled until March 1984 when the Fumarola Circolare collapsed and became extinct (Cioni et al., 1984; Tedesco et al. 1989). Monitoring of Bocca Grande continues to present, whilst gases from Bocca Nuova and Pisciarelli have been regularly sampled and reported since 1995 and 1999 respectively. Compositional analysis of the Solfatara fumaroles was temporarily suspended in September 2017 due to closure of the crater pending legal investigations into the deaths of three people who fell into a ground collapse in the crater floor and asphyxiated due to the high concentrations of CO2. The Public Prosecutor's Office authorised a resumption of monitoring activities by the Vesuvius Observatory in March 2018 (INGV-OV, 2019). Submarine fumaroles and vents external to the Solfatara-Pisciarelli area have not been included in regular monitoring programs.

In addition to the analysis of fumarole gases, systematic measurements of diffuse soil CO_2 degassing have been conducted since 1998 over an array of 30-71 points across the Solfatara-Pisciarelli DDS using the accumulation chamber method, and from three permanent CO_2 flux stations (Fig. 4.4, Granieri et al., 2010; Cardellini et al., 2017). Further surveillance includes continuous infra-red thermal imaging from five stations that make up the Permanent Thermal Monitoring Network (TIRNet, Vilardo et al., 2015). Spot temperature measurements of fumaroles elsewhere in the caldera (Agnano, Monte Nuovo and Mofete) using mobile thermal imaging cameras and a rigid thermo-couple, have been reported semi-regularly since January 2018 (INGV-OV, 2019).

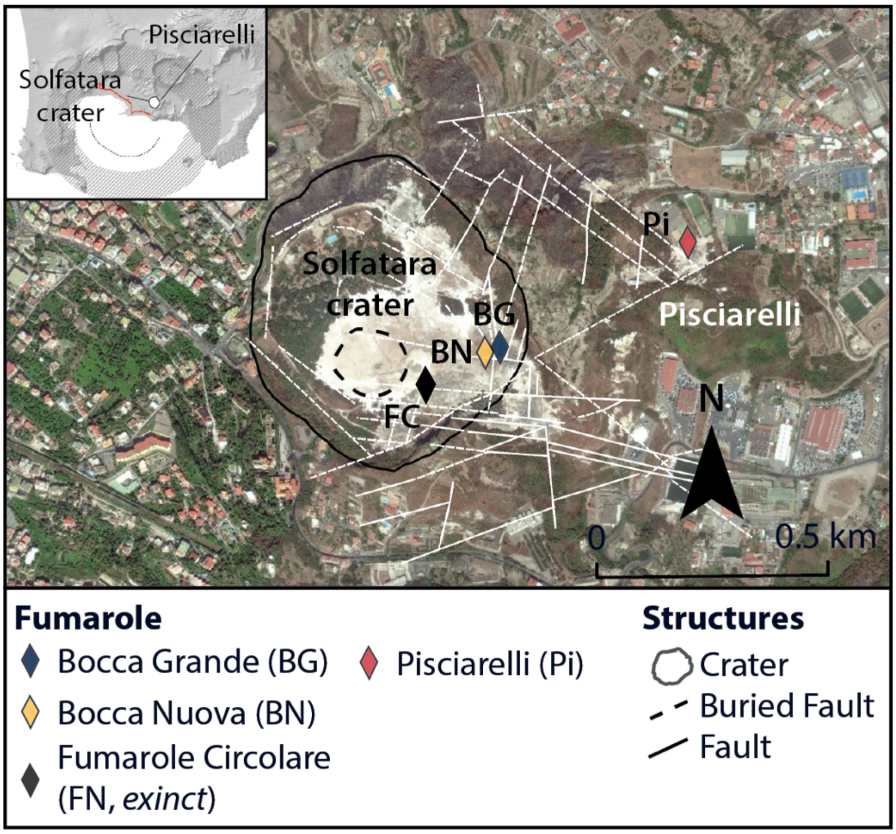
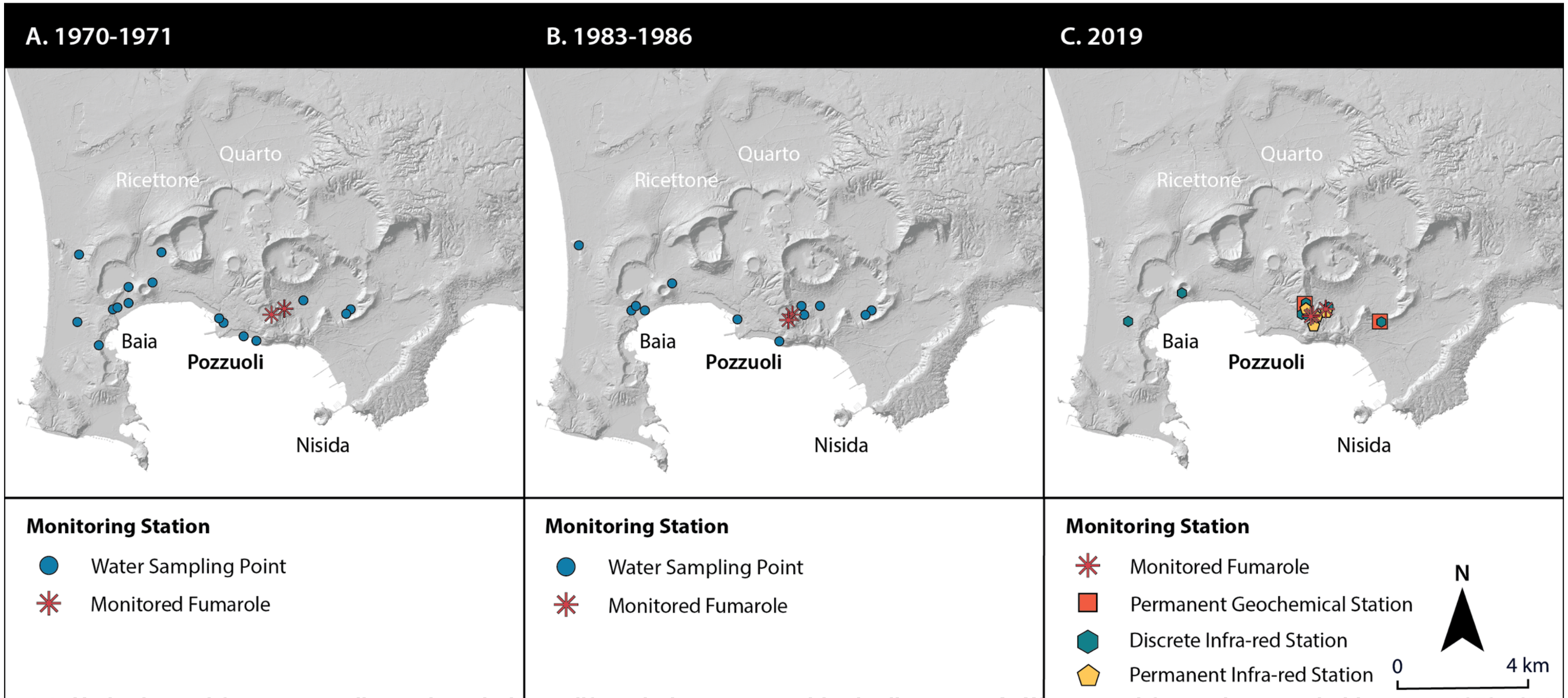


Figure 4.3: Location of monitored fumaroles. The Fumarola Circolare (FC) was monitored between 1978-1983. Bocca Grande (BG) has been continuously monitored since 1983, whilst the Bocca Nuova (BN) and Pisciarelli (Pi) vents have been monitored since 1995 and 1999 respectively. Base map: Esri, HERE, Garmin, © OpenStreetMap contributors, and the GIS user community, Source: Esri, DigitalGlobe, GeoEye, Earthstar Geographic, cNESAirbus DS, USDA, USGS, AeroGRID, IGN, and GIS user community.

The sampling of thermal waters from springs, dug out wells and crater lakes across the caldera (Fig. 4.4), and reporting of their characteristics has been intermittent since initial surveys of minor and trace element contents of waters by Dall'Aglio et al. (1972) that were conducted between 1970-1972. From March 1983 to January 1986 waters were sampled for major and minor element concentrations, and isotopic composition analysis (Ghiara et al., 1988; Ghiara and Stanzione, 1988; Celico et al., 1992b). The literature suggests that following the end of the 1982-1984 uplift, systematic sampling for major, minor and trace element analysis was continued at a varying frequency until at least 1990 by the ENEA (Italian National Agency for New Technologies, Energy and Sustainable Economic Development), the University of Naples, and the University of Rome (Martini et al., 1991; Celico et al., 1992b). Since then the reporting of thermal water analyses from a volcanological perspective has been limited (e.g. Valentino and Stanzione, 2003 and 2004).



~~Figure 4.4: Hydrothermal feature sampling points during uplift periods as reported in the literature. A. Waters and fumaroles sampled between 1970-1971 as~~ reported in Dall'Aglio et al. (1972). B. Water and fumarole sampling locations sampled between 1983-1986 as reported by Ghiara and Stanzione (1988). C. Current sites of monitoring of hydrothermal features. Locations digitised from Dall'Aglio et al. (1972), Ghiara and Stazione (1988) and the Vesuvius Observatory (INGV-OV, 2019) respectively.

4.2 Collating Knowledge of the Behaviour of the Caldera

Knowledge of the behaviour of Campi Flegrei has principally been derived from a review of the scientific literature, which targeted references to observations and interpretations of deformation, seismicity and characteristics of hydrothermal features prior to the onset of unrest in 1950, during and after rapid uplifts between 1950-1984, and after 1984. This was supplemented by information from Vesuvius Observatory reports sourced online and from the Vesuvius Observatory library during a visit in September 2018, as well as qualitative information from the media. A quantitative catalogue of geophysical and geochemical data was also collated from these sources. This was done so that, where possible, temporal trends in the literature could be extended and to examine the relationship between different parameters over different phases of ground movement. A chronology of visual observations of surface activity was also established. The following details the types and sources of data included.

4.2.1 Deformation Data

Sourced deformation data includes; measurements of the height of the ground level at Benchmark 25A/cGPS station RITE (the geodetic station closest to the centre of the deforming area), horizontal deformation measurements, and levelling data from the E-W (from Pozzuoli to Baia, and Pozzuoli to Nisida) and N-S levelling lines (Pozzuoli to Quarto). A sequence of the height of the ground level at Benchmark 25A/cGPS station RITE has been constructed for the period 1905-February 2019. Values for dates between 1905-April 2000 are taken from Del Gaudio et al. (2010), who reconstructed the ground level at this location from 1905-2009 by combining direct measurements with estimates from measurements of sea level, and indirect observations of sea level relative to markers on the Serapeo columns (Roman ruins in Pozzuoli). From April 2000 to February 2019, the sequence is comprised of cGPS measurements collected from the sources given in Table 4.1. Levelling data for the period Jan 1982-Jun 1984 was received in a MS Excel file from C. Kilburn (University College London), whilst that for other periods between 1905 to 2008 was digitised from the references given in Table 4.1. Horizontal deformation data is limited to intervals during the two major periods of uplift in 1969-1972 and 1982-1984, as it was not possible to obtain measurements for other periods. Additionally, a data table was constructed that contains published deformation source depths for intervals between 1970 to 2013 that are estimated from inversions of either deformation or gravity data. The references from which this information was collected are given in Table 4.2.

Table 4.1: Summary of collated deformation data and sources.

Data	Time Coverage	Source	Source Type
Vertical displacement at Benchmark 25a	Mar 1905-Jul 2009	Supplementary file in Del Gaudio et al. (2010)	Research paper
Vertical displacement at RITE cGPS station	May 2000-Jul 2016	Supplementary file in Chiodini et al. (2017)	Research paper
Horizontal deformation	1970-1972	Bonasia et al. (1984)	Research paper
Horizontal deformation	1982-1983	Dvorak and Berrino (1991)	Research paper
Levelling data	1905-1919	Dvorak and Berrino (1991)	Research paper
Levelling data	1970-1995	Orsi et al. (1999b)	Research paper
Levelling data	Jan 1982-Jun 1984	C. Kilburn (UCL)	Pers. comm
Levelling data	1999-2000	Pingue et al. (2006)	Vesuvius Observatory Open File Report
Levelling data	2004-2006	Del Gaudio et al. (2007)	Vesuvius Observatory Open File Report
Levelling data	2006-2008	Del Gaudio et al. (2009)	Research paper

Table 4.2: References from which estimated deformation source depths were collected.

Period	Reference	Period	Reference
Jun 1970 - Aug 1971	Corrado et al. (1977)	Jun 1992 - Dec 2000	Lanari et al. (2004)
Mar 1970 - Jul 1972	Bonasia et al. (1984)	Feb 1993 - Apr 1999	Avallone and Zollo (1998)
Jun 1970 - Aug 1971	Bianchi et al. (1984)	Feb 1993 – Sep 1998	Lundgren et al. (2001)
Jun 1970 - Sep 1972	Bianchi et al. (1984, 1987)	1993 - 1999	Tiampo et al. (2017)
Jun 1970 - Sep 1972	Woo (2007)	1995 - 2000	Amoruso et al. (2014)
Sep 1980 - Sep 1983	Battaglia et al. (2006)	Jul 1999 - Dec 1999	Lanari et al. (2004)
Sep 1980 - Sep 1983	Amoruso et al. (2008)	Dec 1999 - Aug 2000	Lanari et al. (2004)
Sep 1980 - Sep 1983	Amoruso et al. (2014)	Mar - Aug 2000	Lanari et al. (2004)
Jan 1981 - Sep 1983	Dvorak and Berrino	2000 - 2001	Shirzaei and Walter (2009)
Jan 1982 - Jun 1984	Berrino et al. (1984 and 1987)	2001 - 2002	Shirzaei and Walter (2009)
June 1982 -June 1983	De Natale et al. (2001)	Nov 2002 - Nov 2006	Trasatti et al. (2008)
Jun 1982 - Jun 1983	Bianchi et al. (1987)	2004 - 2006	Amoruso et al. (2007)
1982 - 1983	Gottsman et al. (2006a)	Jan 2012 - Jul 2013	D'Auria et al. (2015)
Jan 1982 - Dec 1984	Orsi et al. (1999b)	Mar 2010	Amoruso et al. (2015)
Jan 1982 - Jun 1984	Beaudecel et al. (2004)	2011 - 2013	Trasatti et al. (2015)
Jan 1982 -Jun 1984	Trasatti et al. (2005)	2000 - 2005	Samsonov et al. (2014)
Jan 1982 - Jun 1984	Folch and Gottsman (2006)	2005 - 2007	Samsonov et al. (2014)
1982 - 1984	Woo and Kilburn (2010)	2007	Samsonov et al. (2014)
Jun 1990 – Jan 1995	Battaglia et al. (2006)	2007 - 2013	Tiampo et al. (2017)
Jul 1992 – Dec 1999	Lanari et al. (2004)		

4.2.2 Seismicity Data

A sequence of monthly counts of Volcano-Tectonic (VT) events between March 1970 to February 2019 was created through the combination of three data sources (Table 4.3). Monthly event counts from March 1970 to July 2000 are from a catalogue received as an MS Excel file from the Vesuvius Observatory. Those from August 2000 to July 2016 are from a dataset published in Chiodini et al. (2017), whilst those from August 2016 onwards were sourced from the Campi Flegrei monthly activity bulletins published on the Vesuvius Observatory website (ov.ingv.it). In addition to monthly rates of seismicity, earthquake locations for events that occurred in the Campi Flegrei area (in the monitoring districts of Bacoli, Pozzuoli, Quarto, and the Gulf of Pozzuoli) have been sourced for periods between 1970-2017. Earthquake epicentre locations for 211 events from 1970 to 1974 were digitised from maps of seismicity found in the literature

(Table 4.3). Earthquakes during this period or thought to have occurred between 1-5 km depth but hypocentre locations are not available for this period due to the configuration of the network at the time (Corrado et al., 1977; Orsi et al., 1999b). A catalogue of 3708 hypocentre locations for VT events between January 1983 and September 1984 was obtained from the Vesuvius Observatory. The catalogue is continuous except for a 6-week data gap in January-February 1984 and represents c. 25% of the total number of earthquakes that occurred during this period (c. 14000, Vesuvius Observatory). Events have magnitudes (M_S) between 0.2 and 4.0 and are located at depths less than 6.5 km. Locations for events between 2015-2017 (n= 492) were downloaded from the Vesuvius Observatory online seismological database (sismolab.ov.ingv.it). Included events have magnitudes (M_S) between 0.2 and 4.0 and were located between the near surface at c. 7 km depth. Data after 2017 was not available for download.

To check for the occurrence of volcanic earthquakes prior to the installation of the seismic network in 1970 the ASMI Italian Archive of Historical data (emidius.mi.ingv.it) was consulted and the literature was searched for references to earthquakes. No records of earthquakes occurring at Campi Flegrei between 1900 and 1970 were found. The only reference to felt seismicity within the caldera located was related to a tectonic earthquake in Irpinia, c. 100 km ENE of Campi Flegrei, on 23rd July 1930 (Signore, 1935).

Table 4.3: Summary of collated seismic data and sources. VT stands for Volcano-Tectonic.

Data	Time Coverage	Source	Source Type
Monthly VT event count	Mar 1970-Jan 2013	Vesuvius Observatory	Seismicity catalogue
Monthly VT event count	Aug 2000-Jul 2016	Supplementary file in Chiodini et al. (2017)	Research paper
Monthly VT event count	Aug 2016-Feb 2019	Vesuvius Observatory	Monthly activity bulletin
Earthquake epicentre locations	1970-1974	Digitised from Corrado et al. (1977)	Research paper
Earthquake epicentre locations	1970-1974	Digitised from Orsi et al. (1999b)	Research paper
Earthquake hypocentre locations	Jan 1983-Jul 2015	Vesuvius Observatory	Seismicity catalogue
Earthquake hypocentre locations	Aug 2015-Mar 2017	Vesuvius Observatory (sismolab.ov.ingv.it)	Online Database

4.2.3 Characteristics of Hydrothermal Features

Fumarole gas temperature and compositional data (H_2O, CO_2, H_2S, Ar, N_2, CH_4, H_2, He, $\delta^{13}C$, $\delta^{18}O$, δD) for the Bocca Grande, Bocca Nuova and Pisciarelli vents have been collected from data sets published by Chiodini et al. (2011; 2016) and Caliro et al. (2014) for the periods given in Table 4.4. Additional compositional data pre-dating 1983 were located in the literature but were not included in the catalogue as there were too few points to be able to extend the series and, in some cases, it was not possible to convert the reported values into units consistent with the rest of the dataset. Gas temperatures at Bocca Grande between 1925-1935 and in 1970, however, were retained for comparison with measurements from 1983 onwards (Table 4.4).

The results of the analysis of 1986 thermal water samples collected between 1970-1999 from 23 sites across the caldera, including springs, wells and crater lakes, were collated from the Italian National Geothermal Database (Geothopica) and results published in the scientific literature. Data includes the following; temperature, pH, Total Dissolved Solids (TDS) contents, the concentration of SiO_2, anions (SO_4, HCO_3, Cl), cations (Na, K, Mg, Ca), minor and trace elements (Sr, Li, Rb, Al, Fe, B, As, Sb, Hg, Tl, Pb, U, Br), NH_4, CO_2, and $\delta^{34}S$, $\delta^{18}O$ and δD values. Whilst the dataset has good spatial coverage across the caldera, it was found that the sampling and reporting frequency between sites was highly discontinuous and variable. As such, it was only possible to construct decadal time series for 8 locations (Fig. 4.5). The references used to construct these time series are given in Table 4.4. Data was available from 1970 to 1999 for two of these locations (Stufe di Nerone and Terme Puteolane) and from 1982-1999 elsewhere.

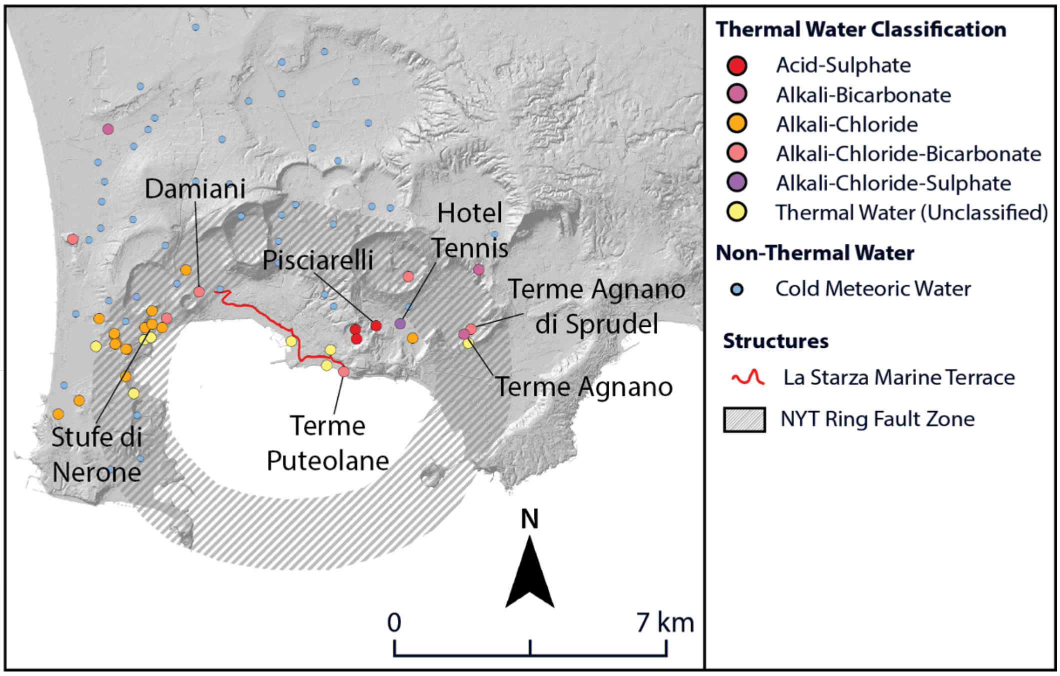


Figure 4.5: Locations of water sampling sites across Campi Flegrei published in the literature. Labelled locations are those for time series of the physical and chemical characteristics of the waters have been constructed. Two sampling sites are located at Stufe di Nerone; Stufe di Nerone (Well) and Stufe di Nerone (Spring). Cold Meteoric water locations from Aiuppa et al. (2006) and are not included within the dataset.

To establish a chronology of the distribution of hydrothermal features in the Solfatara crater and whether hydrothermal surface activity has changed since the onset of unrest in 1950, qualitative information consisting of images and descriptions of activity was collected. Historical paintings of the Solfatara crater from the 17th to 19th centuries were found via online image searches and maps of the location of features in the crater from the 20th and 21st century were collected from the scientific literature and Google Earth. Italian language scientific reports from the late 19th Century onwards located in online archives (e.g. archive.org, luxinfabula.it) and using general internet searches were checked for mentions of changes in hydrothermal surface activity, in addition to newspaper articles related to Campi Flegrei unrest from 1970 onwards that are part of an online collection maintained by Lux in Fabula, a cultural association based in Pozzuoli. Translation of Italian sources into English was done using the online translation tools Google Translate (translate.google.com) and DeepL (deepl.com). The English language scientific literature, Vesuvius Observatory bulletins and observatory reports (collected in the Vesuvius Observatory library and online on the Global Volcanism Program and Vesuvius Observatory websites) were also searched. Additionally, public posts on the social networking platform Twitter were episodically checked for mentions and images of hydrothermal activity in the caldera using the online TAGS tool for Google Sheets (tags.hawksey.info), which archives tweets tagged with a specified word or phrase preceded by the symbol '#' (e.g. #CampiFlegrei, #Pozzuoli, #Solfatara).

Table 4.4: Summary of collated gas and thermal water data sources.

Data	Time Coverage	Source	Source Type
Gas temperature and composition at Bocca Grande	Jun 1983-Jan 2016	Supplementary file in Chiodini et al. (2016)	Research paper
Gas temperature and composition at Bocca Nuova	Mar 1995-Dec 2015	Supplementary file in Chiodini et al. (2016)	Research paper
Gas temperature and composition at Pisciarelli	Mar 1999-Sep 2010	Supplementary file in Chiodini et al. (2011)	Research paper
$\delta^{13}C$ Composition at Bocca Grande, Bocca Nuova and Pisciarelli	Feb 2000 – Nov 2012	Supplementary file in Caliro et al. (2014)	Research paper
Gas temperature at Bocca Grande	1925-1935	Signore (1935)	Observatory Report
Gas temperature at Bocca Grande	Mar-Oct 1970	Dall'Aglio et al. (1972)	Research paper
Thermal water compositions	Oct 1971-Oct 1977	Italian National Geothermal Database (Geothopica)	Database
Thermal water compositions	1970-1971	Dall'Aglio et al. (1972)	Research paper
Thermal water compositions	Apr 1971-Jun 1975	Baldi et al. (1975)	Research Paper
Thermal water compositions	Apr 1971-Jun 1975	Cortecci et al. (1978)	Research Paper
Thermal water compositions	Mar 1983-Dec 1985	Ghiara et al. (1988)	Research Paper
Thermal water compositions	Mar 1970-Dec 1985	Ghiara and Stazione (1988)	Research Paper
Thermal water compositions	1970-1989	Martini et al. (1991)	Research Paper
Thermal water compositions	1978-1989	Celico et al. (1992b)	Research Paper
Thermal water temperature	Mar 1983-Dec 1992	Tedesco et al. (1996)	Research Paper
Thermal water compositions	Sep 1993-Mar 1994	Valentino et al. (1999)	Research Paper
Thermal water compositions	1985-Nov 1994	Valentino and Stanzione (2003)	Research Paper
Thermal water compositions	Feb 1990-Nov 1999	Valentino and Stanzione (2004)	Research Paper

4.3 Analysis of Unrest at Campi Flegrei

4.3.1 Chronological Review of Unrest

The collated data was combined with relevant existing observations and interpretations from the literature and reviewed chronologically in order to establish the relation between deformation and other observations during different periods of ground movements. First, the vertical deformation profile at Benchmark 25A/cGPS station RITE was divided into periods depending on the direction of ground movement. The rates and magnitudes were then compared between each. To determine whether the geometry of post-1984 ground movements was appreciably different from previous episodes of uplift and subsidence between 1905 to 1984, levelling data for intervals between 1905 to 2008 was compared graphically. To allow for the geometry of different magnitudes of deformation to be readily compared, the data for each interval was normalised to the maximum value for that period. This has been done previously (e.g. Orsi et al., 1999b) and this work increases the number of intervals included in the plot. The spatial relationship between the geometry of deformation for periods during uplift in 1969-1972, 1982-1984, 2004-Present (2019) and subsidence after 1984, and the assumed location of the hydrothermal basin below Pozzuoli (as defined by the Vp/Vs anomaly at 1, 2 and 3 km depth identified by Aster and Meyer, 1988, see Chapter 2) was compared using the 3D geospatial visualisation software, ArcScene by ESRI.

Once the characteristics of ground movements through time were established, the occurrence, rates of Volcano-Tectonic (VT) seismicity, and magnitudes of events in relation to ground movements was checked by comparison with the long-term trend in the ground deformation profile at Benchmark 25A/cGPS station and the literature. The distribution of seismicity during the 1969-1972, 1982-1984, and current uplifts, as well as during subsidence after 1984, was compared by mapping the locations of epicentres using ESRI's mapping software ArcMap. Finally, a time series of hypocentre depths was constructed. Time series of fumarole gas temperatures and gas compositions, together with ratios of magmatic and hydrothermal gases were also compared against the long-term trends of both the ground level at Benchmark 25A/cGPS station RITE, and monthly rates of VT seismicity. Additionally, the relative timings of changes in gas concentrations was established by the comparison of normalised trends. This allowed for gas concentrations of different magnitudes to be compared. The data was normalised using two methods, first by dividing concentrations by the maximum value, and second by using the standardised z-score.

In all cases, geophysical and geochemical trends were visualised and compared using the statistical programming language R.

In the case of thermal water compositions, waters at a sampling location were first categorised according to the dominant anion type present (SO_4, HCO_3, Cl) for mean concentrations, using a Piper diagram (Appendix A) and a ternary diagram (Giggenbach, 1980). This is standard in thermal water analysis (Nicholson, 1993). The concentrations of these anions present can indicate the relative contributions of meteoric and hydrothermal fluids, and magmatic steam to fluids. As such, ternary diagrams were used to check for any changes in the type of fluids feeding a sampling point through time. Time series of the physical (temperature, pH and Total Dissolved Solids, or TDS contents) and chemical (major, minor and trace elements) characteristics of waters were then constructed and compared against the long term trends in deformation and VT seismicity, as per the characteristics of fumarole gases, and with observations in the literature.

Finally, to establish whether unrest since 1950 has impacted the distribution of surface activity, images and maps of activity at Solfatara were compared. The chronological catalogue of references to visual observations of changes in hydrothermal features was then reviewed to determine if uplift episodes are associated with changes in the intensity of surface activity, and if these changes are consistent over successive episodes.

4.3.2 Analysis of the Distribution of Seismicity Relative to the Position of the Hydrothermal System Below Pozzuoli

To determine where fracturing and faulting was occurring in the crust, and therefore potential changes in fluid flow paths, the distribution of seismicity relative to location of hydrothermal reservoirs in the assumed model of the crust (Chapter 2) was analysed. Using earthquake location from the Vesuvius Observatory catalogue, a subset of data was extracted in ArcGIS, so that only earthquakes in the main cluster of seismicity, in the region of Pozzuoli, were included (n= 3079 events, Fig. 4.6). This was done to exclude events from outside the main deforming area in other locations in the caldera. The epicentral distribution relative to the location of the hydrothermal basin below Pozzuoli (as defined by the Vp/Vs anomaly at 1, 2 and 3 km depth identified by Aster and Meyer, 1988) was then mapped in ArcMap. The Kernel Density distribution of earthquake epicentres was then calculated using the software to identify where seismicity was concentrated within the primary cluster using all events and in 1 km depth intervals (0-1, 1.1-2, 2.1-3 and more than 3 km depth) through the crust. This was then repeated selecting events of particular magnitudes (Ms 0.2 to 1, 1.1 to 2, 2.1 to 3 and more than 3.1), to see where the largest slip events were located.

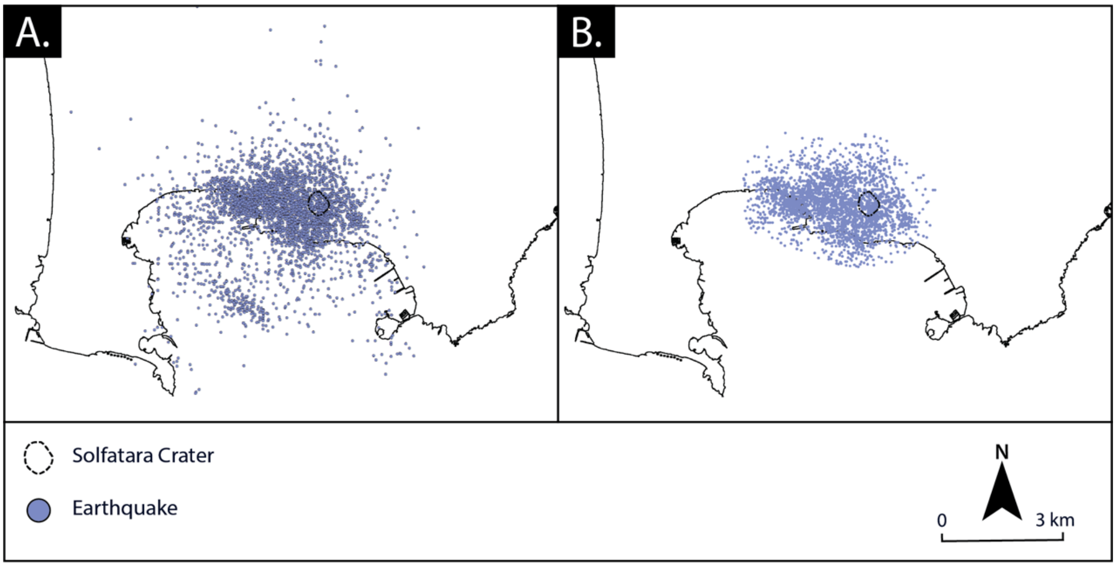


Figure 4.6: Epicentre distribution between January 1983-September 1984. A. includes all events within the catalogue. B. includes events in the main cluster only. Data from the Vesuvius Observatory.

To establish the depths at which seismicity was concentrated in relation to the hydrothermal system, and where the highest magnitude events occurred, hypocentre locations were plotted in 3D space with the aforementioned Aster and Meyer (1988) seismic anomaly in ArcScene. Additionally, histograms with 0.5 km bin widths were constructed to check the frequency of VT events of a particular magnitude with depth. Stress variation in the crust was then investigated by plotting seismic b-values as a function of depth. This parameter describes the geometry of the fault network triggering seismicity for a seismic catalogue where the frequency-magnitude distribution can be described by the Gutenberg-Richter (GR) relation:

$$Log\ \frac{N}{N_R} = -b(\mathrm{M} - M_R) \quad [1]$$

where N is the total number of earthquakes with magnitude greater than or equal to M, N_R is the number of events with magnitudes greater than or equal to a reference magnitude M_R and b is the b-value (Gutenberg and Richter, 1944). The choice of reference scale is arbitrary as the GR distribution is scale-independent (Main, 1996). M_R is commonly taken to be 0 and log N_R expressed as a constant, a, giving the more familiar form of the GR trend, $Log(N) = a - bM$. The b-value can then be considered to be a measure of the dimension D of a fault network and its distribution with depth can be used as an indicator of where stress and fracture growth is concentrated in the crust. Higher values indicate a denser packing and D-values of 1, 2 and 3, or b-values of 0.5, 1 and 1.5, correspond to network development in preferentially one, two and three directions (Turcotte, 1986; Main, 1996). Values of b of 1 ± 0.5 are typical for tectonic earthquake catalogues (Meredith et al., 1990; Frolich and Davis, 1993). However, volcanic seismicity b-values vary from <1 to 3. High values may result from additional fracturing at smaller scales, high temperatures and elevated pore pressures (Wyss et al., 2001; Mc Nutt et al., 2005; Farrell et al., 2009).

To determine the variation of b-values with depth to compare with the expected location of hydrothermal reservoirs, it was first necessary to establish the magnitude of completeness (Mc) of the catalogue (Fig. 4.7). This is the magnitude below which the Frequency-Magnitude Distribution deviates from the GR relation and where the catalogue can no longer be considered representative. A deviation from the trend for low magnitudes is common and is usually interpreted to reflect the limitation of monitoring networks to record all small-magnitude events. The Mc of the catalogue used here was calculated as Ms 1.2. Once events below this magnitude were removed from the data set, seismic b-values were then calculated as a function of depth. The plotting of the Frequency-Magnitude Distribution, calculation of the Magnitude of Completeness and b-values was done using the seismic analysis software ZMap in MATLAB (Wiemer, 2001).

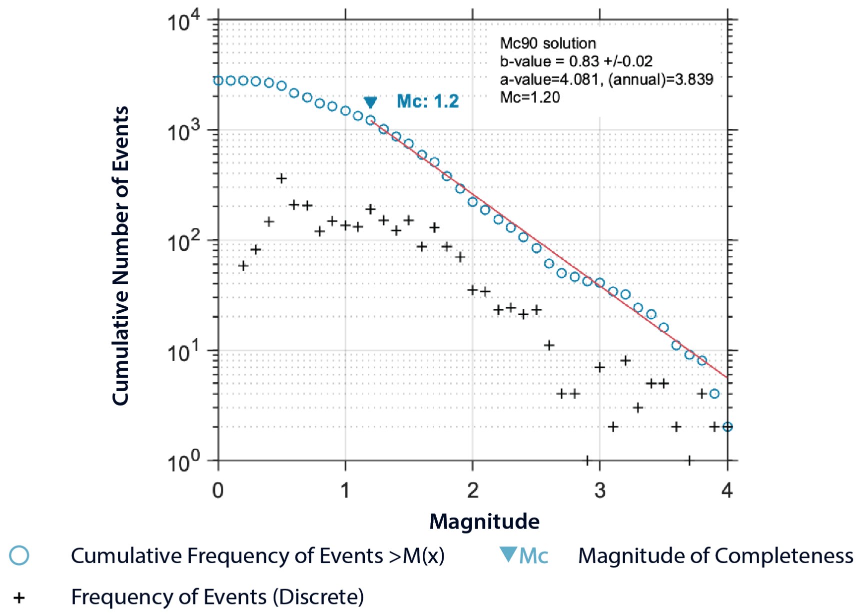


Figure 4.7: Frequency Magnitude Distribution (FMD). The b-value is calculated for events above the magnitude of completeness (Mc = 1.2). Plotted using ZMap (Wiemer, 2001).

4.4 Limitations and Sources of Uncertainty

The primary limitation in the investigation of long-term temporal trends in monitoring observations is that, with the exception of the height of the ground level at Benchmark 25A/cGPS station RITE, it is not possible to extend time series back to the onset of unrest in 1950 or to the present (2019) for all parameters included in the dataset. This is due to differences in the length of time a parameter has been monitored, variations in sampling and in reporting frequency through time, and access to data. However, all datasets sample the 1982-1984 uplift and post-1984 ground movements, which is the critical period of interest, except thermal water data, which are only available until 1999. A second limitation is the use of secondary data sources and the digitisation of points from published materials. To minimise the uncertainty in quantitative data, only values from official sources (e.g. the Vesuvius Observatory) and peer-reviewed journals have been included. Digitised data was checked against the original source for any inconsistencies before inclusion. A final source of uncertainty is in the translation of Italian language materials into English. To ensure that the original meaning was maintained following translation with online tools, any ambiguities were checked with a fluent Italian speaker (C. Kilburn, University College London).

Chapter 5

Behaviour of the Campi Flegrei Caldera

This aim of this chapter is to determine the known recent (20^{th} Century onwards) behaviour of the caldera. It reviews existing observations and interpretations of unrest in the literature in combination with trends in monitoring data, from which the changing characteristics of deformation through time and the relationship between ground movements and other parameters is established. It then goes on to summarise the key observations of activity through time, which form the basis of the reinterpretation of unrest presented in Chapter 6.

5.1 Caldera Unrest at Campi Flegrei

5.1.1 Deformation Since 1905

The vertical deformation profile at Benchmark 25A/cGPS station RITE can be divided into four distinct periods based on the characteristics of ground movements (Fig. 5.1). Caldera-wide subsidence during the quiescent period between 1905-1950 is a continuation of the subsidence that followed the end of the Monte Nuovo eruption in 1538 that, during this period, occurred at a mean rate of c. 2.8 cm yr^{-1}. Between 1950-1984, following the onset of unrest, the profile is dominated by the three major uplifts in 1950-1952 (c. 0.33 m yr^{-1}), 1969-1972 (c. 0.57 m yr^{-1}) and 1982-1984 (c. 0.69 m yr^{-1}). No significant subsidence was recorded after the first uplift, whilst the second episode in 1969-1972 was immediately followed by a lowering of the ground level by c. 0.2 m over a period of three years. The ground level then oscillated about the mean by c. 0.1-0.15 m until 1982. In contrast, the 1982-1984 uplift was followed by a much greater amount of subsidence of 0.9 m that occurred from 1985 to 2004. The rate of displacement decreased exponentially from an initial c. 14 cm yr^{-1} before the ground level stabilised in 2004. The most recent uplift phase is distinct from previous episodes in that it is characterised by a comparatively slow mean uplift rate of 0.035 m yr^{-1} and, so far, has progressed at essentially the inverse rate of the preceding subsidence (Fig. 5.2). As of the present (2019), the ground level at this point is c. 3.4 m higher than at the onset of unrest in 1950.

Superimposed on the post-1984 ground movements are recurring low-amplitude (0.04-0.1 m), inflation-deflation cycles in 1989, 1994, 2000, 2006, 2012–

2013 and 2016-2017 that do not contribute to the overall deformation trend (Gaeta et al., 2003; Troise et al., 2007; Chiodini et al., 2012; D'Auria et al., 2015). They are referred to in the literature as 'mini-uplift' events and are generally regarded to represent minor fluctuations in the pressure of the hydrothermal system below Pozzuoli, although it has been proposed that the 2012-2013 mini-uplift was the result of a small dyke intrusion (D'Auria et al., 2015).

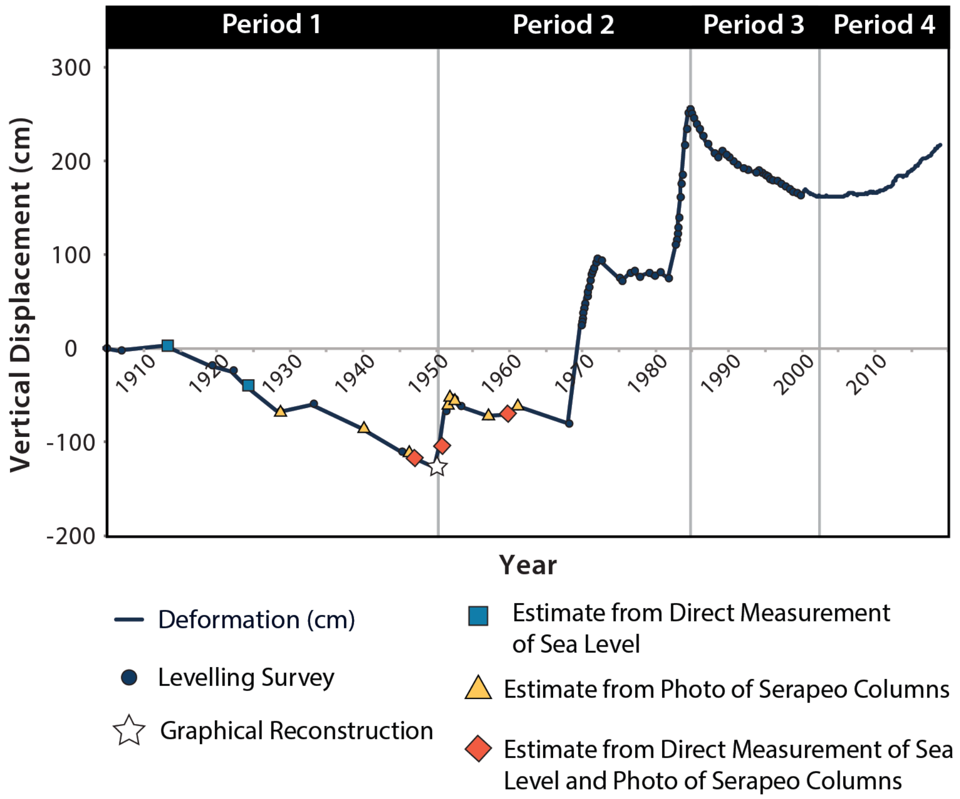


Figure 5.1: Vertical ground movements at Benchmark 25A/cGPS Station RITE. This is the closest station to the centre of unrest near Pozzuoli. Data from 1905 to April 2000 is from the reconstruction of ground level by Del Gaudio et al. (2010) created using a combination of direct and indirect measurements (as indicated in the figure). Data points for May 2000 to July 2016 are cGPS measurements from a supplementary file in Chiodini et al. (2017), whilst those after July 2016 are cGPS measurements digitised from INGV-OV (2019).

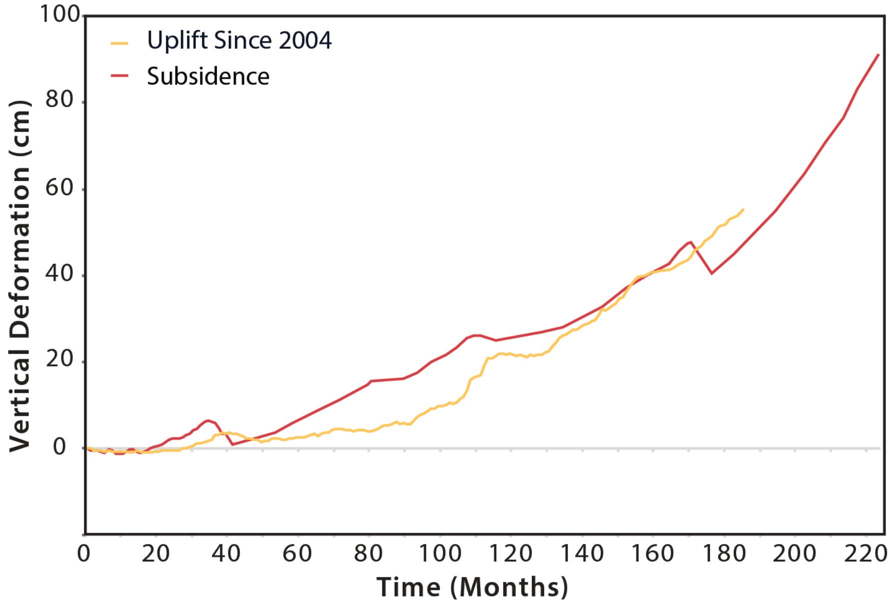


Figure 5.2: Comparison of post-1984 ground movements. The pink line is the subsidence trend reflected on the y-axis. The yellow line is the uplift trend from 2004 to 2019. Data points for the ground level between 1985 to May 2000 are from levelling surveys in Del Gaudio et al. (2010). Data points for May 2000 to July 2016 are cGPS measurements from a Chiodini et al. (2017) and those after July 2016 are cGPS measurements digitised from INGV-OV (2019).

The geometry of caldera-wide vertical ground deformation since 1905 is approximately constant (Fig. 5.3), irrespective of the rate, magnitude and direction of displacement (De Natale and Pingue, 1993; Orsi et al., 1999b; Folch and Gottsman, 2006; Di Vito et al., 2016). The maximum displacements since 1970 have been recorded at Benchmark 25a (cGPS station RITE), c. 0.8 km east of the centre of Pozzuoli, whilst the centre of deformation is located offshore (e.g. Bianchi et al., 1984 and 1987). Deformation decays regularly from the maximum to negligible values at distances at 5.5 km. The field of maximum deformation is elongated NW-SE along the La Starza Marine Terrace fault system and broadly corresponds to the P-S wave anomaly below Pozzuoli that has been interpreted as a pervasively fractured volume containing hydrothermal reservoirs (Fig. 5.4; Aster and Meyer 1988; De Siena et al., 2010; see Chapter 2, section 2.2.2). Horizontal deformation data from surveys conducted during major unrest in 1970-1972 and between 1982-1983 suggests that the geometry of the horizontal deformation was at least also constant between these two episodes. In both cases the displacement pattern was asymmetric, with a greater component of horizontal deformation NW-SE. This is parallel to La Starza, relative to the N-S direction (Dequal, 1972; De Michelis et al., 1978; Corrado et al., 1977; Bianchi et al., 1987; Dvorak and Berrino, 1991; Woo, 2007).

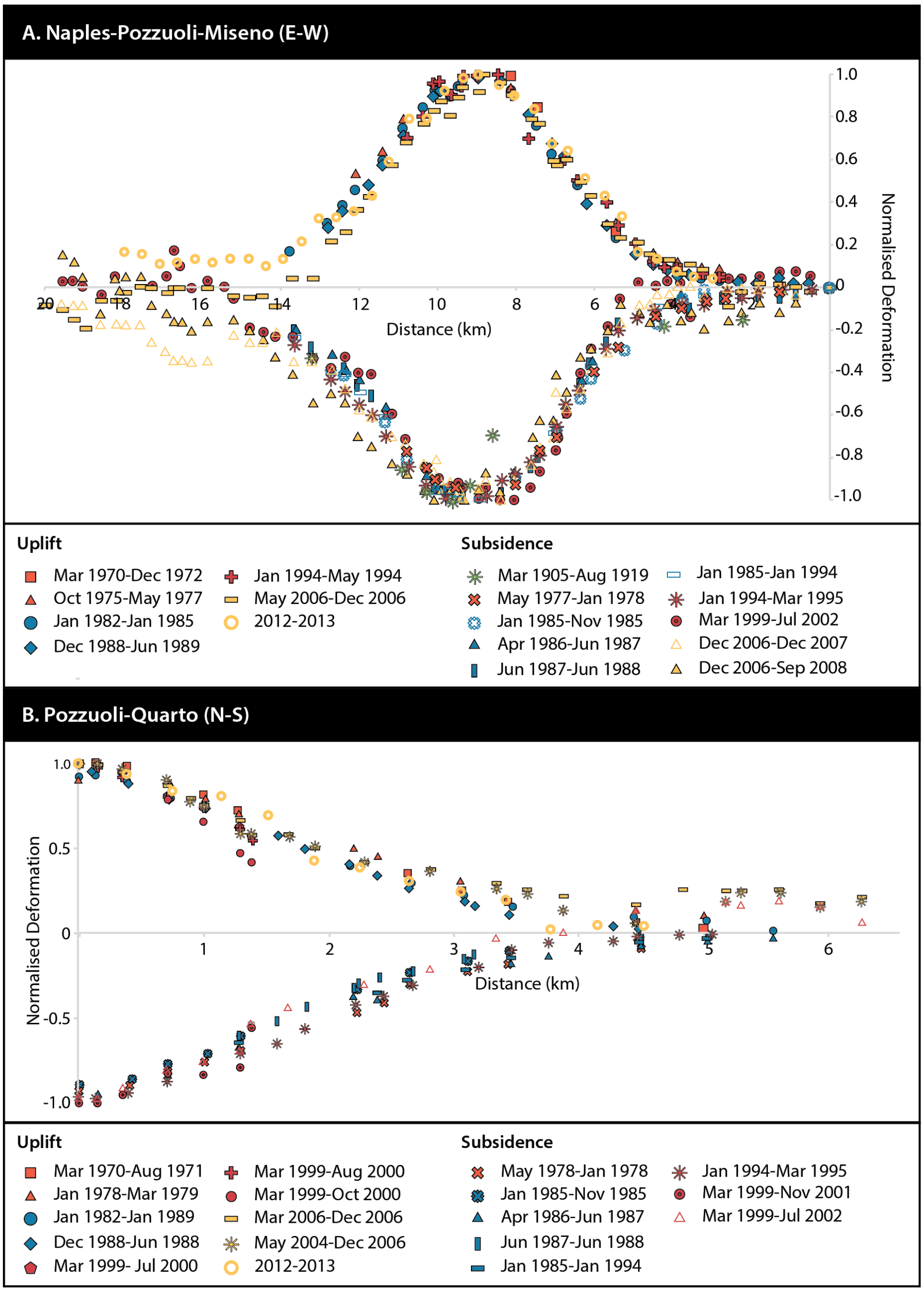


Figure 5.3: Normalised vertical deformation along the East-West (A) and North-South (B) levelling lines for intervals between 1905 and 2008 highlighting the symmetry in up and down displacements and the constancy in shape through time. The data is normalised to the maximum for the period. A degree of the horizontal scatter results from digitisation of poor resolution sources. Modified from Orsi et al. 1999b with the addition of data digitised from Dvorak and Berrino (1991), Pingue et al. (2005), Del Gaudio et al. (2007 and 2009) and Trasatti et al. (2015).

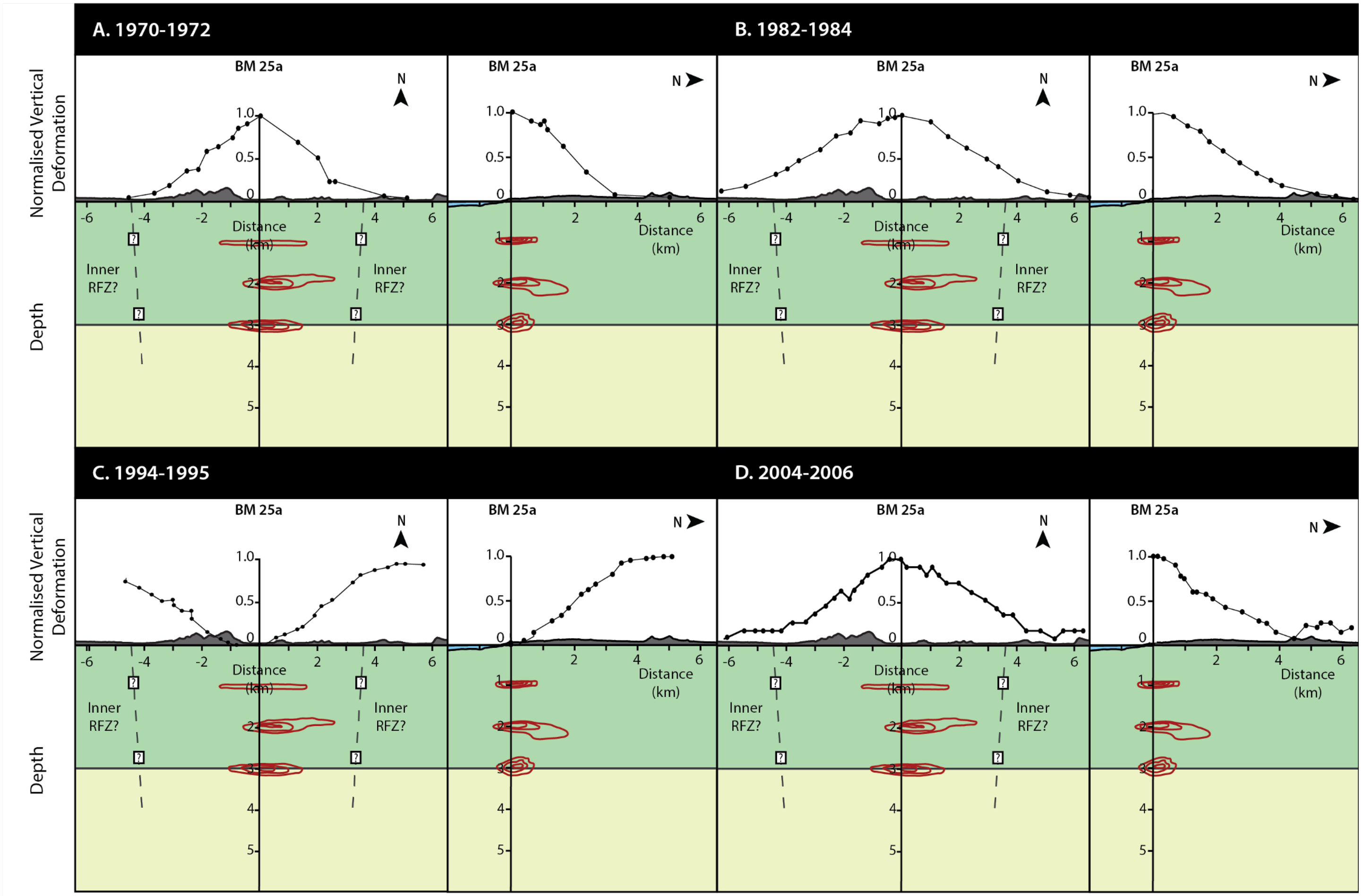


Figure 5.4: The relationship between the maximum displacement during uplift and subsidence phases relative to the position of the hydrothermal system, as inferred from Vp/Vs anomalies at 1, 2 and 3 km depth (red lines). Position of the seismic anomalies are digitised from Aster and Meyer (1988). Levelling data is from Orsi et al. (1999b) and Del Gaudio et al., (2007).

The uplift and subsidence periods are each consistent with the deformation of the crust in response to a pressure change in a discrete source (Woo, 2007). Published estimates of the source depths from inversions of deformation or gravity data for ground movements since 1970 are summarised in Fig. 5.5 and indicate that sources have most likely been located in the shallow crust. The solutions for a given period are non-unique and dependent upon the assumed geometry of the deformation source, the elastic properties of the crust and whether the effects of structural discontinuities and crustal heterogeneity have been taken into account. As such, it is not possible to determine if there has been any migration in the source depth location through time. However, there is a general preference amongst models for source depths between 2-4 km depth for uplifts and 2-3 km depth for post-1984 subsidence. In almost all cases where a single source is assumed, it is located at depths less than 4 km. Such depths are compatible with magmatic, magmatic fluid or hydrothermal pressure sources.

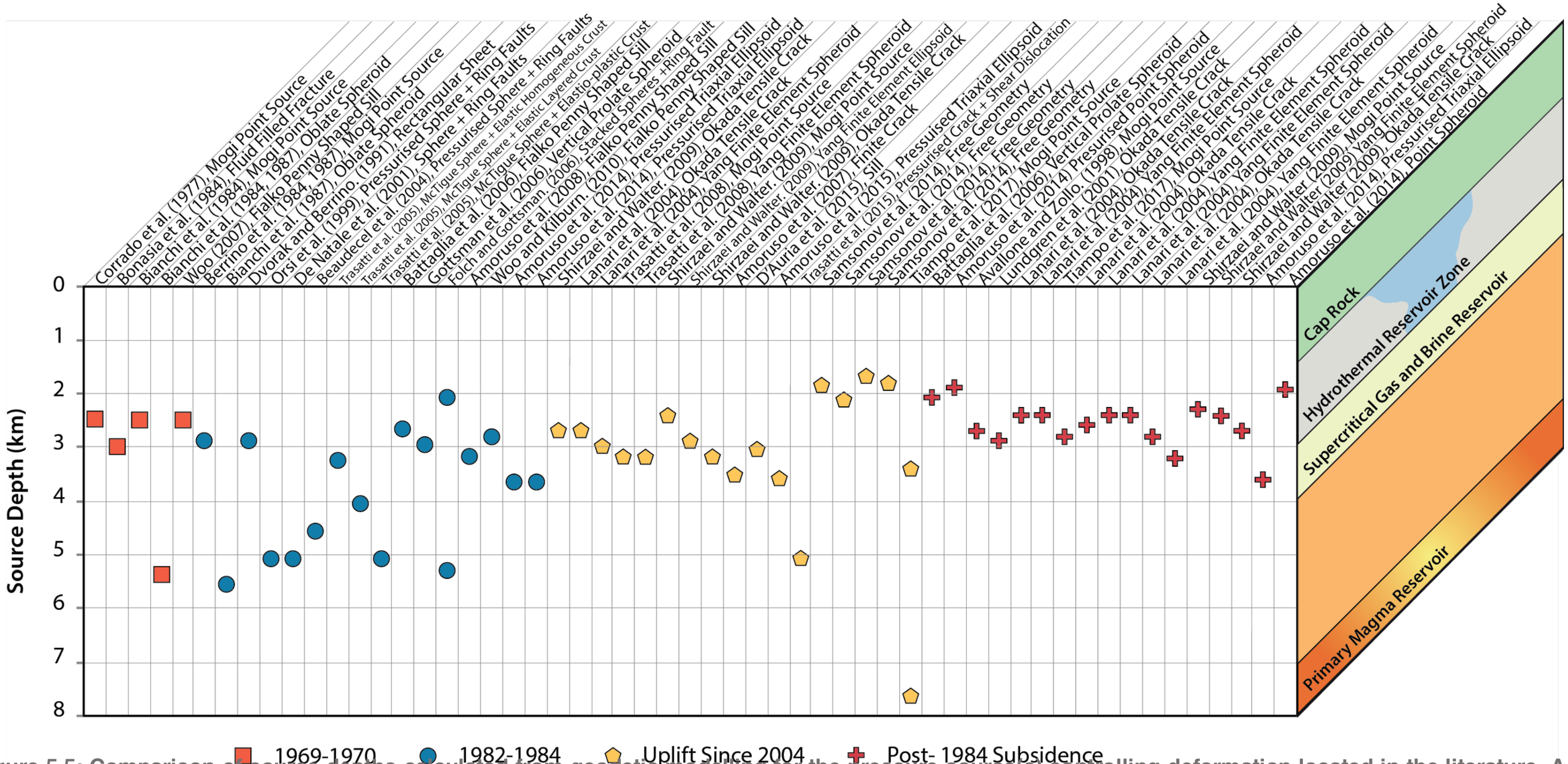


Figure 5.5: Comparison of source depths calculated from geodetic modelling for the pressure source(s) controlling deformation located in the literature. A general preference for between 2-4 km emerges, regardless of assumptions made for the geometry of the pressure source and the nature of the crust. Sources are listed in the figure. The panel on the right-hand side shows a schematic cross section of the crust from Chapter 2.

5.1.2 Relation between Deformation and Seismicity

Seismicity since 1950 has been dominated by low-energy swarms of Volcano-Tectonic (VT) events. 98% have had a magnitude of <2.5, which indicates slip along faults ~0.1-1 km across (D'Auria et al., 2015; Kilburn, 2017). The occurrence of the seismicity is strongly associated with uplift periods (Fig. 5.6), and the cumulative number of VT events correlates with the rate of displacement, consistent with seismicity being induced by the deforming source (Orsi et al., 1999b; D'Auria et al., 2011; Chiodini et al., 2017). Between 1970-1972 more than 5000 events of M <2.5 were recorded, with a peak occurrence of c. 300 in June 1972. The following uplift in 1982-1984 initially progressed aseismically, then from August 1982 seismicity increased steadily until spring 1983 when there was an abrupt escalation in the occurrence of events, as well as an increase in their magnitudes (Corrado et al., 1977; Orsi et al., 1999b; Barberi et al., 1984; D'Auria et al., 2011 and 2015). Over 16 000 events were registered during this period with magnitudes between 0.6 and 4.2 (80% < M 2). Events of magnitude 4 indicate a maximum slip ~1 km (Aster and Meyer, 1992; Orsi et al., 1999b; D'Auria et al., 2015). The initial aseismic ground movement may be attributed to the Kaiser effect during elastic-brittle deformation (Kaiser, 1953), where under cyclic loading and unloading the stress of the previous cycle must be exceeded before seismicity occurs (Kilburn et al., 2017). This effect has also been observed during inflation-deflation events at Krafla (Heimisson et al., 2015). Whether seismicity occurred during the 1950-1952 uplift cannot be confirmed as this uplift pre-dates the start of seismic surveillance (Chapter 4).

Following the 1969-1972 uplift, VT event rates decayed until the onset of the 1982-1984 episode. The decay resembles an aftershock sequence (Corrado et al., 1977). In contrast, seismicity ended abruptly at the end of the 1982-1984 uplift and the following subsidence was aseismic except for minor swarms (M <2.5) that occurred in 1987, 1989, 1993, 1994 and 2000, in association with mini-uplifts. Long Period (LP) events with a maximum depth of 4 km were identified in the July 2000 swarm (Saccorotti et al., 2001; Bianco et al., 2004), the source of which is considered to be the harmonic oscillation of a fluid filled reservoir in response to an increase in fluid pressure (Bianco et al., 2004; Sacrarotti et al., 2007; Cusano et al., 2008). Seismic energy release has remained low throughout uplift since 2004 and the cumulative trend shows the same temporal pattern as the ground movement (Chiodini et al., 2017). More than 2000 VT events of magnitude less than 2.5 (80% <M 1) have occurred and

additional LP swarms have been identified in March 2005, October 2006 and January 2008 (Saccorotti et al 2007; Bianco et al., 2004; D'Auria et al., 2011).

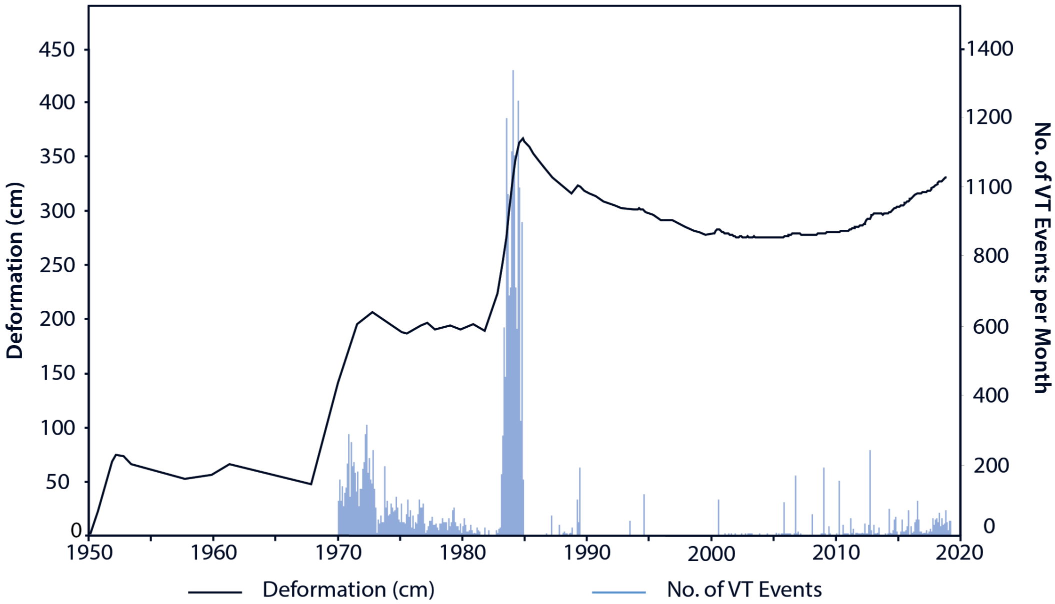


Figure 5.6: Temporal relationship between ground movements and Volcano-Tectonic (VT) seismicity. Ground level data until May 2000 is from Del Gaudio (2010), data points for May 2000 to July 2016 are cGPS measurements from a supplementary file in Chiodini et al. (2017) and those after July 2016 are cGPS measurements digitised from INGV-OV (2019). Rates of VT seismicity are from the combination of data from seismicity catalogues from the Vesuvius Observatory and in Chiodini et al. (2017), and from Vesuvius Observatory monthly bulletins (ov.ingv.it).

The epicentral distribution of VT events through time is given in Fig. 5.7. Since seismic surveillance began in March 1970 the seismicity has been confined to within the Ring Fault Zone (RFZ), as defined from the Bouguer anomaly (Chapter 2, section 2.2.1). The distribution of seismicity has varied over time but the seismogenic volumes were constant throughout individual uplift episodes (D'Auria et al., 2011). During uplift in 1969-1972, VT events were located between 1-5 km depth (predominantly <2km) around the western and northern margins of the main collapse. The highest density of events was recorded in the western sector between Averno and Miseno, rather than within the field of maximum displacement (Rampoldi, 1972; Scherillo, 1977; Corrado et al., 1977). However, this distribution may in part result from the configuration of the network at the time, which was biased to the west, and the low sensitivity of the instruments in operation that could only record nearby events. Following uplift, seismicity at depths greater than 2 km ceased (Corrado et al., 1977).

Seismicity during the 1982-1984 uplift was concentrated in two main clusters; along an offshore NW-SE striking fault dipping at 75-80° to the SW (c. 10% of total, Aster and Meyer, 1992; Di Luccio et al., 2015) and within an elliptical area (6 km x 4 km) centred on Pozzuoli with a long axis elongated parallel to the La Starza Marine Terrace. The Pozzuoli cluster corresponds to the area of greatest uplift. VT events located along the offshore fault occurred between 0-6 km depth and were characterized by reverse to strike-slip focal mechanisms. The highest energy events (M >2.5) along this fault were concentrated at depths greater than 3 km (Orsi et al., 1999b; D'Auria et al., 2015). VT events in the Pozzuoli cluster were also located between 0-6 km depth but were concentrated at depths of less than 4 km (Vilardo et al. 2010). The highest density of epicentres occurred in an area c. 2 km^2 between Solfatara and Agnano that D'Auria et al., (2011) suggest were modulated by the presence of hydrothermal fluids. Focal mechanisms of VT events in the Pozzuoli cluster were found to be dominantly normal with occasional strike-slip and dip-slip events confirming an extensional deformation regime (Orsi et al., 1999b; D'Auria et al., 2015).

During subsidence after 1984, seismicity was confined to an approximately circular area c. 1km^2, centred on a NNE-SSW striking fracture system local to Solfatara (Sacarotti et al., 2001; Bianco et al., 2004). The epicentres cluster in an area that corresponds to the expected location of the vertically extensive fracture zone that connects the main hydrothermal system to the surface at the Solfatara-Pisciarelli Diffuse Degassing Structure (DDS). Events were distributed between the surface and 4 km depth, and clustered at depths shallower than 2.5 km (Fig. 5.8, Orsi et al., 1999b). The deepest earthquakes occurred during the July-August 2000 seismic swarm.

Following the resumption of uplift in 2004, seismicity has been clustered within an area centred on Pozzuoli, with no significant seismicity occurring elsewhere in the caldera. This indicates that the pressure source causing uplift is affecting a smaller volume of the caldera than in previous uplifts. VT event locations broadly coincide with the volume inferred to host the hydrothermal circulation and the highest frequency of events have occurred between 0-2 km depth (Fig. 5.8). A single swarm of c. 200 events (Md ≤1.7) that extended from c. 4 km depth to the surface occurred on 7th September

2012, below the SE flank of Monte Gauro, external to the main seismogenic volume (Amoruso et al., 2014). The highest epicentral density within the main cluster corresponds to the Solfatara-Pisciarelli DDS feeding structure. LP events have generally been of too low energy to locate during this uplift, although some from during the October 2006 LP swarm have been located below the SE rim of the Solfatara crater at c. 500 m depth in a volume distinct from that in which VT events occur (Saccarotti et al., 2007; Cusano et al., 2008). No migration of hypocentres has been observed during any period of uplift or subsidence.

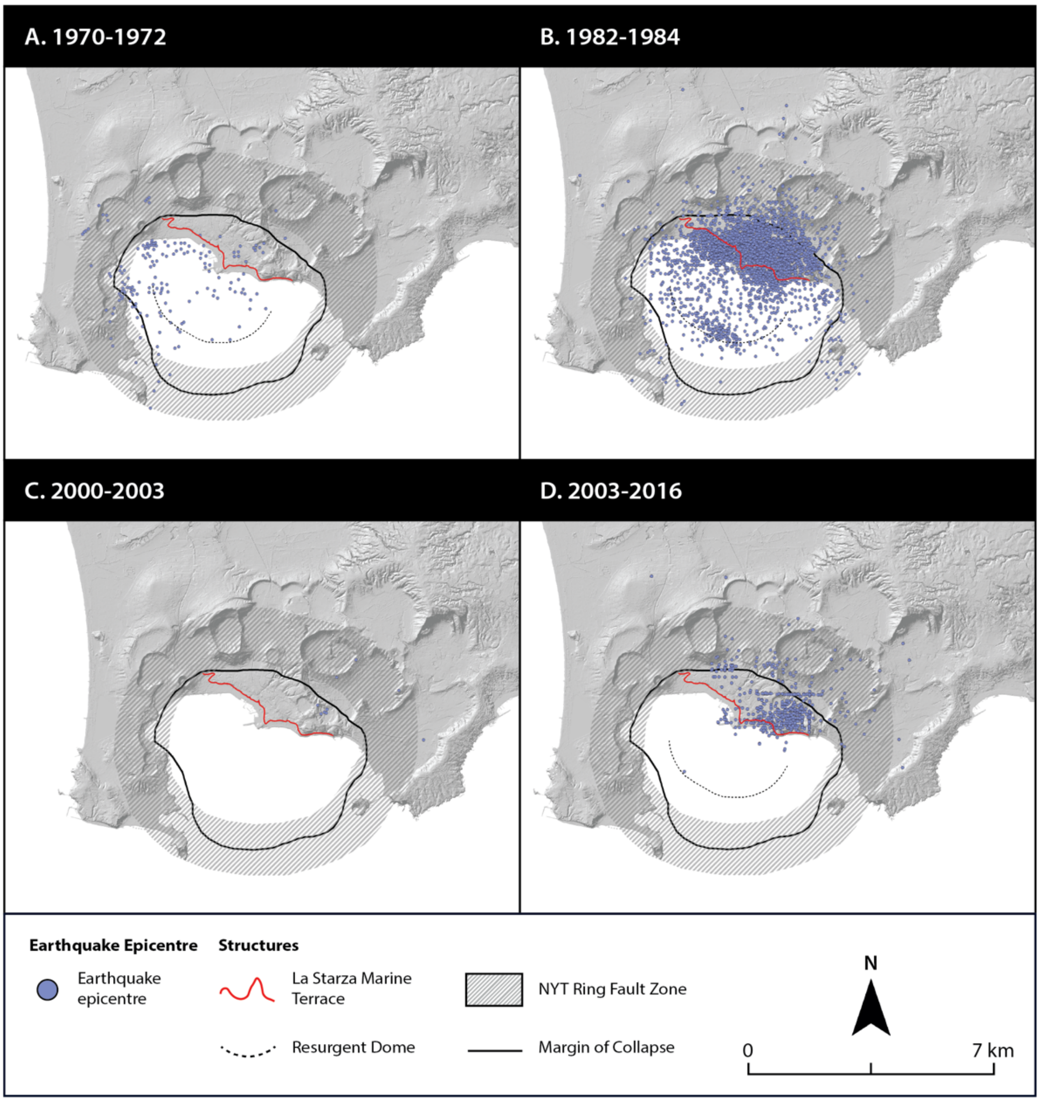


Figure 5.7: Volcano-Tectonic (VT) earthquake epicentres through time. Data for A. is digitised from Corrado et al. (1977) and Orsi et al.(1999b). Epicentre locations in B to D are from an earthquake catalogue from the Vesuvius Observatory.

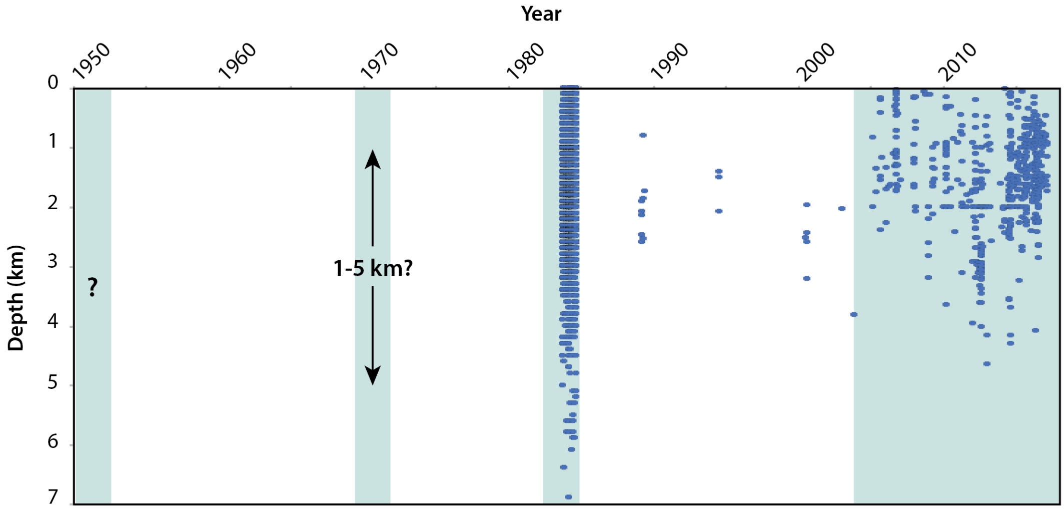


Figure 5.8: Earthquake hypocentre (blue dots) depths through time. The main uplift periods are marked by the shaded boxes. Earthquake locations from an earthquake catalogue from the Vesuvius Observatory.

A significant feature of the long-term trend in rates of VT seismicity is the order of magnitude increase in the number of VT events between the 1969-1972 and 1982-1984 uplift episodes, and the increase in the magnitudes of the largest events, whilst the total displacement in each case is comparable. This may be interpreted in terms of a greater strain rate during the second episode (R. Scandone, pers. comm). However, according to Kilburn et al. (2017) the trend can alternatively be attributed to an increasing component of inelastic deformation with uplift due to the accumulation of stress in an elastic-brittle crust. Under ideal conditions, starting from lithostatic equilibrium, a differential stress applied to the crust (e.g. by a pressure source) is accommodated elastically by stretching unbroken rock. Once a threshold is exceeded, deformation becomes quasi-elastic as a component of the supplied strain energy is lost inelastically by faulting. The inelastic component increases progressively at accelerating rates with applied stress until it becomes the dominant mode of deformation. At this stage, the mean differential stress is held constant as the rate of stress applied by the pressurising source is balanced by the stress lost through faulting (Fig. 5.9 a and b; Kilburn, 2012; Kilburn et al., 2017). The total number of VT events can be considered as a measure of inelastic deformation, whereas uplift is a proxy for the total deformation (the sum of elastic and inelastic components, Kilburn, 2012). By considering the three major uplifts between 1950 and 1984 together, Kilburn et al. (2017) found that the accelerating increase in VT events with total uplift follows the trend expected for a connected sequence of progressive deformation (Fig. 5.9c), and that conditions for the transition from elastic to quasi-elastic behaviour were met during the 1969-1972 uplift, for which the total number of VT events ΣN increased exponentially with uplift Δh. This is described by

$$\Sigma N = (\Sigma N_0)\, e^{(\frac{\Delta h}{\Delta h_{ch}})} \qquad [1]$$

where $\Sigma N_0$is the starting number of VT events and h_{ch} is a characteristic length that describes the specific form of exponential trend. For deformation in extension the ratio $\Delta h/\Delta h_{ch}$ is equivalent to the ratio of applied differential stress to tensile strength, S_d/σ_T, which equals 4 or less for failure in tension and between 4 and 5.6 for failure in extension (Kilburn et al., 2017). The VT events describe changes in the stress field around zones of stress concentration, so that the onset of large-scale rupture (such as the re-opening of a sealed fault) is expected when $\Delta h/\Delta h_{ch}$ approaches its maximum value (Kilburn et al., 2017). According to Kilburn et al. (2017) $\Delta h/\Delta h_{ch}$ had reached a value of 4.2 by the end of uplift in 1984. This would suggest that the crust had approached conditions favourable for bulk failure and therefore the onset of widespread fracturing. This is compatible with both the increasing rates and magnitudes of seismicity over the successive uplifts.

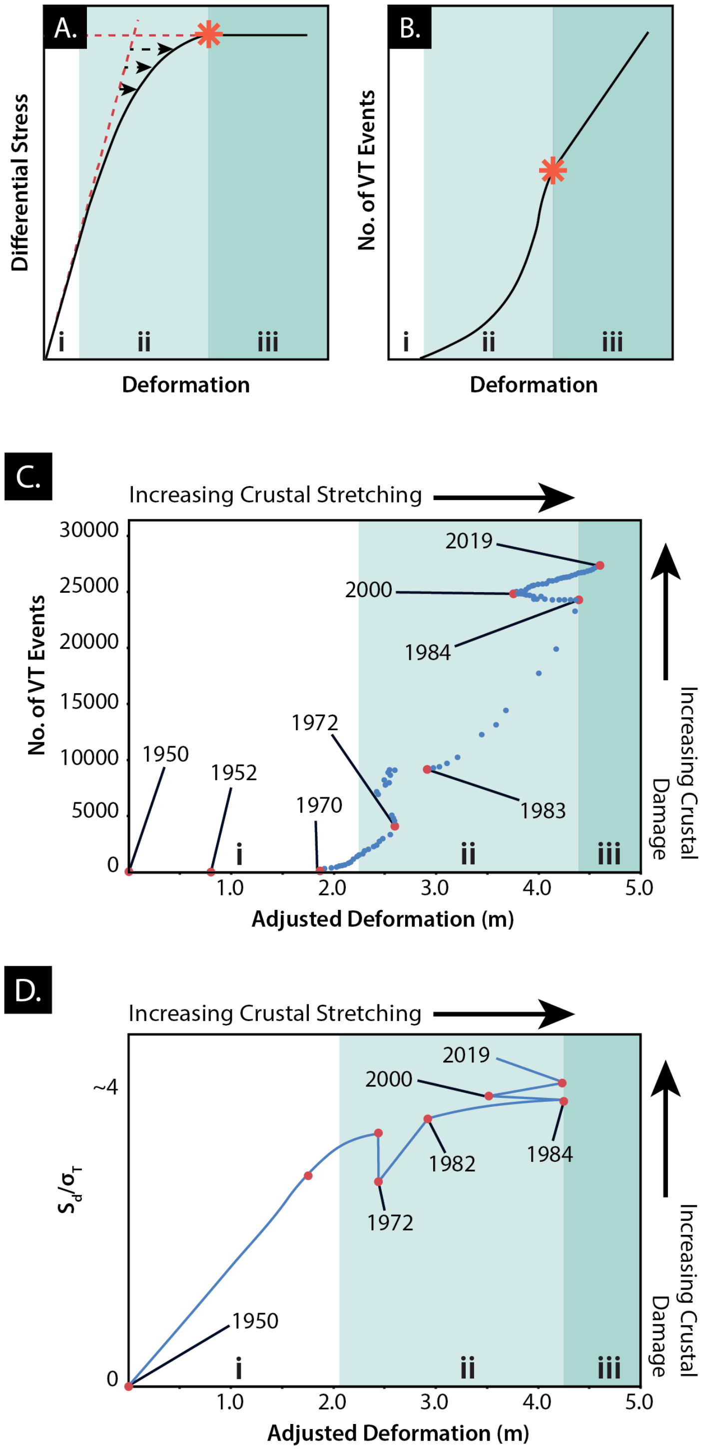


Figure 5.9: Progressive deformation of an elastic-brittle crust (from Kilburn et al., 2017). A. Evolution of the bulk deformation regime with increasing differential stress from elastic (i), to quasi-elastic (ii) to inelastic (iii). B. The evolution is caused by faulting. The total deformation caused by fault movements is represented by the cumulative number of VT events, which increase exponentially in the quasi-elastic regime and linearly with deformation in the inelastic. C. The combined-VT deformation trend for periods of deformation at Campi Flegrei. The exponential trend in VT seismicity suggests deformation has largely occurred in the quasi-elastic regime (ii). D. The trend (blue line) shows: an accumulation of stress; an increase in the proportion of deformation by faulting; and that conditions in 1984 were approaching the inelastic regime. The trend is interrupted by minor relaxation of the crust after the 1969-1972 uplift and the subsidence after 1984. C. and D. assume a deformation trend adjusted for a background subsidence of 1.7 cm yr^{-1}.

5.1.3 Relation between Deformation and Changes in Degassing Features

Fumarole discharge temperatures at the monitored Solfatara vents have been essentially stable throughout the observational period (Fig. 5.10). The mean temperature for Bocca Grande since the onset of monitoring in 1983 is 161.1 ± 2.6 °C, comparable to the mean of 159 °C measured in 1925-1935 and also in 1970 (Signore, 1935; Dall'Aglio et al., 1972). The stability of the temperature is thought to indicate that the temperature of outgassing is controlled by the separation of vapour from a liquid (Cioni et al., 1984) and suggests that there has not been any significant variation in the temperature of fluids entering the gas reservoirs located at c. 100 m depth that feed Bocca Grande through time. This vent and the Bocca Nuova share a gas reservoir but since 2004 a minor heating of c. 5 °C has been recorded at Bocca Grande, whilst Bocca Nuova has cooled by approximately the same amount. This behaviour has been attributed to an increase in gas flux to the surface and the interactions of Bocca Nuova gases with greater volumes of condensates from Bocca Grande, which is upslope (Gresse et al., 2018). Gas temperatures at Pisciarelli are strongly influenced by seasonal effects but between 1999-2005 the mean temperature was observed to increase from c. 95 °C (the boiling temperature at Pisciarelli) to c. 110 °C, thought to be due to an increase in the supply of hot hydrothermal fluids (Chiodini et al., 2011; INGV-OV, 2019).

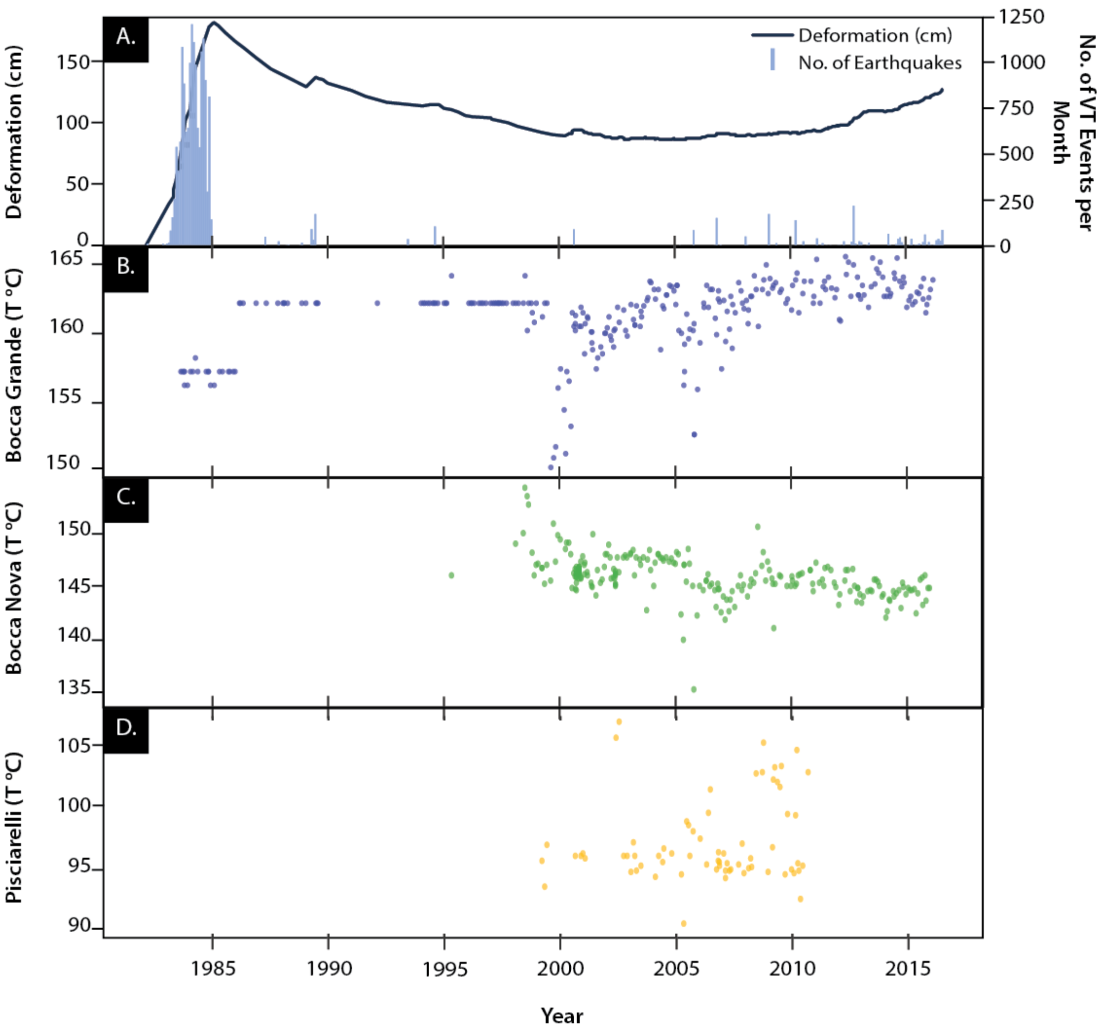


Figure 5.10: Trends in ground level, rates of Volcano-Tectonic (VT) seismicity and fumarole gas temperatures. Ground level data until May 2000 is from Del Gaudio (2010), data points for May 2000 to July 2016 are cGPS measurements from Chiodini et al. (2017) and those after July 2016 are cGPS measurements digitised from INGV-OV (2019). Rates of VT seismicity are from the combination of data from seismicity catalogues from the Vesuvius Observatory and in Chiodini et al. (2017), and from Vesuvius Observatory monthly bulletins (ov.ingv.it). Fumarole temperatures are from Chiodini et al. (2011) and Chiodini et al. (2016).

The temporal trends in fumarole gas compositions from Solfatara-Pisciarelli show systematic variations with ground movements and periods of seismicity (e.g. Chiodini et al., 2003, 2009, 2012 and 2015a; Moretti et al., 2013, 2017 and 2018). The trends discussed here are related to gases from the Bocca Grande vent, which has been continuously monitored for the longest period. The gas concentrations from this vent show essentially the same variations with time as those from Bocca Nuova, as both are fed by the same shallow gas reservoir at c. 60 m (Gresse et al., 2018). Similar trends in gases are also observed from the Pisciarelli vent, as all three fumaroles are supplied by fluids transported to the surface through the Solfatara-Pisciarelli DDS.

At Bocca Grande, in 1983 there was an enrichment in H_2O and the reduced hydrothermal gases H_2S, H_2 and CH_4 (Cioni et al., 1989; Martini et al., 1984 and 1986). An isolated sample from May 1982 was found to have elevated CH_4 relative to samples taken in 1978-1981, suggesting that enrichment occurred after the onset of uplift in 1982 (Barberi et al., 1984). Concentrations of these gases peaked in 1983 then rapidly declined to minimum values in 1985 (Fig. 5.11). The initial increase was attributed to heating and boiling of the deep hydrothermal system and considered to be a precursor to seismicity (Carapezza et al., 1984; Cioni et al., 1989; De Natale et al., 1991). These trends have since been reinterpreted and the favoured explanation today is that the initial enrichment in hydrothermal gases reflects increased fluid flux through the DDS feeding structure ahead of an injection of magmatic gas-rich fluids into the main hydrothermal system (Chiodini et al., 2003). This is based on the progressive increase between 1983 to 1985 in the CO_2/H_2O ratio (Fig. 5.11a), which is controlled by the concentration of magmatic CO_2 and the phase of H_2O in the Solfatara plume. Support for an input of magmatic fluids comes from the concurrent peaks in the redox indicators CO_2/CH_4 and CO_2/H_2S, which indicate that the formation of H_2S and CH_4 was being suppressed by more oxidising conditions (Chiodini et al., 2009). The CO_2 peak coincided with a sharp increase in N_2, which has a $\delta^{15}N$ signature of 6-6.7 ‰, close to values for subduction zone fluids ($\delta^{15}N = 7 \pm 4$‰) and distinct from that of an atmospheric source ($\delta^{15}N = 0$ ‰, Chiodini et al., 2010). Given the isotopic signature of N_2 and that it is less soluble than CO_2 in magma, the increase in concentration can be considered to indicate the transport of N_2-rich magmatic fluids from the magmatic system into the hydrothermal reservoirs below Pozzuoli (Giggenbach, 1980; Caliro et al., 2007 and 2014; Chiodini et al., 2015a). Gases such as SO_2, HCl and HF that are diagnostic of shallow magma were not detected during the uplift.

Between 1985 and 2000 the concentrations of N_2 and CO_2 oscillated together about a mean trend that declined as subsidence proceeded. Opposing interpretations for the depletion in the literature suggest that the trend results from either a change in the composition of gases entering the hydrothermal system from a shallow magma intrusion (<4 km depth) as it depressurised (e.g. Caliro et al., 2014; Chiodini et al., 2015a; 2016), or the progressive removal of CO_2-N_2 rich fluids from the hydrothermal system following the emplacement of a shallow magma intrusion (Moretti et al. 2013; 2017). The downward trend was interrupted by two transient increases in the concentrations of these gases in 1990 and 1994. These coincided with peaks in

CO_2/CH_4, N_2/CH_4 and He/CH_4, indicative of more oxidising conditions. As such they have been interpreted as reflecting inputs of magmatic gas-rich fluids into the DDS feeding system following mini-uplift events and associated seismic swarms (Tedesco and Scarsi, 1999; Chiodini et al., 2009 and 2010). Chiodini et al. (2003) found that peaks in the ratio CO_2/H_2O lagged maxima in ground level by c. 200 days and concluded that this represented the travel time for the gases to reach the surface from a magma body stored between 3-4 km depth. Between 1985-2000 H_2O concentrations consistently acted in the opposite direction to those of CO_2, so that there is an overall increase in H_2O vapour discharged at the surface over this period.

The processes controlling the geochemistry of Solfatara fumarole gases changed in 2000 (Fig. 5.11a). The common behaviour between N_2 and CO_2 trends was lost and replaced by a progressive enrichment in CO_2 that preceded the onset of uplift in 2004, whilst N_2 concentrations have remained stable at minimum values (Fig. 5.12; Chiodini et al., 2015a; 2017; INGV-OV, 2019). The trend in H_2O concentrations has been the inverse of that of CO_2, indicating a progressive reduction in the vapour fraction of hydrothermal fluids feeding surface activity (Chiodini et al., 2015a; Moretti et al., 2017). At present (2019), the concentration of CO_2 is the highest since systematic monitoring at Bocca Grande began in 1983. Other notable changes in gas geochemistry after 2000 include a continuous increase in CO, reflecting heating of the shallow subsurface (c. upper 500 m, Chiodini et al., 2016) and an increased frequency in peaks in CO_2/CH_4 that have smaller magnitudes relative to those that occurred between 1985-2000 (Fig. 5.11c). These peaks have been interpreted as reflecting more regular inputs of oxidising (i.e. magmatic) fluids into the hydrothermal system since 2000 (e.g. Chiodini et al., 2015a). The concentration of He also increased between 2000 and 2010 but because isotopic data are unavailable it is not possible to distinguish whether the trend was controlled by magmatic (^3He) or crustal He (^4He).

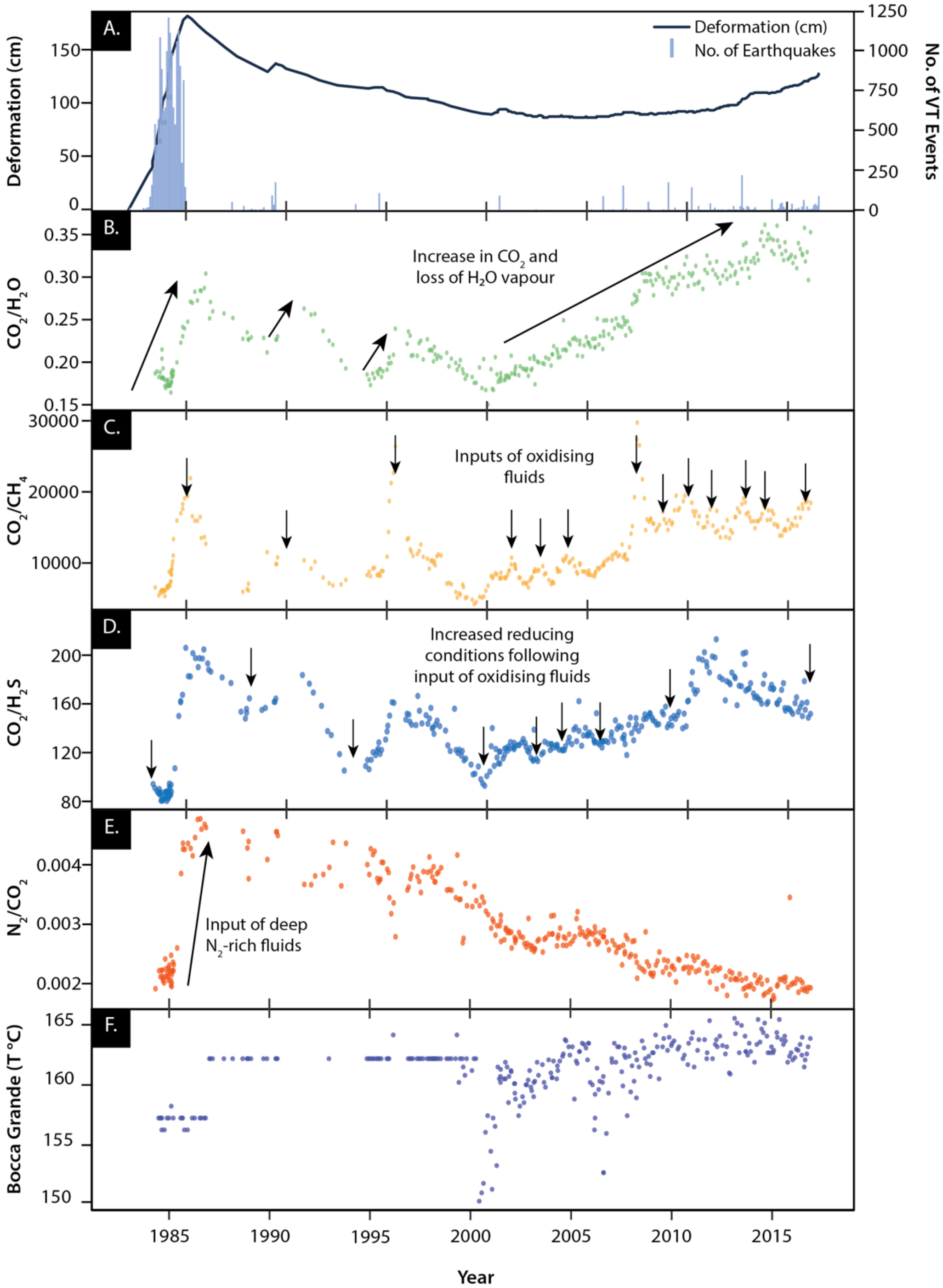


Figure 5.11: Trends in key indicator gas ratios at Bocca Grande fumarole between 1983 to 2016. Panel A is the change in ground level since 1982 and the number of Volcano Tectonic (VT) events, whilst B to E compare changes in gas ratios. Panels B-D indicate an input of oxidising fluids during the 1982-1984 uplift and subsequent decline during the following subsidence. After 2000, conditions progressively became more oxidising, as indicated by the overall upward trends in Panels B-D. Panel E highlights the lack of significant N_2 input after 1985. Data from Chiodini et al. (2017). Interpretations are from the literature as summarised in section 5.1.3. Ground level data until May 2000 is from Del Gaudio (2010), data points for May 2000 to July 2016 are cGPS measurements from a supplementary file in Chiodini et al. (2017) and those after July 2016 are cGPS measurements digitised from the INGV-OV Bulletin. Rates of VT seismicity are from the combination of data from seismicity catalogues from the Vesuvius Observatory and in Chiodini et al. (2017), and from Vesuvius Observatory monthly bulletins (ov.ingv.it). Gas data is from Chiodini et al. (2016).

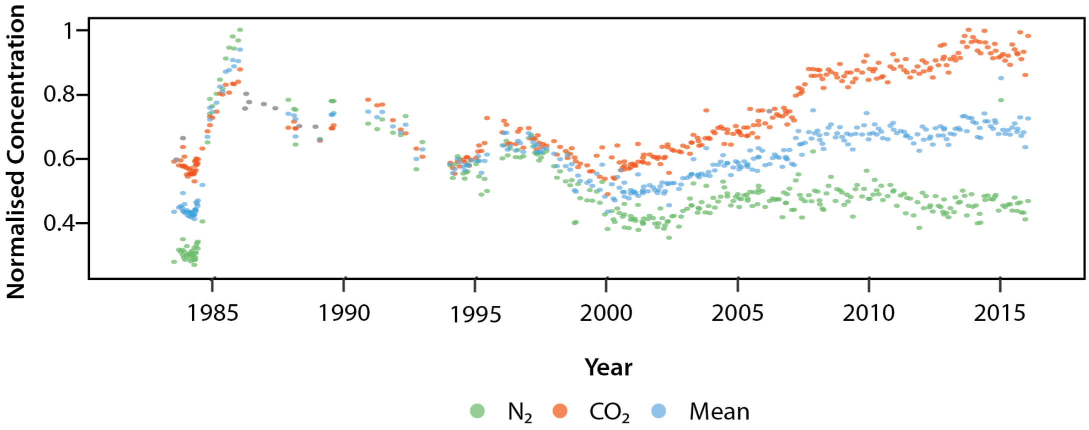


Figure 5.12: Variance in N_2 and CO_2 between 1983 to 2016. Concentrations have been normalised to the maximum concentration of each gas. Data from Chiodini et al. (2016).

In addition to a change in the behaviour of geochemical trends in fumarole gases from 2000, there has been an expansion of the diffuse degassing area, which doubled in 2003-2004 and then increased by a further 30% to 1.2 km^2 in 2011-2012 (Cardellini et al., 2017). The expansion was accompanied by a doubling of the total CO_2 diffuse flux from the ground (i.e. not including from the main fumarole vents) from 750-800 t d^{-1} in 2003 to more than 1500 t d^{-1} after 2014 (peak of 2800 t d^{-1} in January 2015, Granieri et al., 2010; Cardellini et al., 2017). The most significant increases have occurred at Pisciarelli where the flux rate increased from c. 90 t d^{-1} in 2003 to 260 t d^{-1} in 2016 (Cardellini et al., 2017). Temporal trends in the flux from fumarole vents are unavailable but a survey using the CO_2 DIAL remote sensing system calculated fluxes of CO_2 of 266 ± 212 t d^{-1} for Pisciarelli and 715 ± 394 t d^{-1} for the main Solfatara fumaroles in 2015, which were elevated with respect to surveys conducted in 2012-2013 using MULTIGAS and GasFinder techniques that estimated fluxes of 150-200 t d^{-1} and 250-300 t d^{-1} respectively (Aiuppa et al., 2013; Pedone et al., 2014; Quießer et al., 2016). A subsequent survey using the LARSS system in May 2017 at Pisciarelli suggests a possible further increase to 578 ± 246 t d^{-1} (Quießer et al. 2017).

Estimates of CO_2 flux of 15.5-120 t d^{-1} from Solfatara fumaroles during uplift in 1982-1984 are comparatively very low, but these values cannot be reliably compared to modern analyses due to differences in sampling and measurement (Italiano et al., 1984; Allard, 1992). However, increased magmatic gas-rich fluids from fumaroles, reported intensification of fumarolic activity and an expansion of the Fangaia mud pools during the uplift are compatible with increased gas flux during the uplift (Italiano et al., 1984; Bianchi et al., 1990). An intensification of degassing throughout the 1969-1972 uplift and an expansion of the Fangaia was also recorded in 1969-1972 (Casertano et al., 1976; Scherillo, 1977) and an unverified widening of the mud pools in 1950-1952 may also have occurred (Del Gaudio et al., 2010 and references therein).

5.1.4 Relation between Deformation and Changes in Phreatic Waters

The locations of thermal water sampling sites for which monitoring data was located in the literature and temporal trends in their characteristics between 1970-1999 are given in Appendix A. There is no evidence for systematic caldera-wide changes in the chemistry of phreatic waters that can be related to ground movements (Ghiara et al., 1988; Martini et al., 1991). Qualitative references to a greater thermal input to the hydrothermal system during the 1969-1972 uplift relative to that in 1982-1984 have been found in the literature but it has not been possible to confirm these with quantitative data (e.g. Martini et al., 1991; Celico et al., 1992b; Valentino et al., 2004).

Local site responses related to changes in permeability in response to the 1969-1972 and 1982-1984 uplifts were evident at several locations (Ghiara et al., 1988; Martini et al., 1991; Celico et al., 1992b). The most notable changes in water chemistry occurred in thermal waters located at the Stufe di Nerone, which is fed by the Mofete-Monte Nuovo circulation, and at the Terme Puteolane and Hotel Tennis sampling sites, both of which are supplied by hydrothermal fluids from the reservoir below Pozzuoli. At the Stufe di Nerone, uplift in 1982-1984 triggered a long-term change in the permeability of the fault system that feeds the surface waters, which has resulted in a progressive dilution of the Na-Cl rich hydrothermal component by meteoric water (Ghiara et al., 1988). This is evidenced by a continuous decline in the Total Dissolved Solids (TDS) contents from 1984/1985 onwards, and an accompanying increase in HCO_3 (meteoric waters are enriched in HCO_3). Additionally, there has been a shift in $\delta^{18}O$ and δD values towards the meteoric endmember, as well as a change to isotopically lighter $\delta^{34}S$ values from 19.5‰ in April 1971 to 17.4‰ in March 1994 (Ghiara et al., 1988; Celico et al., 1992b; Valentino et al., 1999). At Terme Puteolane waters also record changes in the supply of a deep hydrothermal component over time. This resulted in a decrease in the discharge temperature between the 1969-1972 and 1982-1984 uplifts from 73 °C to 51°C, with a concurrent decrease in SiO_2, NaCl and SO_4 contents (Valentino et al., 1999). Temperatures at this location were stable after the 1982-1984 uplift and NaCl and SO_4 concentrations recovered thereafter. Samples taken in the 1970s, between 1983-1985 and in 1989 are enriched in an HCO_3-rich fluid relative to samples from 1990-1999. Given the location of the Terme Puteolane on the margin of the outflow from the Solfatara plume this is may be attributed to a greater input of CO_2-rich steam during unrest in 1969-1972 and 1982-1984, and the mini-uplift in 1989, although the data are insufficient for a definitive conclusion. On the basis of a low temperature measurement of 51 °C made in 1970 and a greater component of magmatic S in samples from 1971 and 1975 relative to samples collected in 1994 (based on $\delta^{34}S$ values), Valentino et al. (1999) suggested that chemical and isotopic trends could be accounted for by an input of hydrothermal fluids from the deep, magmatic gas rich reservoir, into the Term Puteolane feeding system during the 1969-1972 uplift.

The greatest compositional changes through time have been observed in waters from the Hotel Tennis sampling site, close to Pisciarelli. They are strongly correlated with ground movements (Fig. 5.12). During the 1982-1984 uplift the NaCl rich hydrothermal component of these waters was diluted by an H_2S and CO_2 rich steam, causing the anion composition of the water to move towards that typical for steam heated waters (Fig. 5.13), similar to those found at Pisciarelli (see Chapter 2, section 2.2.2). Dilution is inferred from peaks in HCO_3/Cl and SO_4/Cl that correspond to peaks in CO_2 and H_2S from the Solfatara fumaroles (Ghiara et al., 1988). Once uplift and seismicity ended, there was an immediate decline in the concentration of HCO_3 and SO_4, and an increase in Cl consistent with a reduction in the steam component. The compositional response to the 1982-1984 uplift appears to be faster than at the Solfatara fumaroles and may be related to increases to permeability, but this cannot be confirmed. Subsequent peaks in temperature in 1989 and 1993 indicate inputs of hot fluids into the shallow subsurface, whilst a peak in both HCO_3/Cl and SO_4/Cl in 1993 was recorded prior to a seismic swarm in 1993, compatible with an input of steam (Valentino and Stanzione, 2004).

Water flux records from liquid dominated hydrothermal activity at the surface and piezometric level measurements across the caldera are not available in the literature. However, Celico et al. (1992a) refer to measurements of the piezometric level in 48 wells between December 1985 and November 1986. The exact locations of the wells are unknown but a continuous increase in the height of the water table in the central caldera, south of the Quarto Plain appears to have occurred.

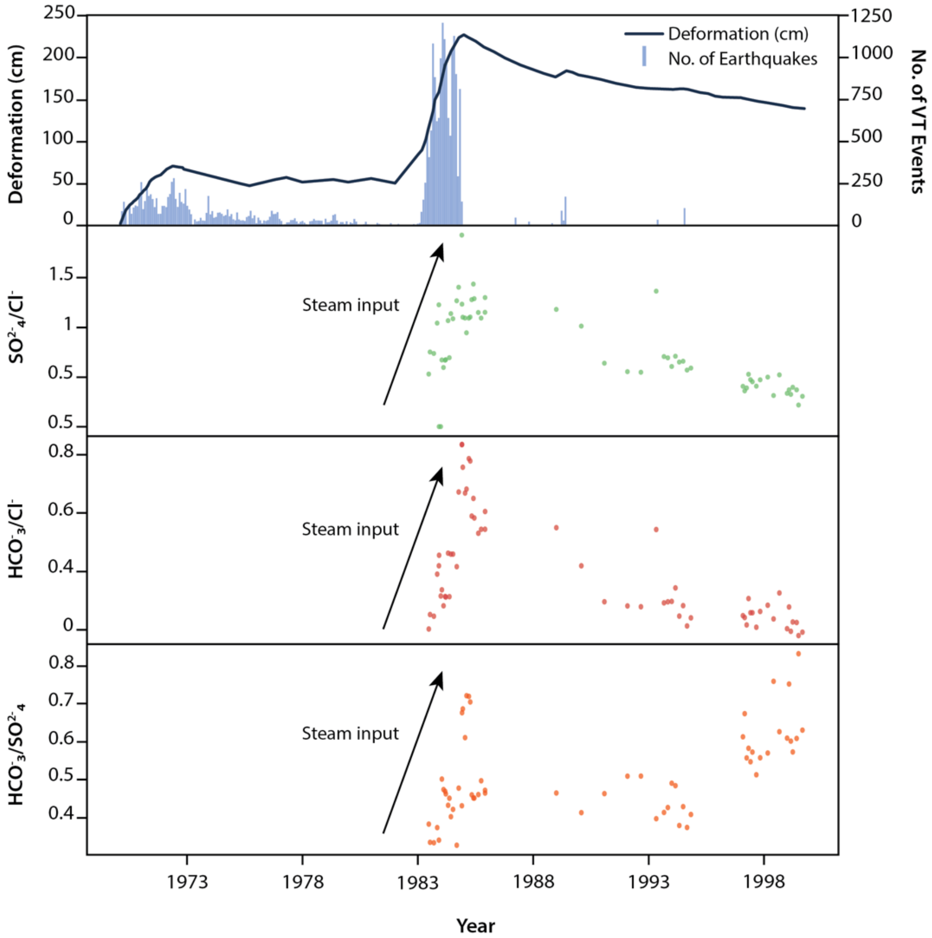


Figure 5.13: Correlation between anion ratios in waters from Hotel Tennis and the 1982-1984 uplift. The increase in HCO_3 and SO_4 relative to Cl is indicative of the dilution of the Na-Cl rich hydrothermal component with magmatic gas-rich steam (Ghiara et al., 1988). Ground level data until May 2000 is from Del Gaudio (2010), data points for May 2000 to July 2016 are cGPS measurements from a supplementary file in Chiodini et al. (2017) and those after July 2016 are cGPS measurements digitised from the INGV-OV Bulletin. Rates of VT seismicity are from the combination of data from seismicity catalogues from the Vesuvius Observatory and in Chiodini et al. (2017), and from Vesuvius Observatory monthly bulletins (ov.ingv.it). Anion concentration data is from Ghiara et al. (1988), Ghiara and Stanzione (1988), Valentino and Stanzione (2003; 2004).

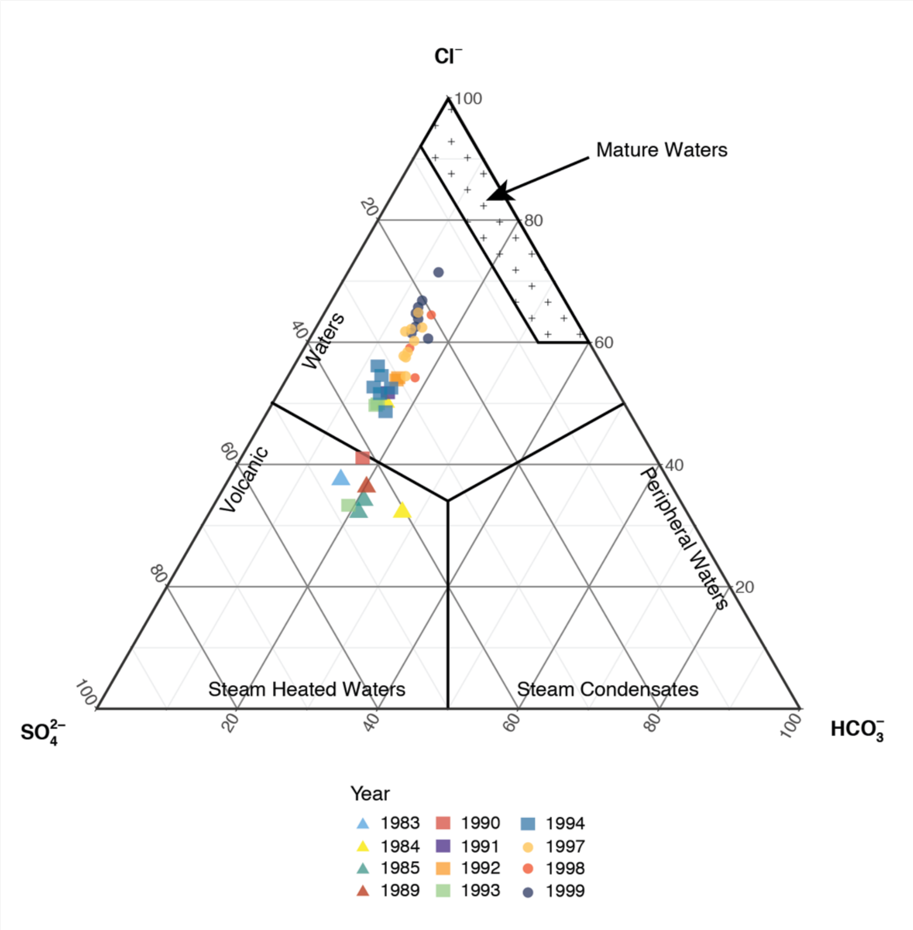


Figure 5.14: Ternary diagram after Giggenbach (1980) showing the relative concentrations of major anions. It can be seen that there is a periodic enrichment of SO_4 and HCO_3, confirming an input of magmatic gas-rich steam in samples during or soon after the 1982-1984 uplift and the mini-uplift in 1990. Data from Ghiara et al. (1988), Ghiara and Stanzione (1988), Valentino and Stanzione (2003; 2004).

5.1.5 Relation between Deformation and Changes in Surface Activity

Comparison of images of the Solfatara crater since the 17th Century indicates that the distribution of the main degassing areas and the location of the Fangaia have remained stable through time (Fig. 5.15). It can therefore be inferred that the locations of the fluid reservoir feeding these features have not been significantly affected by recent ground movements. Prior to the onset of uplift in 1950, small hydrothermal explosions (footprint ~10^1 m), the explosive opening of fumaroles accompanied by ground heating, and an intensification of degassing occurred episodically in the crater (e.g. the opening of the Fumarola Aguilar in 1904 and a vent on 21st April 1921 at the base of Monte Olibano). Opening of fractures of unknown dimensions in the crater floor, most likely related to alteration and fluid movement, were regularly observed and the formation of at least nine *vulcanetti* were recorded between 1874 and 1935 (Signore, 1935). These short-lived features (from months to a few years) were funnel shaped depressions, several metres wide and deep, that appeared explosively and ejected mud spatter several metres into the air. They were produced by the sudden release of pressurised fluids in the shallow subsurface from below a liquid cap and were observed to appear after periods of heavy rainfall or following regional seismicity (e.g. 7th June 1910, Signore, 1935). With the exception of a single unverified event in 1970 (Di Giacomo, 1994), no reports have been located of these features appearing since. If this is the case, it may be speculated that this is due to some combination of a net-increase in permeability related to deformation since 1950, or a lowering of the water table due to uplift. Regardless, it is apparent that activity at Solfatara is dynamic, even during periods of quiescence.

Figure 5.15: Images of the Solfatara crater showing the positions of the main degassing area on the SE crater wall and the Fangaia mud pools towards the centre of the crater. A. 16-17th century engraving in the Civitates Orbis Terrarum by Georg Braun (1541-1622) and Franz Hogenburg (1540-1590). Downloaded from http://archeoflegrei.it/campi-flegrei-civitates-orbis-terrarum. B. Engraving in Vera antichitá di Pozzuoli by G. C. Carpaccio (1607). Image from Sicardi (1956). C. Engraving in the Histoire Naturelle (Anonymous) printed by Bartolomeo Narici in Naples (17th century). Image from Sicardi (1956). D. Oil on canvas titled 'Veduta della Solfatara di Pozzuoli' by Tommaso Ruiz c. 1750. Downloaded from: https://bidtoart.com/en/fine-art/veduta-della-solfatara-di-pozzuoli/1641443. E. Engraving by R. Liberatore (1838-1840). Image from Sicardi (1956). F. Image from 2017 (L. Smale, 2017).

No significant changes in surface activity were located in the literature for the period 1950-1952, although Parascandola (1952) observed a visible increase in degassing at Solfatara and an enlargement of the Fangaia between 1944 to 1952. An increase in degassing and in the extent of the mud pools was also reported during the 1969-1972 uplift (Casertano et al., 1976; Scherillo, 1977), whilst in February 1970 a new vigorously venting fumarole opened in the crater (Puntillo, 1970). Scherillo (1977) also referred to the opening of fractures associated with gas emission outside the Solfatara-Pisciarelli area, the emission of H_2S at Monte Nuovo during the uplift, and an increase in submarine fumarole activity in the West of the Gulf of Pozzuoli between 1970 to 1974 but it has not been possible to verify these observations. Newspapers in 1970 reported increased submarine fumarole activity that was killing fauna, and that fishermen had pulled up burnt nets and found cooked fish. However, these claims were discounted during intensive scientific investigation of the unrest (Versino, 1972; Global Volcanism Program).

As for the previous uplifts, an increase in degassing and expansion of the Fangaia was observed during the 1982-1984 unrest episode (Italiano et al., 1984). Further changes in surface activity included the aforementioned extinction of the Fumarola Circolare, the explosive opening of a fumarole vent on 16-17th November 1984 and the opening of a NE-SW trending extensional fracture across the Fangaia. This feature was c. 100 m in length, along which a new mud pool opened that grew to 12.5 m in length with an average width of 0.5 meters (Italiano et al., 1984; Bianchi et al., 1990). The fracture was subsequently infilled and the pool dried up. Divers also documented photographically an increase in submarine fumarolic activity offshore of Monte Nuovo in 1984 (Global Volcanism Program, 2013).

No records have been published to suggest that significant changes in the distribution of activity or the opening of new features have occurred at the Solfatara crater since the onset of uplift in 2004. Two minor ground collapses occurred in 2014 (adjacent to the northern wall of the Solfatara crater) and in September 2017 (in the area of the Fangaia); neither can be directly attributed to the uplift, because ground collapses in areas of such extensive alteration are common due to the low mechanical strength of the rock. However, an intensification of activity at Pisciarelli has been related by Chiodini et al., (2011) to an increased fluid flux along the fault system that channels fluids to the surface at this location, as a direct result of the uplift and associated seismicity (Fig. 5.16). The activity has included small hydrothermal explosions, the formation and extension of bubbling pools, expansion of the fumarolic area and the opening of a new vigorously venting fumarole vent in 2009 (Chiodini et al., 2011; 2015a; Vilardo et al., 2015; INGV-OV, 2019). No reports of similar increases in activity at this location during previous uplifts or significant changes in surface activity elsewhere in the caldera since 2004 been located in the literature.

Figure 5.16: Temperatures at Pisciarelli fumarole and hydrothermal activity since January 1999 (from INV-OV, 2019). The white circles are discrete temperature measurements and the red line refers to the daily mean for continuous measurements. The inset is the vent that opened in 2009.

5.2 Discussion

Each episode of uplift since 1950 and the subsidence between 1984-2004 is consistent with the deformation of an elastic-brittle crust in response to a pressure change in a discrete source (Woo and Kilburn, 2010) that is most likely located in the shallow crust. A well-established feature of the deformation is that throughout the period of instrumental monitoring since 1905, both during quiescence before 1950 and after the onset of the unrest sequence, the geometry of caldera-wide deformation has been constant. A recent reconstruction of ground movements prior to the Monte Nuovo eruption in 1538 suggests that the shape of deformation has been maintained since at least the 15th Century (Di Vito et al., 2016). The constancy in the shape of the deformation and between subsidence and uplift phases is not exceptional to Campi Flegrei and has been observed at other calderas, such as Yellowstone in the United

States (Dzurisin et al., 1994). Proposed explanations for the similarity in the ground movements at Campi Flegrei include: (i) all ground movements are controlled by volume changes in one or more permanent strain sources (e.g. Amoruso et al., 2017), (ii) ground movements are controlled by different strain sources that are always of similar shape and location in the crust (e.g. Woo and Kilburn, 2010), and (iii) the geometry is constrained by structural discontinuities. In the latter case, published interpretations suggest that the geometry is controlled either by a confining effect of the caldera ring faults (e.g. De Natale and Pingue, 1993; Beaudecel et al., 2004; Folch and Gottsman, 2006), or by the accommodation of ground movements along faults that define a resurgent block (e.g. Orsi et al., 1999b). A single source controlling all ground movements is unlikely given the differing characteristics of deformation through time. It is also difficult to reconcile a common geometry for these movements if they result from different strain sources. The preferred explanation here is for a structural control and that the location of the maximum field of displacement is determined by the high density of faults located in the centre of the caldera in the region of Pozzuoli. In addition to the constant geometry, such a control can also account for the elongation of the deforming area parallel to the La Starza marine terrace fault system, and the greater component of horizontal deformation in the E-W direction relative to the N-S observed during both the 1969-1972 and 1982-1984 uplifts, without invoking an additional process such as regional tectonic ESE-WSW extension (e.g. Woo and Kilburn, 2010). These area containing these faults broadly corresponds with the region of the crust below Pozzuoli expected to contain hydrothermal reservoirs based on the location of the seismic anomaly inferred by Aster and Meyer (1988) as a pervasively fractured volume saturated in fluids (Chapter 2). The implications of this are that regardless of the nature of the source, the geometry of caldera-wide deformation will always be the same, and that structures that act as fluid flow paths in the hydrothermal system are likely to be involved in accommodating ground movements through time. It may also be reasonably inferred that movements along these structures may have an intrinsic role in maintaining permeability in this region of the crust.

The similar rates and durations of displacement during rapid uplifts between 1950-1984 imply a common causative process that cumulatively resulted in a permanent displacement of the central caldera of 2.9 m. Both the 1969-1972 and 1982-1984 episodes were of comparable magnitude and the rates of displacement during each were similar. However, the deformation that followed each was distinctly different. After the 1969-1972 uplift, subsidence was minor (c. 0.2, 11% of the preceding uplift) and lasted three years, after which the ground level was essentially stable until the start of the next uplift. Throughout this period VT seismicity persisted and is regarded as an indicator of an elastic relaxation of the crust by slip along small (< 0.1 km long) faults as it adjusted to a new lithostatic equilibrium (e.g. Orsi et al., 1999b). In contrast, subsidence after 1982-1984 was greater (c. 0.9 m, 50% of the preceding uplift), longer lasting (1985 to 2004), and largely aseismic. It is generally attributed to a loss of pore pressure in the hydrothermal system (Chapter 3). The emergence of this signal suggests a critical change in the crust occurred during the 1982-1984 uplift. The most notable difference in monitoring parameters between the two uplifts is the increase in the number and maximum magnitude of VT events over the successive episodes. As such it can be concluded that the amount of brittle deformation, or crustal damage, during the 1982-1984 uplift was greater than the preceding episode and that slip also occurred along longer structures. According to Kilburn et al. (2017) conditions were approaching bulk failure in the crust. As such the amount of fracturing and faulting during the 1982-1984 uplift is expected to be significantly greater than during the previous uplift.

Evaluation of the impact of unrest on the hydrothermal system is limited by the comparatively shorter records of geochemical sampling and reporting relative to those for ground level and seismicity. There is no evidence to suggest any major hydrological organisation in the shallow hydrothermal system from either thermal water compositions or the distribution of activity over the 1969-1972 and 1982-1984 uplift episodes, although permeability changes at shallow depths did cause minor compositional changes in thermal waters at some locations. There are also no records in the available literature that would suggest that there was any notable increase in the frequency of the opening of fumarole vents or hydrothermal explosions that may suggest significant pressure changes in the near surface during the uplifts. Observations of increased degassing and an expansion of the Fangaia mud pools during both the 1969-1972 and 1982-1984 uplifts could suggest that there was an increase in the flux of gas through the Solfatara-Pisciarelli DDS during both episodes, which in the absence of any significant meteorological events, can most simply be attributed to an increased supply of gases to the shallow reservoirs feeding fumaroles and condensates reaching the surface at the Fangaia (Chapter 2). This is reasonable given that thermal water compositions at Terme Puteolane, at the margin of the outflow of the Solfatara plume, were enriched in CO_2 during these periods, and in 1982-1984 there was also a recorded increase in the concentration of magmatic gases (e.g. CO_2,

N_2) from Solfatara fumaroles (Chiodini et al., 2003), and dilution of thermal waters marginal to the DDS by steam (e.g. at Hotel Tennis).

The longest continuous geochemical trends are available for gases from the Bocca Grande vent at Solfatara from 1983 to present (2019). As a single sampling point the trends cannot be considered representative of the whole hydrothermal system, however they do act as indicators of the composition of fluids being transported through the Solfatara-Pisciarelli DDS, which is connected to the main hydrothermal reservoirs below Pozzuoli (Chapter 2). Between 1983-2000, peaks in CO_2 and N_2 of Solfatara fumaroles were superimposed on an overall declining trend and occurred in response to geophysical changes in the crust during the 1982-1984 uplift and mini-uplifts in 1989 and 1994 (e.g. Chiodini et al., 2003). The temporal relationship between geochemical trends in fumarole gases and the geodetic signal then changed in 2000. At this time the progressive enrichment of gases in CO_2 that continues to present (2019) began but there was no increase in the ground level recorded in 2004. This suggests an increase in magmatic gases reaching the surface but contrary to the period between 1983-2000, this increase in CO_2 has not been accompanied by an enrichment in N_2, indicating that either this gas has been exhausted in the source of magmatic gases (e.g. Chiodini et al., 2015a), or that the source of CO_2 has changed (e.g. Moretti et al., 2017). The relative timing of the compositional changes in fumarole gases and the change in the ground level suggests that the uplift since 2004 is a response of the crust to the processes causing the compositional change and as such the ground movement must result from a different mechanism to the preceding episodes. This supported by the comparatively slow rate of deformation, the reduction in the seismogenic volume relative to earlier episodes and the clustering of seismicity at shallower depths. In order to achieve the approximate symmetry between subsidence after 1984 and uplift since 2004, the process controlling the uplift must be acting on the crust at the inverse rate of that responsible for the preceding subsidence.

3.3 Summary

The long-term deformation profile at Campi Flegrei and the relation between deformation and other monitoring parameters since 1950 suggest that the controls on uplifts have changed over time. The similarities in the characteristics of rapid uplifts between 1950-1984 is consistent with a common causative process. However, the emergence of the slow ground oscillation post-1984 is indicative that a change in the conditions in the crust occurred during the third uplift episode in 1982-1984. The timing of this change follows a period of inelastic deformation greater than observed during earlier uplift episodes (based on rates of VT seismicity) and the approach of stress conditions in the crust towards those favourable to bulk failure and widespread fracturing. The characteristics of uplift since 2004 suggest that the mechanism of uplift is not the same as for previous episodes. This is supported by a change in the behaviour of compositional trends in fumarole geochemistry that preceded the onset of this ground movement and relatively low rates of seismicity accompanying uplift. Throughout the unrest sequence since 1950 the field of maximum deformation appears to have been centred on Pozzuoli, implying the involvement of the region of the crust containing the hydrothermal circulation and that the same region of the crust has accommodated ground movements through time. In order to be successful, a new model of unrest must be consistent with the observations of the behaviour of the caldera and hydrothermal features summarised in Table 5.1. Critically it must be able to account for the following key observations;

- the accumulation of net-uplift between 1950-1984 and the emergence of the slow ground oscillation after the 1982-1984 uplift,
- the rate of uplift since 2004, which is essentially the inverse of the preceding subsidence,
- the change in the behaviour of the CO_2/H_2O ratio from an oscillatory to continuously increasing trend in 2000 without a concurrent enrichment in N_2.

Table 5.1: Summary of key observations since 1950

	Period	Observation
Deformation	1905-Present	Constant geometry of deformation
	1950-1984	Cumulative net-uplift of 2.9 m
	1985-2003	Long-term subsidence above background rates and recovery of 0.9 m of deformation
	2004-Present	Reversal of subsidence trend and slow uplift
Seismicity	1969-1984	Progressive increase in the rates and maximum magnitudes of VT seismicity over successive uplifts
	1972-1982	Decelerating rates of seismicity following uplift. End of seismicity below 2 km.
	1984	Abrupt cessation of seismicity at the end of uplift
	1984	Stress conditions approach those for bulk failure of the crust
	1970-1984	Seismicity during uplift distributed around margins of caldera collapse and along offshore faults between the surface and depths up to c. 7 km
	2004-Present	Seismicity during uplift confined between the surface at c. 4 km below Pozzuoli
	2004-Present	Low seismic energy release relative to previous uplift episodes
Geochemistry	1969-1972	Possible input of magmatic gas-rich fluids into the hydrothermal system
	1982-1985	Input of magmatic CO_2-N_2 rich fluids into hydrothermal system during uplift and seismicity causing more oxidising conditions. Increase in magmatic-gas rich steam from fumaroles and in thermal waters closest to the Solfatara-Pisciarelli DDS.
	1985-2000	Progressive decrease in the magmatic gas-rich fluid component discharging at fumaroles and increase in reduced species. Magmatic CO_2-N_2 rich fluid inputs in 1990 and 1994 following mini-uplifts and seismic swarms. Increase in H_2O, indicative of boiling.
	1982-2000	Geochemical changes lag changes in ground level and seismic swarms
	2000	Change in magmatic-gas rich fluid component from CO_2-N_2 rich to CO_2 rich, N_2 depleted.
	2000-Present	Monotonic increase in CO_2 with opposite trend observed for CO2. Increase in He, stable N_2 concentrations.
	2000-Present	Higher frequency of magmatic gas-rich fluid injections into hydrothermal system relative to 1983-2000
Fluid Flux	1985-1986	Increase in water table height in the central caldera
	2003-Present	Increase in CO_2 flux from the Solfatara-Pisciarelli DDS after 2003
Surface Activity	1944-1952	Unconfirmed expansion of Fangaia mud pools
	1969-1972	Expansion Fangaia mud pools, and intensification of fumarolic activity at Solfatara.
	1982-1984	Expansion Fangaia mud pools, and intensification of fumarolic activity at Solfatara. Possible intensification of activity at offshore fumaroles
	2000-Present	Intensification of activity at Pisciarelli

Chapter 6

A Crustal Damage Model for Unrest

Existing models of post-1984 ground movements discussed in Chapter 3 have shown that subsidence from 1985 to 2004, and uplift since then are compatible with pore pressure changes in the shallow crust (<3 km depth). However, a common limitation between them is that they are unable to account for either the timing of the emergence of this signal, or why a decadal ground oscillation was observed after rapid uplift in 1982-1984 but not following earlier episodes in 1950-1952 and 1969-1972. In each case these models are constrained by monitoring parameters since 1982, assume that uplifts are independent events requiring the transport of pressurised fluids originating in the magmatic system to shallower depths, and that subsidence after 1984 represents a return to equilibrium conditions in the crust.

In this chapter a new conceptual model for post-1984 ground movements is proposed that accounts for the change in the characteristics of deformation after 1984 for the first time. Contrary to existing models, it considers the full deformation profile since 1950 to represent a single, long-term evolutionary sequence. Ground movements after 1984 are then attributed to a loss and subsequent recovery of pore pressure in the hydrothermal reservoirs below Pozzuoli, triggered by the exceedance of a critical value of permeability resulting from the repeated stretching of the crust over successive uplifts between 1950-1984. The primary implications of the model are that post-1984 ground movements can be explained by considering long-term changes in conditions in the crust and that the subsidence may represent a departure from, rather than return to, equilibrium conditions.

6.1 Model Starting Assumptions

The lack of direct observations of the subsurface in the main deforming area of the caldera and the absence of diagnostic indicators of the processes controlling ground movements in monitoring parameters, means that any conceptual model must make fundamental assumptions. The primary assumptions made here are outlined in this section and are principally based on the results of the reviews in Chapters 2 and 5.

6.1.1 Structure of the Hydrothermal System Below Pozzuoli

Deformation throughout the unrest sequence has been concentrated on Pozzuoli (Chapter 5). As such, processes controlling ground movements are expected to operate in the subsurface below this area in the caldera. The assumed model of the crust is based on the results of the literature review in Chapter 2 and assumes the presence of a hydrothermal basin in the shallow crust the position of which is stable through time (Chapter 2, section 2.2.2). The essential features of the hydrothermal system (Fig. 6.1) are; a CO_2-rich hydrothermal reservoir zone located between 1.5-3 km depth, a low permeability caprock formed by hydrothermal alteration that limits upflow, and the Solfatara-Pisciarelli Diffuse Degassing Structure (DDS), which acts as a conduit for fluids to the surface. Magmatic fluids are transported into the hydrothermal system from an underlying zone of supercritical gas and brine at 3-4 km depth across a transition zone of rapid permeability loss that defines the limit of meteoric circulation. Here they undergo decompression boiling and mix with meteoric water, flashing it to vapour. Transport of fluids out of the reservoir zone is via lateral outflow and degassing through the Solfatara-Pisciarelli DDS.

The location of the reservoir zone is based on the results of geophysical imaging of the caldera (Aster and Meyer, 1988; De Lorenzo et al., 2001; Vanorio et al., 2005; Battaglia et al., 2008; De Siena et al., 2010; Capuano et al., 2013; Caló and Tramelli, 2018) and the clustering of hydrothermal surface activity in this region of the caldera (Chapter 2, section 2.2.2). The caprock depth, position of the Solfatara-Pisciarelli DDS and the presence of a storage zone of overpressured magmatic fluids are also assumed from geophysical imaging studies (e.g. Vanorio et al., 2005; Zollo et al., 2008; De Siena et al. 2010; Troiano et al., 2014; Vanorio and Kanitpanyacharoen, 2015). Whilst the caldera collapse is piecemeal (Scandone et al., 1991; Bruno et al., 2004; Capuano et al., 2013; Steinmann et al., 2018), so that observations from boreholes drilled elsewhere in the caldera cannot be directly applied, the assumed depths of the reservoir zone overlap that intersected by geothermal exploration boreholes at Mofete in the west of the caldera (Rosi and Sbrana, 1987). The presence of CO_2-rich hydrothermal reservoirs and the origin of fluids in crust below Pozzuoli is based on the prevailing geochemical model of the Solfatara feeding system by Caliro et al. (2007; 2014).

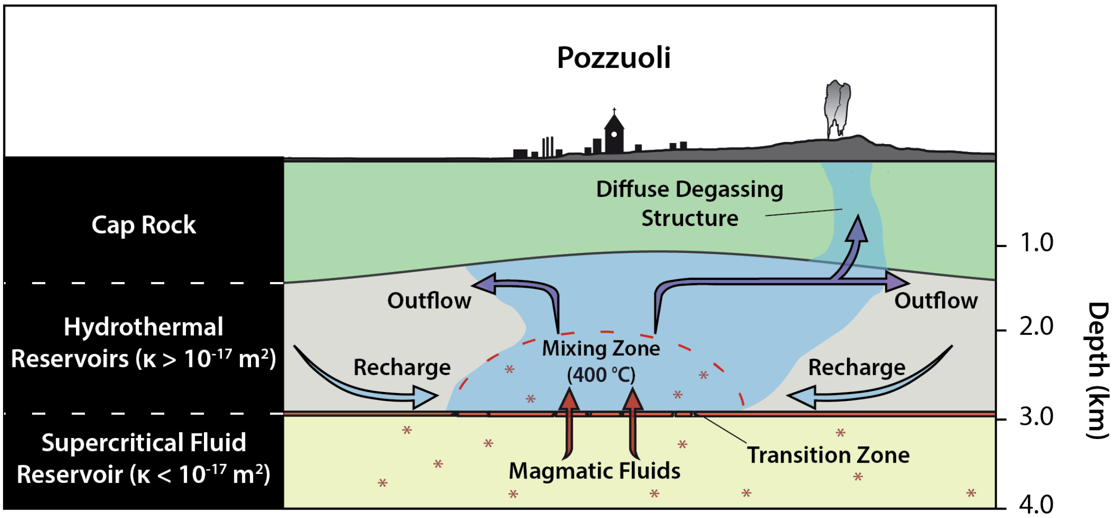


Figure 6.1: Schematic of the primary features of the hydrothermal system below Pozzuoli in the model assumed here. The mean permeability (κ) of the supercritical fluid reservoir is assumed to be $\leq 10^{-17}$ m^2, which is sufficiently low for the accumulation of overpressured fluids as inferred by Zollo et al. (2008) to be present at 3-4 km depth. The bulk permeability in the hydrothermal reservoir is expected to be $\geq 10^{-17}$ m^2 and thus sufficiently high for advection at hydrostatic pore pressures.

6.1.2 Nature of the Source of Ground Movements Since 1950

The second set of assumptions that must be made are related to the nature of the pressurised fluids driving ground movements over time. The similarity in the characteristics of uplifts between 1950-1984 (Chapter 5, Fig. 6.2) implies a common source mechanism, whilst the net-uplift over the successive episodes requires the pressurisation of a permanent strain source during each of the three uplifts. This constraint may be satisfied by either successive injections of magmatic fluid below a hydrological barrier, requiring a period of accelerated degassing of the primary magma reservoir or, most simply, by the repeated intrusion of magma at shallow depths. Intrusions are the preferred mechanism here because related deformation is permanent once they solidify. The involvement of magma is also supported by the observed enrichment in gases considered to have a magmatic origin (e.g. CO_2 and N_2) and depletion in reduced gases such as CH_4 and H_2S at Solfatara fumaroles during the 1982-1984 uplift (Caliro et al., 2007; 2014; Chiodini et al., 2015a). Total uplift at the end of the 1982-1984 unrest can then be considered as the sum of the movements caused by magma intrusions and pore pressure in the crust. Post-1984 subsidence therefore must reflect changes in either or both of these components. Whichever the controlling mechanism, it must be able to produce the approximately symmetric pattern of post-1984 ground movements, act over the observed timescales of both phases of deformation and cannot have operated to a significant extent after the major uplifts in 1950-1952 and 1969-1972.

The intrusions are assumed here to be sills as this geometry can account for the observed deformation at overpressures smaller than the tensile strength of the crust (<10 MPa) unlike other source shapes (Woo and Kilburn, 2010). Geodetic inversions of the 1982-1984 uplift that assume a sill geometry suggest intrusion thicknesses of ~1-10 m and source depths located between 2.6-3.1 km (e.g. Battaglia et al., 2006b; Amoruso et al., 2008; Woo and Kilburn, 2010). Temperatures at these depths are expected to be c. 400 °C (Rosi and Sbrana, 1987). For post-1984 ground movements to have been controlled by changes in intrusions, a volume loss in a magma body, followed by a further period of magma intrusion would be required. Mechanisms for volume loss include magma degassing, lateral migration of the magma and contraction on cooling. As such, a necessary condition for magma to control subsidence is that it remains mostly fluid for the duration (i.e. from 1985 to 2004). The maximum solidification time can be approximated by estimating the time for the solidus to migrate from the margins of an intrusion to the centre assuming cooling by conduction only. Jaeger (1968) modelled conductive cooling of a sill as an infinitely wide rectangular sheet that cools from the upper and lower margins using the governing equation:

$$\frac{(T-T_1)}{(T_0-T_1)} = \left(\frac{1}{2}\right)\left\{\frac{erf(\xi+1)}{2\tau^{\frac{1}{2}}} - \frac{erf(\xi-1)}{2\tau^{\frac{1}{2}}}\right\} \quad [1]$$

where T, T_0 and T_1 are the temperature at the distance in the sill from its margin (x), the initial temperature of the magma and the temperature of the host rock, $\xi = \frac{x}{a}$, $\tau = \frac{\kappa t}{a^2}$, a is the half thickness of the sill, κ is the thermal diffusivity of the magma, τ is a Fourier number and dimensionless, and t is time. The term erf denotes the error function, which is a form of infinite series. Jaeger (1968) presented the results graphically and tabulated the ratio $\frac{(T-T_1)}{(T_0-T_1)}$ against corresponding values of τ. Magmas at Campi Flegrei are typically trachytic in composition (Chapter 2), a representative liquidus temperature for which is 1040 °C, whilst the solidus temperature is typically 200 °C less than the liquidus (C. Kilburn, pers. comm). Following Jaeger, the ratio $\frac{(T-T_1)}{(T_0-T_1)}$ for a trachytic sill of maximum thickness of 10 m intruded into a host rock at 400

°C would be 0.69. The corresponding value of τ is 0.5, which, assuming a thermal diffusivity of magma of 4 x 10^{-7} m^2 s^{-1} (Murase and McBirney, 1973), gives a solidification time, t, of a year. As such, a volume loss in the sill as the control on subsidence is unlikely. Furthermore, if magmatic processes controlled post-1984 ground movements then the same processes would be expected to have operated after the 1950-1952 and 1969-1972 uplifts, but no comparable ground movements were observed after either episode.

The aseismic subsidence after the third rapid uplift episode, suggests the operation of a new process controlling ground movements that had not occurred previously. Similarly, the characteristics of uplift since 2004 are distinctly different to the earlier episodes, including; the slow rate of displacement, a slower seismic energy release, a reduction in the size of the seismogenic volume, and the enrichment in CO_2 in Solfatara fumarole gases that preceded, rather than lagged the geodetic signal (Chapter 5). These ground movements are compatible with a pore pressure loss and subsequent recovery related to changes in fluid flow. The emergence of the slow ground oscillation is therefore considered here to represent a transition from a magmatic to pore pressure control on ground movements after 1984. This is in agreement with interpretations of compositional trends in Solfatara fumarole gases (Chiodini et al., 2015a; Moretti et al., 2017). The timing of the transition implies that it is an effect of the third sill intrusion in 1982-1984.

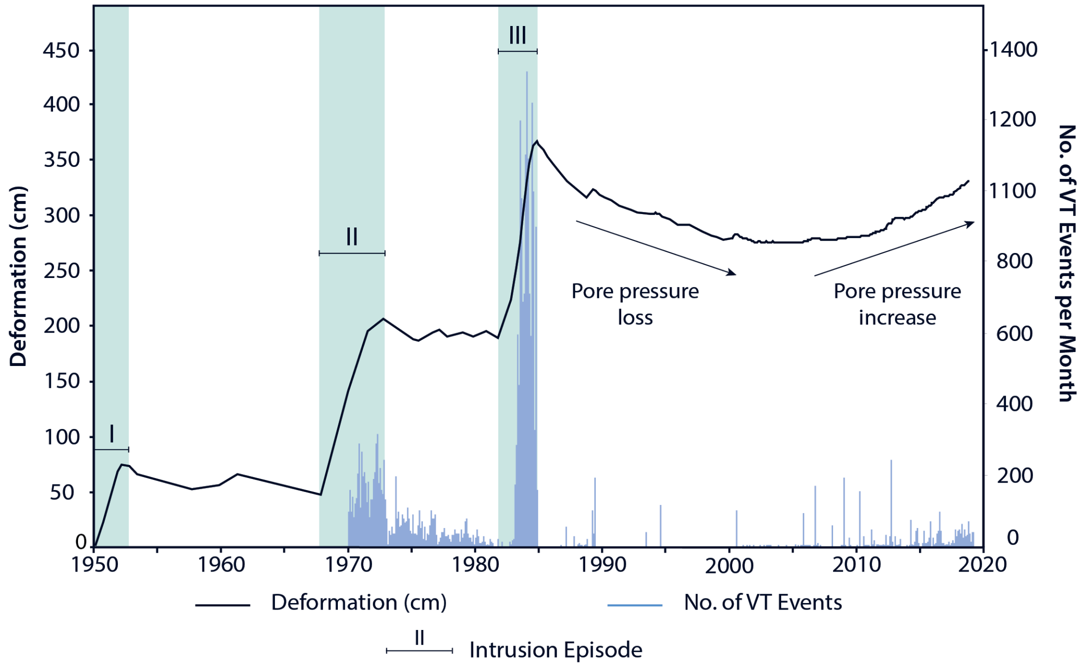


Figure 6.2: The deformation profile since 1950 and the assumed controls on ground movements. I, II and III indicate periods of sill intrusion in 1950-1952, 1969-1972 and 1982-1984. Post-1984 subsidence and uplift are interpreted as a depressurisation and re-pressurisation of the hydrothermal system respectively. Ground level data until May 2000 is from Del Gaudio (2010), data points for May 2000 to July 2016 are cGPS measurements from a supplementary file in Chiodini et al. (2017) and those after July 2016 are cGPS measurements digitised from the INGV-OV Bulletin. Rates of Volcano-Tectonic (VT) seismicity are from the combination of data from seismicity catalogues from the Vesuvius Observatory and in Chiodini et al. (2017), and from Vesuvius Observatory monthly bulletins (ov.ingv.it).

6.1.3 Fracture Controlled Permeability

The primary control on the pore pressure distribution in the crust is permeability (κ). This parameter regulates fluid flow, transport of heat and mass into, and out of, a hydrothermal system, and therefore pore pressures (Elder 1981; Hurwitz et al., 2007; Todesco et al., 2010). It determines the generation of superhydrostatic overpressures ($\kappa \leq 10^{-17}$ m^2), the transition from conduction to advection dominated heat transport ($\kappa \geq 10^{-16}$ m^2), the dimensions of the volume through which fluids circulate, and the rate at which a pore pressure disturbance propagates (Hayba and Ingebritsen, 1997; Hutnak et al., 2009; Todesco et al., 2010). The constitutive equation that describes fluid flow is Darcy's Law (Table 6.1), the form of which depends on the fluid phases present (Darcy, 1856; Ingebritsen et al., 2010). Each phase has a different relative permeability depending on its density and kinematic viscosity (Norton and Knight, 1977; Elder et al., 1981; Ingebritsen et al., 2010). For example, the permeability of gas in volcanic materials is 2-5 times greater than that for liquid water (Heap et al., 2018). Where gas and liquid co-exist, an interference effect occurs, increasing the resistance to flow. This is the result of gas preferentially occupying larger void spaces and fractures, whilst liquid is held in smaller pores by capillary pressure and surface effects (Elder, 1981; Todesco, 2008).

Table 6.1: The constitutive equations for single- and two-phase fluid flow

Form	Equation	Parameter
Single-Phase Flow	$q = -\left(\frac{\kappa\rho g}{\varepsilon\mu}\right)\frac{dp}{dl}$	q is the flux (ms $^{-1}$), κ is the intrinsic permeability of a porous medium (m^2), ρ is the fluid density (g cm^3), g is acceleration due to gravity (m s^{-2}), ε is the porosity of the host rock, μ is the fluid viscosity (Pa s) and $\frac{dp}{dl}$ is the hydraulic gradient (Pa m^{-1}).
Two-Phase Flow (gas and liquid)	$q_\beta = -\frac{k_{r\beta}k}{\mu_\beta}\left(\frac{\partial P}{\partial z} + \rho_\beta g\right)$	q_β is the volumetric flux per unit area for phase β, k_r is a dimensionless value describing the relative permeability of the phase, μ_β is the dynamic fluid viscosity for phase β (Pa s), P is pressure (Pa), z is depth (km), and ρ_β is the phase density (g cm^3).

The intrinsic permeability of a caldera fill is dependent on the lithological units present. Generally, it is low as tuffs and clays have high porosity but poor connectivity between pore spaces, whilst lavas and intrusions can act as hydrological barriers (Cathles et al., 1997; Manning and Ingebritsen, 1999; Jasim et al., 2015; 2018 and references therein). Faults can also disrupt lateral flow as a result of the low permeability of fault cores (Jasim et al., 2015). The bulk permeability of a hydrothermal system is controlled by the presence of fractures and their characteristics, i.e. density distribution, aperture, tortuosity, length, degree of linkage and infilling (Hurwitz et al., 2007). Their fundamental role in fluid transport is well established from laboratory experiments and field studies of exhumed systems (e.g. Norton and Knight, 1977; Ingebritsen and Manning, 2010; Rowland and Simmons, 2012; Cox, 2016; Farquahson et al., 2017).

Permeability is a dynamic parameter that is continually modified by physical and chemical processes. Processes that act to reduce permeability through closure of fractures operate over a range of timescales and include: compaction (10^1-10^3 years), mineral alteration and precipitation (10^0-10^3), and changes in the stress field (instantaneous, Browne, 1978; Fournier et al., 1993; Reed, 1997; Curewitz and Karsson, 1997; Manning and Ingebritsen, 1999, Cox, 2005; Preisig et al., 2015). Of particular importance is fracture sealing by precipitation of SiO_2. At the base of a hydrothermal system (i.e. the limit of meteoric circulation) temperatures are typically ~350-420 °C, where quartz enters the field of retrograde solubility and precipitates out (Fournier, 1985; Saishu et al., 2014). At shallower depths and cooler temperatures, it precipitates from supersaturated fluids during boiling or cooling (Grindley and Browne, 1976; Lowell et al., 1993). Calcite precipitation is also important in regions where CO_2-rich fluids circulate (Chiodini et al. 2015b).

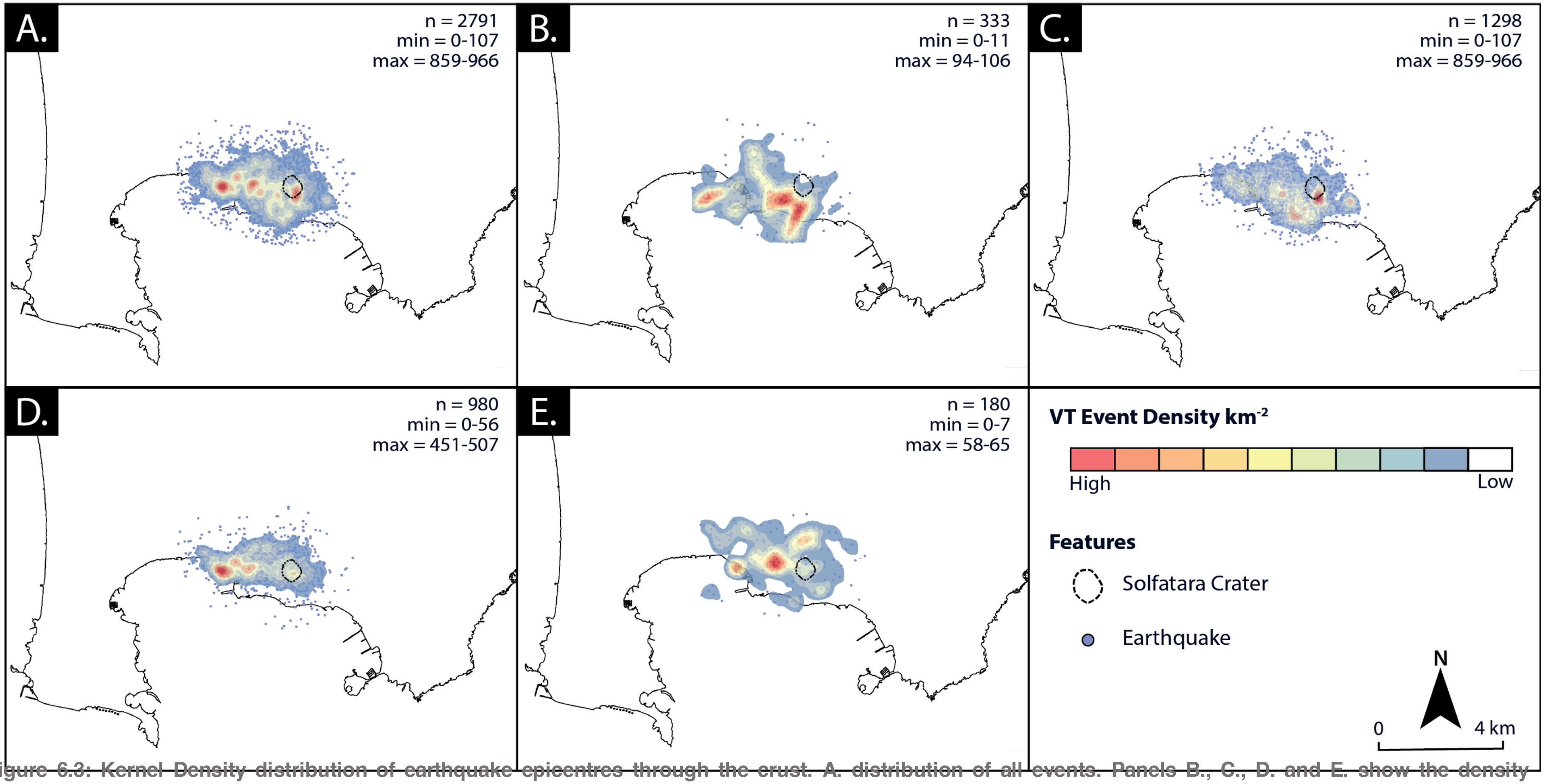
In order to maintain hydrothermal circulation through time, episodic permeability creation through fracture reactivation, nucleation, propagation, and coalescence must occur. It has been shown that fracturing can increase permeability by 1-4 orders of magnitude (Farquharson et al., 2016; Heap and Kennedy, 2016), which may sufficiently change flow, and thus pore pressure distribution, for deformation to be observable at the surface, depending on the size of the affected volume (Hurwitz et al., 2007; Todesco et al., 2010). Mechanisms of fracturing include; hydrofracturing by overpressured fluids (e.g. Cox, 2005; 2016), co-seismic fracture (Curewitz and Karsson, 1997), hydrofracturing due to injection of lithostatically pressured magmatic fluids (Weiss, 2015), thermal contraction and expansion of the host rock (Cathles et al., 1997), and brittle deformation induced by magma intrusion (e.g. Chang et al., 2007).

In line with the general understanding of fluid transport in hydrothermal systems and in the absence of direct observations of the hydrothermal system below Pozzuoli, it is assumed that permeability and therefore the pore pressure distribution, is controlled by the presence of fractures. This assumption is supported by the dependence of the location of hydrothermal reservoirs on the presence of fracture zones elsewhere in the caldera at Mofete (Chapter 2 and references therein), the correspondence of the hydrothermal circulation below Pozzuoli with the highest density of faults in the caldera, and the fracture-dependent location of outflow at the surface (AGIP; De Siena et al., 2010; Acocella et al., 2010 and Vitale and Isaia, 2014). Furthermore, geochemical modelling of Solfatara fumarole gases suggests that fluids circulate at temperatures increasing from c. 220 °C to c. 400 °C with depth (Caliro et al., 2007). Based on these temperatures the host rock would be expected to have undergone calc-aluminium silicate to thermometamorphic alteration, and therefore be sufficiently competent and brittle to support flow through dynamically maintained fracture networks (Browne, 1978; Stimac et al., 2015). In order for convection to be maintained, permeability must be in the order of 10^{-16} m^2 or more (Fig. 6.1). The pore pressure within the reservoir zone is expected to be regulated by the rate at which heat and mass is transferred by fluids from the underlying magmatic system and transported out of the hydrothermal reservoir. Increased rates of magmatic fluid inflow would therefore be favoured by fracturing across the base of the hydrothermal system, and increased rates of outflow by fracturing within the system itself. Pore pressures, in turn, would be expected to increase with faster rates of inflow and to decrease with faster rates of outflow. As a result, new fracturing below the hydrothermal reservoir may promote uplift and fracturing within the reservoir would favour subsidence.

6.1.4 Bulk-Permeability Increase in the Shallow Crust in 1984

The final assumption is that to trigger the post-1984 deformation sequence a bulk permeability change is required during the 1982-1984 uplift in a region of the crust below Pozzuoli that results in a redistribution of pore pressure. Adopting the Kilburn et al. (2017) model of progressive deformation of the crust at Campi Flegrei (Chapter 5, section 5.1.2), it is expected that (i) the inelastic component of deformation (i.e. fracturing and faulting) increased exponentially with uplift between 1969-1984, and (ii) conditions in the crust at the end of 1984 were favourable for bulk failure and the onset of widespread fracturing large enough to change the bulk permeability of the crust being deformed. To evaluate where in the crust permeability changes during the 1982-1984 uplift were most likely occurring, the distribution of Volcano-Tectonic (VT) seismicity in the primary seismogenic volume during this period was considered using a catalogue of 3708 located events between January 1983 and September 1984 from the Vesuvius Observatory.

The epicentral distribution shows that the area over which brittle deformation was occurring was larger than but encompassed the area of Pozzuoli below which hydrothermal reservoirs are expected (Fig. 6.3). The density distribution highlights the presence of two "hotspots" of seismicity, suggesting a concentration of activity on particular structures. These hotspots have also been identified by De Siena et al. (2017) and Castaldo et al. (2019). Both are located within regions of high strain identified by Acocella et al. (2010) from a structural analysis of surface fractures. The greatest density of events occurred in a small (c. 1 km^2), approximately circular area, to the SE of the Solfatara crater, that broadly corresponds with the fractures that connect the main hydrothermal reservoir to the surface at Solfatara-Pisciarelli. The second maximum is located in the western portion of the main cluster, close to low temperature hydrothermal discharges. The highest magnitude events ($M_s \geq 3.1$), associated with the greatest amount of slip (on faults up to 1 km long) were located within 2 km of the Solfatara crater (Fig. 6.4). The depth distribution of seismicity (Fig. 6.5 and Fig. 6.6) confirms that seismicity was concentrated within the volume containing the seismic anomaly inferred by Aster and Meyer (1988) as a zone of hydrothermal reservoirs. The greatest number of events and largest magnitude events occurred between 1.5-3 km depth, which corresponds to the region below the caprock and the hydrothermal reservoir zone in the assumed model of the crust.



~~Figure 6.3: Kernel Density distribution of earthquake epicentres through the crust. A. distribution of all events. Panels B., C., D. and E. show the density~~ distribution of epicentres located between 0 to 1 km, 1.1 to 2 km, 2.1 to 3 km and more than 3 km depth respectively. Two main hot spots of seismicity are apparent at Solfatara and in the west of the main cluster. Earthquakes in the west tend to be distributed deeper than those at Solfatara.

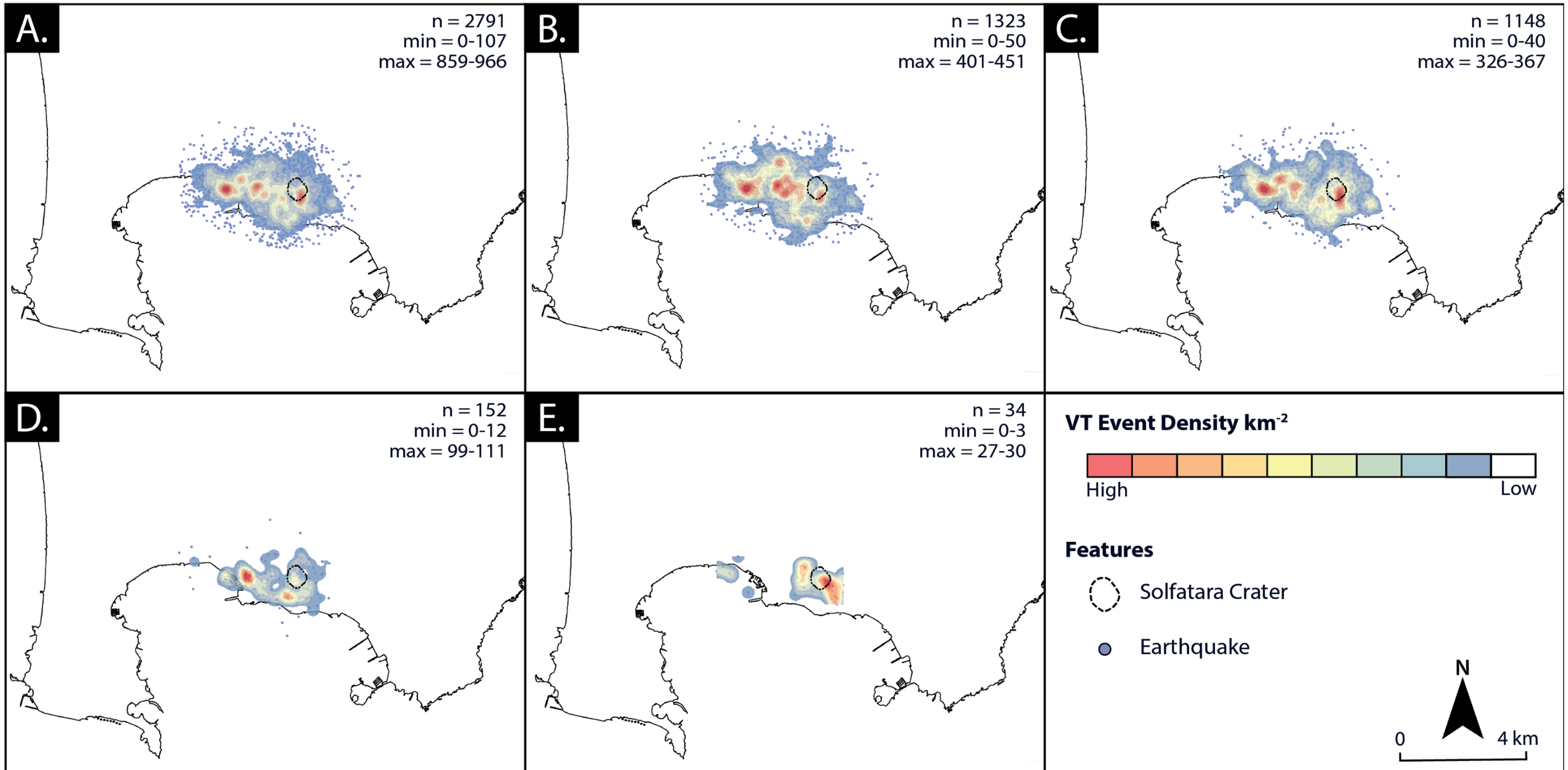


Figure 6.4: Kernel Density distribution of earthquake epicentres of particular magnitudes. A. distribution of all events contained in the catalogue. Panels B., C., D. and E. show the density distribution of epicentres of magnitude (Ms) 0.2 to 1, 1.1 to 2, 2.1 to 3 and more than 3.1 respectively. The largest magnitude events (Ms ≥3.1) are concentrated within 2 km of Solfatara.

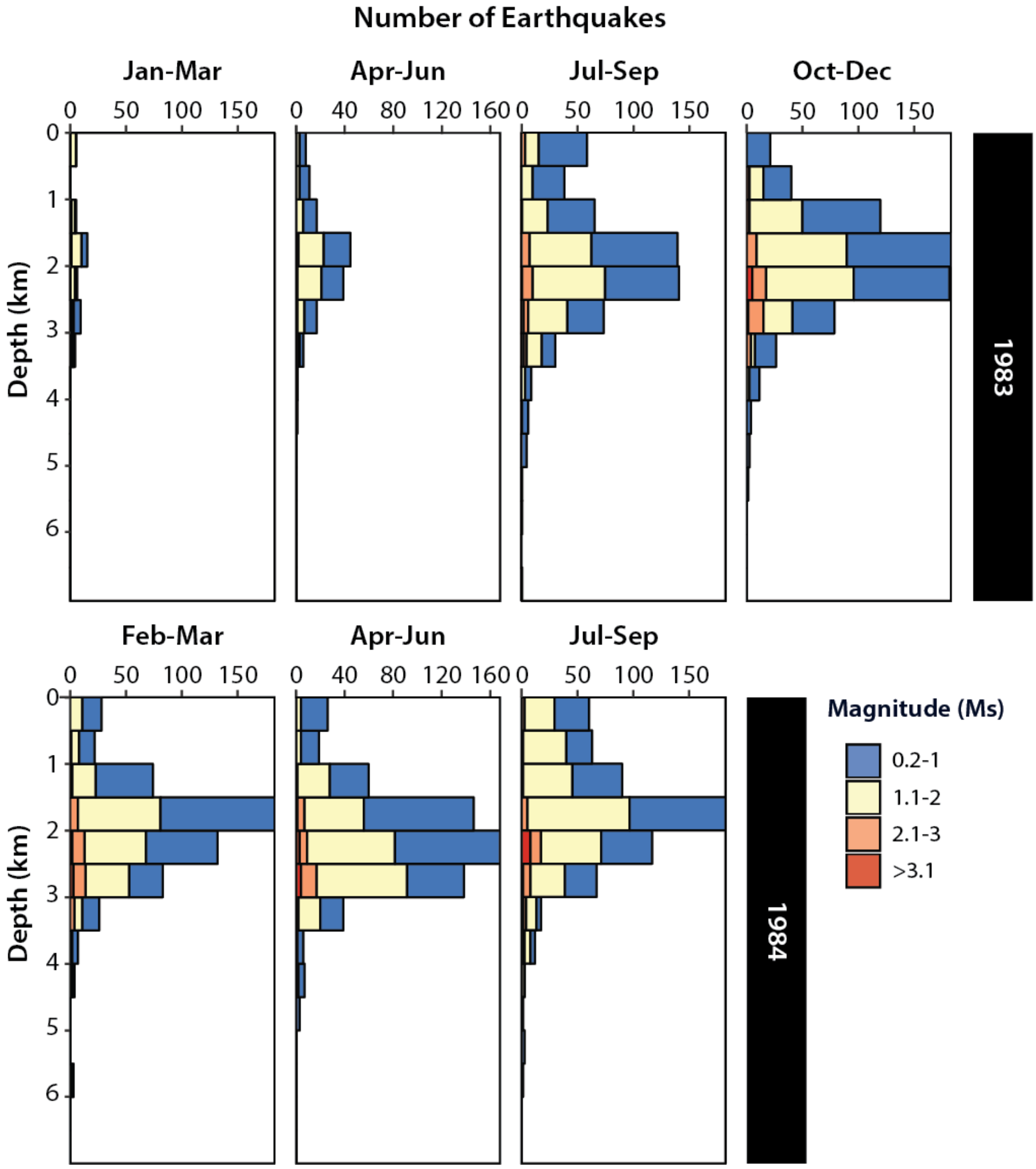


Figure 6.5: Earthquake frequency through time by depth and magnitude. Most events are clustered at depths shallower than the expected limit of the hydrothermal system at 3 km depth, whilst the largest events are clustered between 1.5 and 3 km, below the expected limit of the caprock.

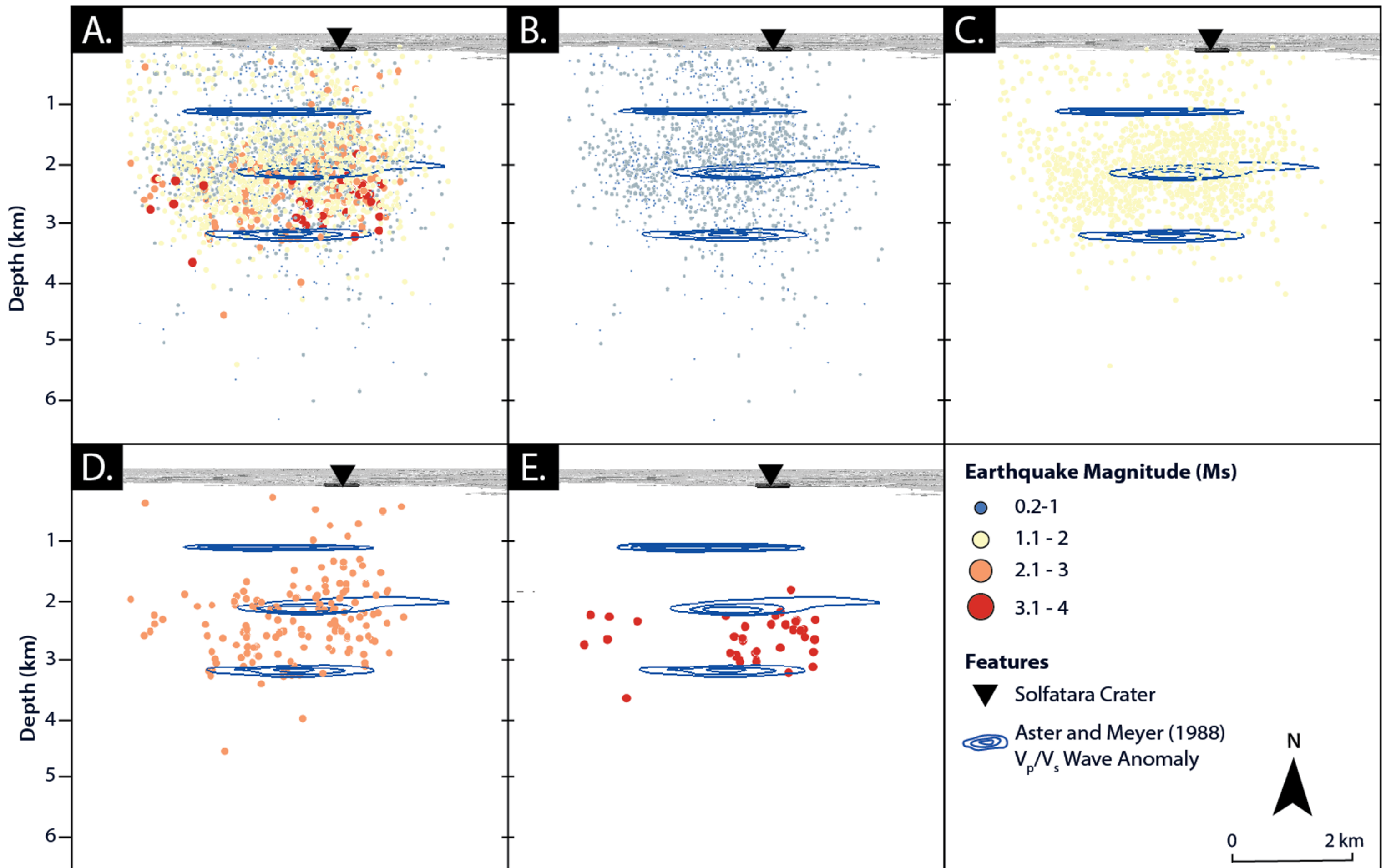


Figure 6.6: Distribution of hypocentres below Pozzuoli looking N from the Gulf of Pozzuoli. The seismogenic volume broadly corresponds with the seismic anomaly identified by Aster and Meyer (1988) as a fractured volume saturated in fluids. The largest events are concentrated below the caprock (c. 1.5 km) within the region expected to contain the hydrothermal reservoirs. Events are coloured by magnitude.

Seismic b-values can be used as an indicator of where stress and fracture growth is concentrated in the crust. When plotted as a function of depth (Fig. 6.7) are observed to decrease from about 1 to 0.6, between c. 1.4 to 3 km depth. Castaldo et al. (2019) have confirmed the trend from a more extensive data set, but for slightly different b-values of 1.2 and 0.8. Below 3 km the value increases to approximately 1 at 4.5 km depth, then remains essentially constant to the limit of detected seismicity at 6.1 km depth. The depths across which the b-values decline (c.1.4-3 km) corresponds to the region where seismicity was concentrated within the hydrothermal reservoir zone. Decreases in b-values with increasing stress are also commonly seen in rock-physics experiments (Main, 1996), for which minima coincide with the onset of rupture. The decrease has been interpreted to reflect an increasing concentration of stress within a smaller volume of the rock being deformed, until a new failure plane is formed or a locked failure plane is reactivated. The deviation in b-values between 1.4-3 km depth is consistent with the concentration of stress in this region and the decay of stress in magnitude away from a laterally extensive pressure source near the base of the hydrothermal system, across the reservoir zone. Incidentally the depth range of 2.5-3 km for minimum b-values agrees well with the depths of sill-shaped pressure sources independently estimated from ground deformation (2.6-3.1 km, Battaglia et al., 2006b; Amoruso et al., 2008; Woo & Kilburn, 2010).

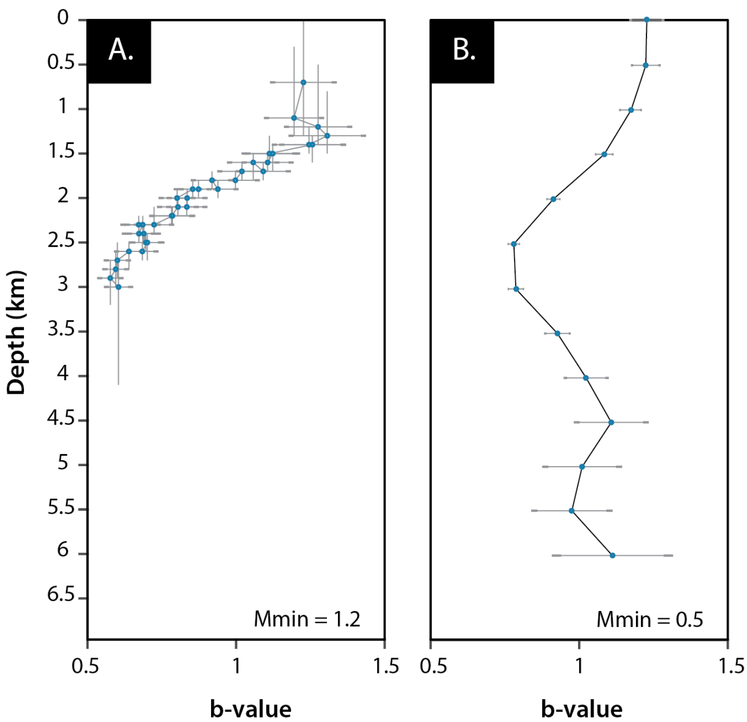


Figure 6.7: The b-value as a function of depth. Panel A. is for the Vesuvius Observatory catalogue. Panel B. is the trend from Castaldo et al. (2019). Panel A. was plotted using ZMap (Wiemer, 2001).

Finally, focal mechanisms of the earthquakes were dominantly normal with occasional strike-slip and dip-slip events confirming deformation in extension. The maximum principal stress was oriented subvertically and the minimum subhorizontally, striking NE-SW (Zupetta and Sava, 1991; De Natale et al., 1995; Chiodini et al., 2001; Acocella et al., 2010; D'Auria et al., 2015). As such the stress field was oriented favourably for the opening of subvertical fractures during the uplift and a permeability increase in this direction. This was observed at the surface, in particular at Solfatara where a NE-SW trending extensional fracture c. 1 m wide and c. 100 m long opened across the crater floor (Fig. 6.7; Italiano et al., 1984; Rosi and Sbrana, 1987).

Figure 6.8: NE-SW trending extensional fracture in the Solfatara crater that opened during the 1982-1984 uplift. The discontinuity is c. 100 m in length. Image modified from Acocella et al. (1999).

Overall the key features of seismicity are summarised as follows.

i. Earthquakes between 1983-1984 show a concentration within the region of the crust hosting hydrothermal reservoirs below Pozzuoli.
ii. The variation in b-values with depth, together with focal mechanisms of earthquakes, suggest that seismicity was caused by extension around a pressure source located towards the base of the hydrothermal system.
iii. The acceleration in the VT event rate with uplift since 1970 implies that inelastic component of deformation increased over time, so that the degree of fracturing and faulting towards the end of the 1982-1984 uplift would be expected to be greater than at any other time since the onset of unrest in 1950.

Together these factors suggest that it is reasonable to assume that the crust hosting the hydrothermal system became increasingly fractured as the ground was uplifted. Given the expected control of fractures on permeability, a redistribution of pore pressure in the hydrothermal reservoirs would necessarily follow.

6.2 Defining a New Model for Hydrothermal Deformation at Campi Flegrei

6.2.1 A Crustal Damage Model for Triggering the Pore Pressure Signal in Deformation

Previous models (Chapter 3) that consider the post-1984 sequence of ground movements to have been triggered by magma intrusion into the shallow crust in 1982-1984 (e.g. Chiodini et al., 2015a; Moretti et al., 2017) have assumed that subsidence and the following uplift represent a depressurisation and re-pressurisation of the hydrothermal system caused by changes in the supply of magmatic fluids. In contrast, it is suggested here that the pore pressure changes, and therefore deformation, were triggered by an increase in permeability caused by the mechanical effect of sill intrusions on the crust. In this context a three-stage response of the hydrothermal system can be recognised.

Stage 1: Initial Uplift timing

Permeability is expected to be enhanced in three ways. First, the bulk permeability in the hydrothermal reservoir zone will progressively increase as crustal damage accumulates. This will occur via reactivation of clogged and sealed fluid pathways, crack growth and linkage at all scales, as well as the opening of existing flow paths related to crustal extension. Second, the hydraulic gradient will be augmented. This is the expected net-effect of the increase in vertical permeability and pressurisation of the base of the hydrothermal system related to the combination of sill inflation and increase in the supply of magmatic fluids due to sill degassing. Any associated poroelastic inflation is assumed to have a minor contribution to the overall uplift relative to that of the sill as the creation of permeability will limit the accumulation of pore pressure. Indeed, an increase in pore pressure will reduce the effective stress, promoting brittle failure (Terzaghi, 1923). Finally, resistance to flow is expected to further decrease over the duration as the H2O boiling zone expands (due to the reduction in confining pressure by fracturing and heating from the increased flux of magmatic fluids), and then as the fraction of magmatic gas progressively increases in the system. Permeability generation ends with uplift at the end of 1984.

Stage 2: Mid-stage Subsidence

Permeability and therefore hydraulic connectivity between the hydrothermal reservoir, the surrounding caldera and the surface will be at a maximum at the end of Stage 1. The reservoir zone is expected to have extended, allowing fluids to escape the circulating system, into the surrounding caldera whilst flow through outflow structures will be enhanced. Pressurisation of the base of the hydrothermal system is expected to have ended with sill inflation at the end of 1984, and any permeability increase from fracturing across the transition between the magmatic and hydrothermal systems at c. 3 km depth is expected to be recovered by SiO_2 precipitation. Outflow will therefore be the dominant control on pore pressure in the hydrothermal system from this point. Recharge of the system may also be limited during this time by the outward transport of pressure.

As pore pressure is lost from the hydrothermal reservoirs, the rate of subsidence will decrease with the hydraulic gradient. Concurrently, a permeability decay is expected from crack closure as pore pressures decline, sealing of fractures and porosity loss by mineral precipitation from fluids, as well as alteration of fresh rock surfaces created by fracturing during the preceding uplift. Subsidence ends once the hydraulic gradient has been reduced sufficiently by pore pressure loss so that enhanced outflow can no longer be sustained. At this time permeability in the hydrothermal system will be at minimum values. An excess of pressure is expected to have been lost from the hydrothermal reservoirs, so that the total subsidence exceeds the contribution of the hydrothermal system to the uplift. Net-uplift will be equivalent to that caused by the sill intrusion.

Stage 3: Late-stage Uplift

A consequence of permeability recovery is the re-establishment of normal flow patterns, allowing pore pressures to be restored by the background supply of magmatic fluids from the primary magma reservoir and recharge. A slow uplift is therefore expected from re-pressurisation of the hydrothermal reservoirs, without invoking any change in the magmatic system, which may eventually return to conditions similar to those prevailing before Stage 1.

6.2.2 A Crustal Damage Model for Triggering the Pore Pressure Signal in Deformation

Stage 1 deformation corresponds with the rapid uplift in 1982-1984, Stage 2 with subsidence between after 1984 to 2004, and Stage 3 with uplift since 2004 (Fig. 6.9). The Stage 2 subsidence represents an excess pore pressure loss from the hydrothermal reservoirs since the end of 1984, whilst net-uplift of c. 0.9 m (measured) in 2004 is attributed to the permanent deformation caused by sill intrusion between 1982-1984. Stage 3 uplift represents a recovery of pore pressure in the hydrothermal system. To be consistent with the observed deformation signal, the model must be able to account for: (i) the timing of the onset of the slow ground oscillation after 1984, and (ii) the symmetry between the following subsidence and uplift since 2004.

In the context of the model the hydrothermal system is only expected to be significantly disturbed once fracturing has increased permeability above a critical threshold. The VT-deformation trend since 1950 suggests that conditions for widespread fracturing were only approached in 1984 (Chapter 5, section 5.2.2). If so, no significant hydrothermal response is expected to have occurred after the previous major uplifts, which is in agreement with the absence of comparable slow ground movements following the 1950-1952 and 1969-1972 episodes. The model therefore meets the first constraint.

The subsidence and the uplift are consistent with the superposition of two trends with time: an exponential decay and an exponential increase (Moretti et al., 2018). The best-fit trends for Moretti et al. (2018) give for the rate of subsidence, dh_s/dt:

$$\frac{dh_s}{dt} = \left(\frac{h_0}{t^*}\right) e^{-\left[\frac{(t-t_0)}{t^*}\right]} \qquad [2]$$

and for the rate of uplift, dh_u/dt :

$$\frac{dh_u}{dt} = \left(\frac{h'}{t^*}\right) e^{-\left[\frac{(t-t_0)}{t^*}\right]} \qquad [3]$$

where h_0 is the ground level at the start of 1985 (time, t_0) and h' is $e^{(\frac{-37}{6})}$, or 2.1 x 10^{-3} m, a notional starting value of the uplift component. Elevations are in metres and time is in years. Equations 2 and 3 are characterised by the same exponential timescale, t^*, of 6 years (Moretti et al., 2018). Given that the timescale is controlled by the physical mechanism driving the ground movements, it follows that uplift represents a reversal of the process controlling the preceding subsidence. An increase in permeability in the hydrothermal reservoir, followed by a loss of pore pressure due to fluid flow, would naturally be succeeded by a period of permeability recovery by crack closure and mineral sealing processes. Such processes are known to occur over relevant time scales (e.g. Lowell et al., 1993; Fournier et al., 1993). Provided background rates of degassing from the magma reservoir and flow paths for meteoric recharge were not permanently altered, a re-pressurisation of the hydrothermal reservoir and uplift would necessarily follow, satisfying the second constraint. This is reasonable to assume due to the constant supply of CO_2 to the surface, and the lack of evidence for a major hydrological reorganisation from trends in thermal water characteristics and the distribution of surface activity (Chapter 2). Furthermore, the clustering of low magnitude earthquakes at depths less than 2.5 km depth, within the volume of the crust containing the hydrothermal reservoirs (Appendix B) during uplift since 2004 is compatible with a pressurisation that results from internal processes.

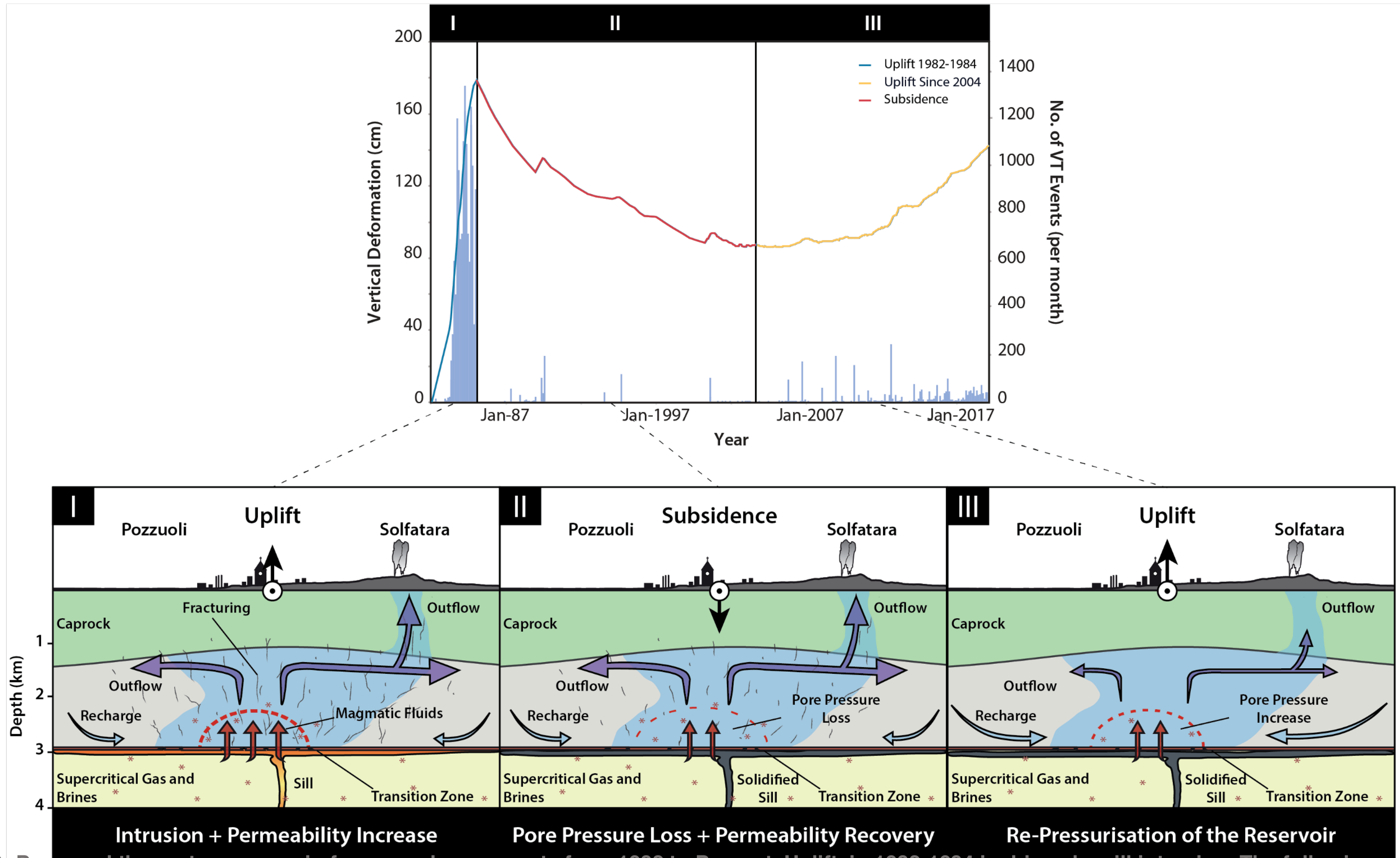


Figure 6.9: Proposed three-stage scenario for ground movements from 1982 to Present. Uplift in 1982-1984 is driven by sill intrusion. The following subsidence and slow uplift is attributed to a pore pressure loss and subsequent recovery triggered by a permeability enhancement resulting from widespread fracturing during the sill intrusion.

6.3 Discussion

6.3.1 Deformation Since 1950 as a Single Evolutionary Sequence

Contrary to existing interpretations, the pore pressure signal after 1984 is considered here to be part of a single long-term evolutionary sequence since 1950 and is a consequence of progressive deformation of the crust by successive sill intrusions in 1950-1952, 1969-1972 and 1982-1984. The principal requirement of the model is that the accumulation of stress in the crust between 1950-1984 was sufficient for the exceedance of a critical threshold of fracturing in 1982-1984 to trigger a response of the hydrothermal system. As such, it is dependent on the emplacement of a permanent strain source in each uplift during this period. At no point is the model dependent on the presence of shallow magma for deformation to occur. The necessity of a permanent strain source could therefore be satisfied by either magma intrusions, or the accumulation of magmatic fluids below a hydrological barrier, similar to the process of uplift invoked by De Vivo and Lima (2006) and others (Chapter 3).

The depth distribution of seismicity during the 1982-1984 uplift (maximum depth of c. 6 km) and the enrichment of fumarole gases in N_2, a gas expected to originate at the primary magma reservoir at c. 7 km depth (Caliro et al., 2007), support deep fluid transport to shallow depths during this unrest. However, it is not possible to definitively establish whether this fluid was magma or exsolved gases and brines. The assumed involvement of magma in the model is therefore subjective but is considered reasonable as it can simply account for net-uplift since 1950 and the episodic nature of the rapid uplifts. Furthermore, this is in agreement with the results of expert elicitation exercises used to parameterise the Bayesian Event Tree for forecasting at Campi Flegrei, which concluded that the rates of deformation of the magnitude observed during 1982-1984 unrest are most probably due to magma intrusion (Selva et al., 2012).

The exceedance of a critical threshold of fracturing and therefore permeability in 1984, was inferred from the long-term trend in rates of VT seismicity with increasing uplift since 1950, which were interpreted by Kilburn et al. (2017) as indicating an approach towards conditions for bulk failure in the crust in 1984. The timing of the emergence of the slow ground oscillation in the model after 1984 is therefore dependent on these conditions having been met. An alternative scenario where the 1982-1984 sill was intruded into a region of the crust that was more favourable for brittle deformation and a response of the hydrothermal system than previous intrusions is rejected on the basis of: the accelerating rates of VT seismicity with increasing uplift between 1970-1984, the overlapping source depths estimated for the 1969-1972 and 1982-1984 uplifts from inversions assuming a sill shaped source (2.5-3.1 km, Bonasia et al., 1984; Woo, 2007; Battaglia et al., 2006b; Amoruso et al., 2008), and the constant geometry of ground movements through time, which implies the same region of crust accommodated both ground movements (Chapter 3).

6.3.2 Compatibility of the Model with Compositional Changes in Solfatara Fumarole Gases

In Chapter 3 it was established that there was a distinct change in the behaviour of compositional trends in Solfatara fumarole gases in 2000. For the model to be a viable scenario for post-1984 ground movements, it must be able to account for: (i) the change from an oscillatory trend in the CO_2/H_2O ratio to a continuous increase; (ii) the loss of co-variance between CO_2 and N_2 from this time; and (iii) an increase in the frequency of peaks in the redox indicator CO_2/CH_4.

Geodetic inversions of the 1982-1984 uplift that assume a sill-like geometry, suggest that such a source would be ~1-10 m thick, with volumes ~0.02-0.04 km^3 (Battaglia et al., 2006; Woo and Kilburn, 2010). Such bodies would be expected to have completely solidified within a few years of emplacement (section 5.1). The peak in the CO_2/H_2O ratio following the 1982-1984 unrest is assumed to represent an input of magmatic fluids into the hydrothermal system during sill emplacement, in agreement with existing interpretations (e.g. Chiodini et al., 2015a; Moretti et al., 2017). The following decline between 1986 to 2000 is then attributed to a combination of the solidification of the sill, the progressive transport of magmatic-gas rich fluids out of the main hydrothermal reservoir by outflow and degassing and an increase in the H_2O vapour fraction caused by decompression boiling of hydrothermal fluids. Peaks in the CO_2/H_2O ratio superimposed on the main trend during this period correspond to mini-uplifts in 1989 and 1994. Contrary to existing interpretations (e.g. Chiodini et al., 2003), it is suggested that they are related to sealing-rupture processes in the hydrothermal system that resulted in the accumulation and release of gas rich fluids into the Solfatara plume from the main reservoir, rather than being transient accelerations in magma degassing (e.g. Chiodini et al., 2003).

By 2000, the primary source of magmatic fluids in the magmatic reservoir is expected to be the main magma reservoir at c. 7 km, in agreement with Moretti et al. (2017). At this time fluid transport rates out of the hydrothermal system are assumed to be at a minimum due to the decline in pore pressure, and the non-condensable gas fraction (e.g. CO_2) in the main hydrothermal reservoir to be at a maximum due to a combination of the loss of H_2O by boiling and background degassing. Permeability is also expected to be at minimum values since the end of the 1982-1984 unrest. From this point onwards, the process of restoration of equilibrium conditions in the hydrothermal system is expected to begin. As pore pressures in the hydrothermal reservoir are re-established, rates of fluid transport into the Solfatara plume would be expected to increase as flow paths are reactivated and/or dilate, resulting in an enrichment of CO_2 in fumarole gases as observed. This would also account for the expansion of the CO_2 diffuse degassing area at Solfatara-Pisciarelli since the early 2000s and the increase in activity at Pisciarelli. The decrease in H_2O can be related to a combination of the increase in the CO_2 fraction in the main reservoir and condensation in the Solfatara plume as pore pressures increase (e.g. Todesco et al., 2003).

The assumption of two degassing sources (i.e. the intruded sill and the primary magma reservoir) is supported by the trend in N_2. Before 2000 the compositional trend in this gas was the same as that for CO_2. This is taken to indicate a common source (Chapter 3), which is assumed to be the N_2 gas-rich sill. N_2 concentrations would therefore be expected to remain stable at minimum values from once magmatic fluids from the sill had been removed from the hydrothermal system.

The final constraint that must be satisfied is the increase in the frequency of peaks in CO_2/CH_4 in fumarole gases after 2000 identified by Chiodini et al. (2015a). Recalling that conditions in the main hydrothermal reservoir are oxidising (Chapter 2; Caliro et al., 2007), an increase in pore pressure in this region would favour a more regular enrichment of the Solfatara plume in oxidising, magmatic-gas rich fluids, suppressing the formation of CH_4. This can account for the observed trend but would require that CH_4 equilibrates in the plume, in agreement with Moretti et al. (2017), rather than in the main reservoir as suggested by Caliro et al. (2007) and assumed by Chiodini et al. (2015a; 2016). Overall the proposed crustal damage model for post-1984 ground movements appears to be compatible with geochemical changes in fumarole gases over the same period.

6.3.3 Implications for Hazard and Expected Evolution of Uplift

Contrary to previous models (Chapter 3), subsidence after 1984 is interpreted here as a departure from, rather than a return to, equilibrium conditions in the hydrothermal system. Instead, the current uplift since 2004 is considered to represent a progressive return to equilibrium. The model does not require an increase in the rate of transport of magmatic fluids into the hydrothermal system in order for this uplift to occur (e.g. Moretti et al., 2017) or, the presence of shallow magma (e.g. Chiodini et al., 2015a; 2016). As such, the eruption hazard is expected to be at its lowest since the 1982-1984 unrest episode. Additionally, the mechanism of uplift is unlikely to lead to the generation of overpressures in the hydrothermal system that are great enough to significantly increase the probability of a phreatic eruption or hydrothermal explosion.

Unless the system is disturbed by another intrusion, and assuming that there has been no permanent increase in connectivity between the magmatic-hydrothermal systems during the 1982-1984 uplift, it is expected that uplift since 2004 will end once equilibrium conditions have been restored. It is suggested that this will occur as the ground level approaches that at the end of the 1982-1984 uplift. Assuming the exponential trend from Moretti et al. (2017) in section 6.3.2 continues, this would be expected to occur within 5 years. At this point stress conditions in the crust conditions in the crust would have returned to those in 1984, so that in the event of a future magma intrusion a greater component of deformation would be expected to be accommodated by faulting and fracturing than at any point previously along the evolutionary sequence (Kilburn et al., 2017). The likelihood of a disturbance of the hydrothermal system and a pore pressure signal in deformation would therefore be greater than for previous episodes. If the current uplift continues the current trend beyond the maximum ground level, then this would favour alternative models that consider this ground movement to result from increased transport of pressurised fluids originating in the magmatic system to shallower depths (e.g. Chiodini et al., 2015a; 2016; Moretti et al., 2017).

6.4 Conclusions

The slow ground oscillation after 1984 is compatible with a depressurisation and re-pressurisation of the hydrothermal system, triggered by the mechanical effect of successive sill intrusions in 1950-1952, 1969-1972 and 1982-1984. Deformation since the onset of unrest in 1950 can be therefore be considered as a single evolutionary sequence, where the probability of a response of the hydrothermal system to unrest increased with uplift between 1950-1984, as stress accumulated in the crust and it became increasingly fractured. The implication is that for the pore pressure signal to emerge, a critical threshold of stretching of the crust had to be exceeded. As such, post-1984 ground movements can be considered to represent a continued adjustment of the hydrothermal system to the 1982-1984 unrest and an indicator of strain in the crust.

The compatibility of the model with both geodetic and geochemical parameters suggests, albeit qualitatively, that: (i) an increased coupling between the magmatic-hydrothermal systems is not a pre-requisite for slow uplift, contrary to existing models, (ii) uplift since 2004 is not dependent on the evolution of the magma body intruded in 1982-1984, (iii) the hydrothermal system is moving towards, rather than away from, equilibrium conditions, and (iv) past uplift episodes may not provide a reliable basis for defining future scenarios of unrest without consideration of long-term cumulative changes in the crust. Critically, the model does not require magma to be located at shallow depths to generate ongoing uplift since 2004. As such, the eruption hazard is expected to be at its lowest since the 1982-1984 unrest episode.

Chapter 7

Scientists Perceptions of the Role of the Hydrothermal System in Unrest at Campi Flegrei

Campi Flegrei is one of the most intensely monitored volcanoes in the world. The Vesuvius Observatory has the primary legal responsibility for volcano surveillance and managing the monitoring network, as a section of the Istituto Nazionale di Geofisica e Vulcanologia (INGV). Additional monitoring activities are conducted by the Istituto per il rilevamento elettromagnetico dell'ambiente (Irea) and the Consiglio Nazionale delle Ricerche (CNR). Together with the Observatory, the three institutions comprise the Centres of Competence (Centri di competenza) that advise the Italian Department of Civil Protection (DCP) about activity and hazard assessment at Campi Flegrei. The DCP have the responsibility for deciding the Volcano Alert Level (VAL), which they do based on the information provided by the Centres of Competence, in consultation with the group of volcano experts that comprise the Major Risks Committee – Volcanic Risk Section (Commissione Grandi Rischi - Settore Rischio Vulcanico, Protezione Civile, 2019).

The perceived role of the hydrothermal system and the processes controlling its contribution to ground movements at Campi Flegrei has implications for short- to long-term hazard assessment. For example, the latent heat of condensation model for uplift since 2004 (Chapter 3, Chiodini et al., 2015a; 2016) considers there to be a magma body currently located at shallow depths in the crust that is evolving towards conditions for eruption. In common with this model, both the CO_2-drying model (Chapter 3, Moretti et al., 2017) and that proposed in the previous chapter, consider this uplift to result from a pressurisation of the hydrothermal system, but contrast in that they expect the potential for eruption to be at its lowest since the 1982-1984 unrest.

The academic debate as to the cause of the current and past episodes of uplift is well established, but the extent to which it exists amongst those who may be involved in the scientific operational response to a future escalation in activity, has yet to be evaluated. Expert elicitation exercises have focused on the definition of thresholds in monitoring observations for the parameterisation of a Bayesian Event Tree for probabilistic forecasting of eruptions (BETEF-CF, Selva et al., 2012), whilst hazard centred perception studies have principally been related to understanding public knowledge of, and preparedness for, volcanic hazards from Vesuvius (e.g. Davis et al., 2005; Barberi et al., 2008; Carlino et al., 2008; Ricci et al., 2013a) and Campi Flegrei (Ricci et al., 2013b). An exception is a study of the perception of volcanic hazards and emergency plans at Vesuvius amongst monitoring scientists from the Vesuvius Observatory and local authorities by Solana et al. (2008).

Here the results of a survey conducted amongst the monitoring scientists at the Vesuvius Observatory, and other experts on the Campi Flegrei volcanic system is presented with the principal aims of establishing; the extent to which perceptions of the role of the hydrothermal system in ground movements differ, perceptions of the hazard associated with uplift since 2004, and perceived challenges for communication of the volcano's status between stakeholders.

7.1 Methodology

To evaluate the degree to which differing conceptual models of unrest at Campi Flegrei exist, a survey was conducted among monitoring scientists based at the Vesuvius Observatory and a group of researchers from both governmental and academic organisations. Data collection occurred in two phases. In the first, a paper-based survey was distributed to scientists at the Vesuvius Observatory in June 2019 using a purposive (non-probability) sampling approach in collaboration with Dr. E. Marotta from the Observatory. Distribution via a recognised individual is a common practice (Bird et al., 2009) and was advantageous in that it facilitated greater access to the scientists, whilst the presence of a native Italian speaker ensured that the nature of the study was fully understood by potential participants. To avoid partiality in participant selection, all 60 scientists present at the Observatory during the fieldwork period were provided with a copy of the survey. 50 surveys were returned, 45 of which were useable, giving an overall response rate of 75%. During the second phase of data collection an online version of the survey was constructed using the web-based tool Opinio™ (Version 7.11) and hosted on the University College London server. A link to the survey was sent with an invitation to participate in the study to a list of 41 Italian researchers identified from the literature who have previously published work on Campi Flegrei. 17 provided responses.

Both surveys were translated into Italian by a native speaker, then reverse translated into English to check that the meanings of the questions had been maintained. All data collection was carried out in Italian and survey answers were translated into English using an online translation tool (deepl.com), then checked with a fluent Italian speaker for accuracy (C. Kilburn, UCL). Both the paper and online versions of the survey were self-administered and completed anonymously, in the respondents' own time. Approval to carry out the study was obtained from the University College London (UCL) Research Ethics Committee prior to any data collection (Appendix C) and permission to conduct the survey at the Vesuvius Observatory was given by the director, Dott.ssa Francesca Bianco. The paper and online versions of the survey are provided in Appendix D with the participant information sheets.

The survey distributed at the Observatory consisted of 27 variables that were formatted as either open or closed questions, or seven-point Likert scales (Likert, 1932) with specified end points. Where closed questions were used, options for *other* and *not sure* were given, in addition to free text boxes, so as not to restrict answers. The questions were grouped into 5 sections (Table. 7.1). The demographic section (questions 1-8) was used to classify the sample according to respondents age, expertise and experience of volcanic crises. It also established whether they expected to advise the Civil Protection in the event of a future volcanic crisis at Campi Flegrei, and how they rate the scientific understanding of the behaviour of the volcano relative to other examples. Questions 9-19 specifically addressed participants views as to the causative processes controlling ground deformation during the periods 1969-1972, 1982-1984, 1985-2004 and 2004-Present, whilst 20-21 were related to criteria for differentiating magmatic unrest. The final two sections asked participants about their perceptions as to other stakeholder groups views of uplift since 2004 (i.e. the Civil Protection, local authorities, the public and the media), and the existence of difficulties related to the communication of the behaviour of Campi Flegrei between stakeholders.

A preliminary analysis of the results from the Vesuvius Observatory identified a potential ambiguity in the phrasing of questions 18 and 24. These questions were related to physical hazards associated with unrest, but it was found that some participants gave responses related to risk and unrest management. The use of the phrase *magmatic unrest* in question 21 was also highlighted by one respondent as being ambiguous, as to the role of magma movement. The phrasing of these questions was modified accordingly in the online survey prior to is publication, but their original intended meaning was maintained. The survey was also shortened by the removal of items 10, 12 and 14.

Completed survey responses were assigned a numerical code and analysed using the statistical software package, IBM SPSS Statistics Version 26. The responses to closed questions were coded directly, whereas responses to open questions were first grouped into categories based on a thematic analysis of the answers (Sarantakos, 2012; Bryman, 2012). Data analysis consisted of frequency and cross-tabulation tables in SPSS. Statistical tests based on p-values (e.g. Chi-squared, Fisher's exact tests) were then applied to test for association between participant responses and the variables: age, workplace, field of expertise, years worked on Campi Flegrei, experience of working on a recently erupted volcano; past experience of a volcanic crisis, and expected involvement in advising Civil Protection in the event of a future volcanic crisis at Campi Flegrei. These variables are referred to as indicator variables throughout. A p-value of <0.05 was taken as indicative of a significant association, and <0.01 as highly significant, in accordance with convention (Fisher, 1925, Bryman, 2012).

Table 7.1: The survey consists of 27 variables divided into 5 sections based on theme (1 = demography, 2 = cause of unrest, 3 = interpretation of monitoring parameters,4 = perception of stakeholders, 5 = communication of volcano status between stakeholders). Question formats include; open (Op), closed (Cl) and seven-point Likert scales (Li).

No.	Question	Theme	Format
1	Age	1	Cl
2	Where do you work? (e.g. INGV section, university)	1	Op
3	Which of the following options best describes your primary field of expertise?	1	Cl
4	How many years have you worked on the following volcanoes?	1	Op
5a	Have you experienced a volcanic crisis?	1	Op
5b	If so, in what role?	1	Cl
6	In the event of a future volcanic emergency at Campi Flegrei do you expect to be involved in advising the Civil Protection about the status of the volcano and the expected evolution of the unrest?	1	Cl
7	Rate the level of scientific understanding of the behaviour of the following Italian volcanoes. 1 is very poor and 7 is very good.	1	Li
8	Rate the level of scientific understanding of the behaviour of the following calderas. 1 is very poor and 7 is very good.	1	Li
9a	In your opinion, what was the most likely cause of the uplift in 1969-1972? Why?	2	Cl
9b	Why?	2	Op
9c	At approximately what depth(s) in the crust do you expect that the pressure source or sources were located?	2	Cl
10	To what extent do you think the following groups are in agreement about the cause of the 1969-1972 uplift. 1 is no agreement and 7 is total agreement.	2	Li
11a	In your opinion, what was the most likely cause of the uplift in 1982-1984?	2	Cl
11b	Why?	2	Op
11c	At approximately what depth(s) in the crust do you expect that the pressure source or sources were located?	2	Cl
12	To what extent do you think the following groups are in agreement about the cause of the 1982-1984 uplift. 1 is no agreement and 7 is total agreement.	2	Li
13a	In your opinion what was the most likely cause of subsidence after 1984?	2	Cl
13b	Why?	2	Op
13c	At approximately what depth(s) in the crust do you expect that the pressure source or sources were located?	2	Cl
14	To what extent do you think the following groups are in agreement about the cause of the subsidence. 1 is no agreement and 7 is total agreement.	2	Li
15a	In your opinion what is the most likely cause of the uplift since 2004?	2	Cl
15b	Why?	2	Op
15c	At approximately what depth(s) in the crust do you expect that the pressure source or sources were located?	2	Cl
16	To what extent do you think the following groups are in agreement about the cause of uplift since 2004. 1 is no agreement and 7 is total agreement.	2	Li
17	What do you think is the most likely scenario for how the current uplift will end?	2	Op
18	What do you consider to be the main hazards associated with the current uplift?	2	Op
19	At approximately what depths do you expect that magma is being stored in the crust at Campi Flegrei today?	2	Cl
20	What changes in monitoring parameters would you expect to be characteristic of a magma intrusion at Campi Flegrei?	3	Op
21a	What do you consider to be the key criteria for determining whether unrest is magmatic or non-magmatic in origin?	3	Op
21b	Are there any new datasets or tools not currently in use that could help refine interpretations?	3	Op
22	How well would you rate the understanding of the behaviour of Campi Flegrei by the following groups? 1 is very poor and 7 is very good.	4	Cl
23	What do you think the following groups think is the source of the current uplift?	4	Cl
24	What hazards do you think the following groups are most concerned about during the current uplift?	4	Op
25a	At present are there any challenges or difficulties regarding communication of the volcano's behaviour between scientists, the Civil Protection, the local authorities, the media and the public?	5	Cl
25b	If yes, what are they?	5	Op
26	In the event of a future volcanic emergency at Campi Flegrei, how well do you expect the following groups to understand information they receive about the volcano? 1 is not at all and 7 is very well.	5	Cl
27a	In a future volcanic emergency at Campi Flegrei do you anticipate any particular difficulties or challenges related to communicating the cause of unrest and the volcano status between scientists, the Civil Protection, the local authorities, the media and the public?	5	Cl
27b	If yes, what are they?	5	Op

7.2 Results

7.2.1 Survey Sample

The sample included all age categories (18-30 to 61+), but the modal age was 51-60 years (46.8%, n= 62, Fig. 7.1a). 79% of participants currently work at the Vesuvius Observatory, and a further 8.1% work at other sections of the INGV or at the Consiglio Nazionale delle Ricerche (CNR). The rest of the sample are researchers at academic institutions, except for one, who is retired (Fig. 7.1b). Participants came from a range of disciplines (Fig. 7.1c). The most frequently selected were *Geophysics (Seismology)* (31.1%, n= 61) and *Geology (Earth Sciences)* (27.9%). 14.8% of respondents selected *Other* and given examples included: physical volcanology, hazard assessment, and eruption modelling.

Overall, the sample had a high degree of familiarity with Campi Flegrei. 95.2% (n= 62) have worked on the volcano (2 participants did not give an answer, only 1 has not previously worked on Campi Flegrei) and 76.7% (n= 60) have worked on it for more than 5 years (Fig. 7.1d). All participants who are based at organisations other than the Vesuvius Observatory have worked on it for more than 10 years. 55% (n= 60) and 51.3% of the total sample have worked on the persistently active volcanoes of Stromboli and Etna respectively. 65% (n= 60) have previously experienced a volcanic crisis (Fig. 7.1e); 51.7% as a monitoring scientist and 13.3% in an advisory role to those responsible for managing the crisis response (e.g. civil protection). 51.7% had experienced a crisis on a recently eruptive volcano (e.g. Stromboli, Etna or non-Italian volcano) and 36.7% stated that they had experienced a crisis at Campi Flegrei. Of those that listed Campi Flegrei (n= 22), 12 gave dates during the most recent uplift episode since 2004, and 9 had experienced the major unrest in 1982-1984 (5 were monitoring scientists at the time and 4 were members of the public and not in an operational role at the time). 1 respondent stated they had been a monitoring scientist during a crisis in 1978 and another in 1980. Most likely these dates relate to tectonic activity, rather than volcanic unrest. The comparatively low number of responses that stated they had experienced a crisis at Campi Flegrei since 2004 compared to the number of people currently working at the Vesuvius Observatory suggests that different definitions exist amongst the sample as to what constitutes a crisis; but this cannot be tested here. Overall, 20 respondents (32.8%, n= 61) expect to advise the Civil Protection as to the status of the volcano in the event of a future crisis at Campi Flegrei. 6 are from organisations other than the Vesuvius Observatory.

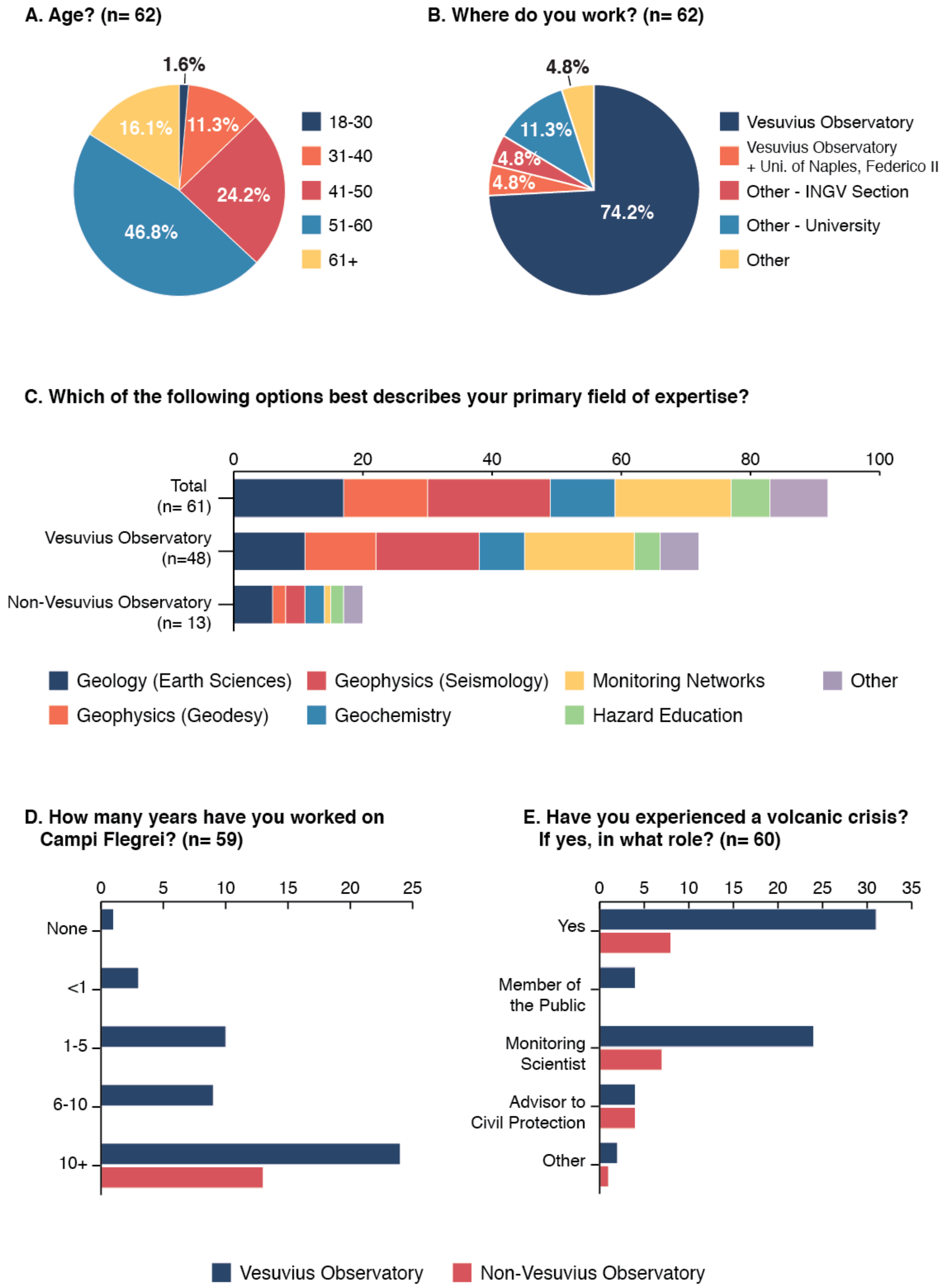


Figure 7.1: Sample characteristics. A. Age of participants, B. Where participants work, C. Participant fields of expertise, D. Participant experience of working on Campi Flegrei, E. Participant experience of volcanic crises.

Participants were asked to rate the scientific understanding of Campi Flegrei, a list of Italian volcanoes and other large calderas on a seven-point scale where 1 is *very poor* and 7 is *very good* (Fig. 7.2). Overall the understanding of the behaviour of Campi Flegrei was considered to be moderate-good with mean and modal responses of 5.2 (n= 61, SD= 1.412) and 5 respectively. These scores are almost identical to those for Vesuvius and the mean is comparable to those for the regularly eruptive centres of Stromboli and Etna, although more people rated the understanding of these volcanoes to be higher than Campi Flegrei. Campi Flegrei was considered to be the most well understood of the caldera-type volcanoes listed (i.e. Ischia, Long Valley, Nisyros, Yellowstone and Rabaul). Italian volcanoes were all rated as being better understood than non-Italian examples. No association was found between how participants rated the understanding of volcanoes and the indicator variables listed in section 7.1.

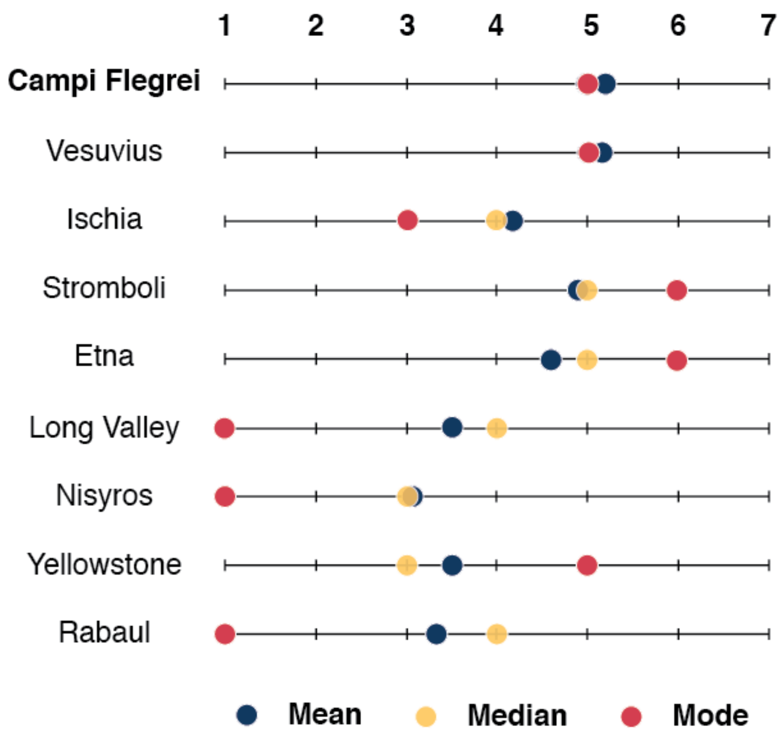


Figure 7.2: Perceived scientific understanding of Campi Flegrei compared to other Italian volcanoes and large calderas. 1= very poor, 7= very good.

7.2.2 Unrest in 1969-1972

Question 9a asked participants *in your opinion, what was the most likely cause of the uplift in 1969-1972?* 70.6% (n=51) expect that this uplift was either totally (27.5%) or partially (43.1%) the result of a magma intrusion (Fig. 7.3a). Most (61.1%, n= 36) consider that a component of the deformation resulted from a pressurisation of the hydrothermal system. 15 respondents (29.4%, n= 51) preferred a non-magmatic scenario for uplift, with 10 selecting *inflation of the hydrothermal system due to an injection of magmatic fluids* and 3 choosing *injection of magmatic gas and brine below an impermeable layer in the crust* (Fig. 7.3b). 2 suggested that uplift resulted from a combination of these two processes, whilst 7 selected *Not Sure*. 1 stated that it was not possible to say what the most likely cause of the uplift was due to insufficient monitoring data. Those that expect to advise the Civil Protection in a future crisis were most likely to consider the cause of unrest to be magmatic but with a component of the uplift resulting from an inflation of the hydrothermal system (n= 17). 20 participants provided a justification for their preferred scenario of unrest (Table. 7.2), 18 of whom expect the unrest was magmatic in origin. Most of the reasons given were related to the characteristics of deformation, in particular the permanent component of deformation, and seismicity. 2 respondents stated that the permeability of the crust was too low for the unrest to have a non-magmatic cause. When asked to rate the level of agreement amongst scientists as to the cause of unrest on a seven-point scale where 1 represents *no agreement* and 7 *total agreement*, responses ranged from 1 to 7 but most selected between 4-6, indicating a perception that scientists somewhat to mostly agree (Table. 7.3).

55 participants provided a response to the question *at approximately what depth(s) in the crust do you expect that the pressure source or sources were located?* The majority expects that the source was shallow (24.4% between *3-4 km* and 33.3% between *4-5 km* depth where n= 45), irrespective of the preferred scenario for uplift (Fig. 7.3c). Participants from the Vesuvius Observatory were most likely to select *4-5 km*, whilst non-Vesuvius Observatory participants were more likely to select *3-4 km* (p= 0.001). Those that consider the unrest to have been magmatic and those that expect to advise the Civil Protection in the event of a future crisis were most likely to select *3-4 km* depth. Those that preferred a non-magmatic scenario for unrest generally considered the source to be deeper at *4-5 km* depth (Fig. 7.3d). 4 respondents provided a range of depths (1-5 km, 2-4 km, 3-6 km and 4-6 km) and 2 stated that there were

two pressure sources, one at *2-3 km* depth and one at *6-7 km* depth. These responses were categorised as *Other*. There is no clear relationship between preferred scenarios and source depths, although those that selected *magma intrusion* as the source of uplift were more likely to select *3-4 km*.

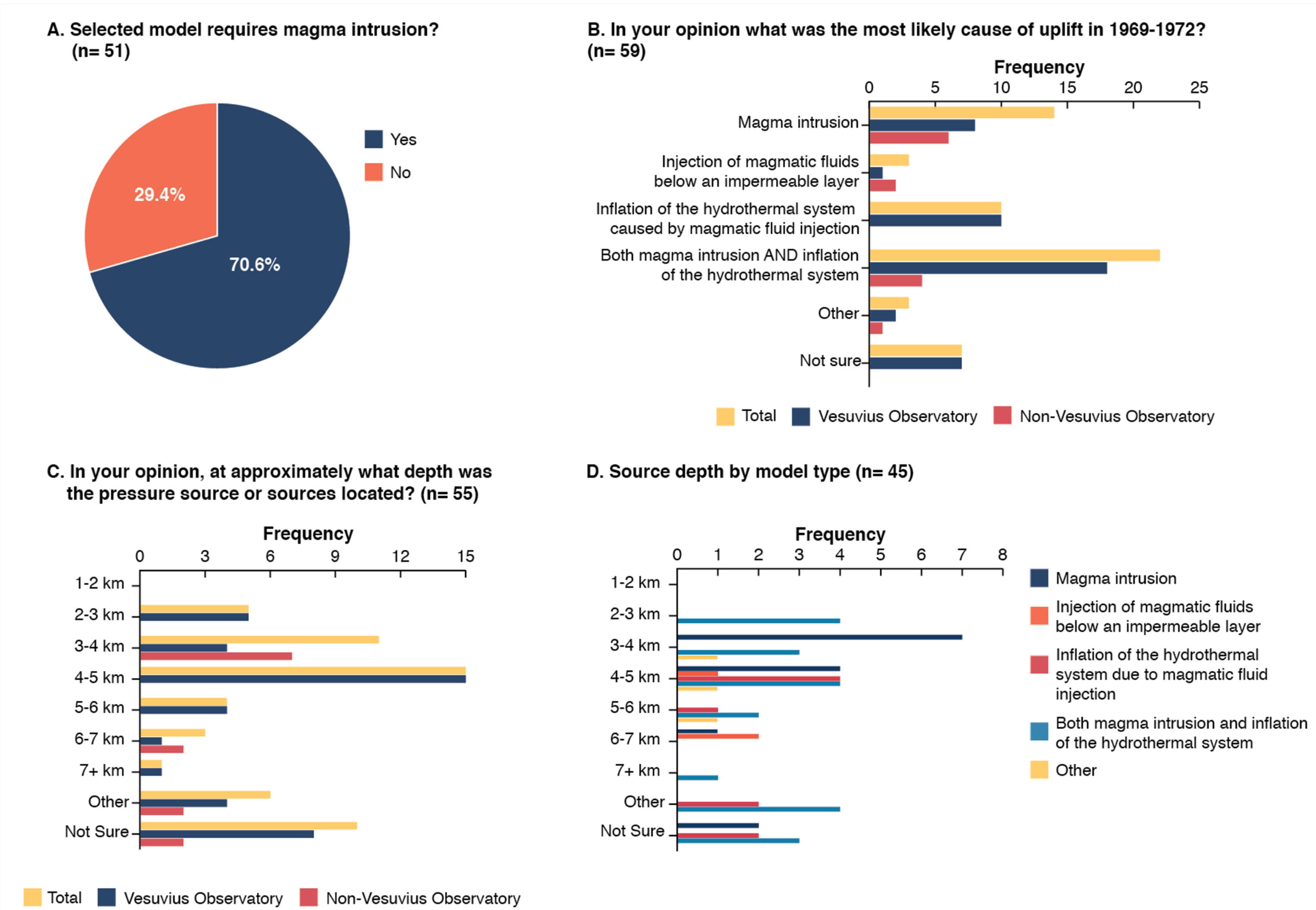


Figure 7.3: Survey participants preferred scenarios and source depths for the 1969-1972 uplift. A. Proportion of participants who consider the uplift to be partially or totally the result of a magma intrusion. B. Participants preferred scenarios for the cause of uplift. C. Preferred source depths for uplift. Responses that selected multiple depths ranges are categorised as other. D. Preferred source depths by uplift scenario.

Table 7.2: Reasons given for participants response to the question in your opinion, *what was the most likely cause of the uplift in 1969-1972?* The scenario column indicates the participants preferred causal mechanism of the deformation. (Mi = magma intrusion, Inf = inflation of the hydrothermal system due to an injection of magmatic fluids, Mi + Inf = both magma intrusion and inflation of the hydrothermal system, Ot = Other).

Scenario	Reason for Selection
Mi	Cannot be explained by fluid injection due to low permeability.
	Low permeability - quickly returned to previous state.
	Rapid uplift, similar to the 1982-1984 episode.
	We now know from other evidence (petrology, seismic tomography) that Campi Flegrei is characterised by shallow magma pockets…
	Results of recent studies and no subsequent subsidence.
	There was no subsidence...
	…a sill can most realistically explain the ground deformation, seismicity and hydrothermal activity. It is also the most consistent with the history of the caldera, which is characterised by episodes of intrusion that, in most cases, do not lead to eruptions.
Inf	Accepted Hypothesis.
Mi + Inf	Earthquakes, localisation of deformation, shallow pressure source. Low seismic energy and shallow depths can be associated with hydrothermal processes.
	Ground level did not return to initial level.
	On the basis of knowledge of bradyseismicity and the following crisis.
	Based on knowledge of later crises.
	… subsequent deformation models are only compatible with a magmatic source at shallow depths.
	Permanent deformation; hydrothermal system has secondary role represented by subsidence of c. 20 cm
	Depth and magnitude of seismicity, deformation behaviour.
	By exclusion. Other possibilities do not adequately explain all observations.
	…samples from boreholes indicate many thin sills are emplaced at shallow depths. [Sills] are emplaced at some discontinuity…then release gas upward into the hydrothermal system. This induces a pore pressure increase…contributing to deformation. The magma causes the permanent residual deformation. …
	Ground deformation rate and earthquake locations are consistent with a magmatic intrusion at shallow depth, minor geochemical signals were detected.
	…uplift after the 1950s has only previously been observed in the last 2000 years prior to the 1538 eruption… In addition, the hydrothermal system is highly developed so it certainly has a role to play.
Ot	Phreatic activity.

Table 7.3: ***To what extent do you think the following groups are in agreement about the cause of the 1969-1972 uplift?*** **1= no agreement, 7= total agreement. Values are in % except for the Mean and Standard Deviation (SD).**

	1	2	3	4	5	6	7	Mean	SD
Neapolitan Scientists (n= 32)	3.1	6.3	3.1	28.1	28.1	28.1	3.1	4.69	1.355
Italian Scientists (n= 30)	-	3.3	6.7	26.7	43.3	20	-	4.7	0.988
International Scientists (n= 26)	7.7	3.8	7.7	34.6	26.9	19.2	-	4.27	1.402

7.2.3 Unrest in 1982-1984

As for the 1969-1972 uplift, most participants (69.1%, n= 55) consider that the 1982-1984 uplift was either totally or partially the result of a magma intrusion (Fig. 7.4a). Of those that prefer a magmatic source (Fig. 7.4b), most expect that a component of the uplift resulted from an inflation of the hydrothermal system (63.2%, n= 38). Non-Vesuvius Observatory participants were more likely to select *magma intrusion* (p= 0.008), as were those who identified having expertise in *Geology (Earth Sciences)*. Participants that expect to advise the Civil Protection in the event of a future crisis were equally likely to select *magma intrusion* and *both magma intrusion and an inflation of the hydrothermal system* as the cause of unrest. Those that prefer a non-magmatic source for unrest were most likely to select an *inflation of the hydrothermal system due to an injection of magmatic fluids* as the cause of uplift (58.8%, n= 17). 23 people provided a response when asked *why* they had selected a particular scenario for unrest (Table 7.4). As for the 1969-1972 unrest, there was no relationship between the response given and participants' preferred scenario for unrest. Most responses referred to the characteristics of deformation, in particular the rate, localisation and net-uplift, as well as the results of geochemical and deformation modelling studies. Reasons for assuming a hydrothermal component included: subsidence following uplift gravimetric changes, and the results of geochemical modelling. When asked to rate the level of agreement amongst scientists as to the cause of unrest on a seven-point scale where 1 represents *no agreement* and 7 *total agreement*, the responses were comparable to those for the 1969-1972 uplift and ranged from 1 to 7. Most selected between 4-6 (Table. 7.5).

There is a general agreement amongst the sample that the source of the 1982-1984 uplift was located at depths shallower than 5 km (78.9%, n= 57, Fig. 7.4c) and the most frequent response was *4-5 km* depth (42%, n= 50). 7 participants selected *Not Sure*. Participants who do not work at the Vesuvius Observatory were most likely to select *3-4 km* depth, as were those who indicated that they expected to advise the Civil Protection in the event of a future crisis. 3 people considered two pressure sources to have been active during the uplift; 2 located the sources at *2-3 and 6-7 km* depth, and 1 chose *2-3 and 4-5 km* depth. 4 gave a depth range (1-4, 2-4, 3-5 and 3-6 km depth), which has been classed as *Other*. There is no significant relationship between participants preferred scenario of uplift and source depth (Fig. 7.4d), although those who consider uplift to have a non-magmatic source were more likely to locate the source at *4-5 km* depth.

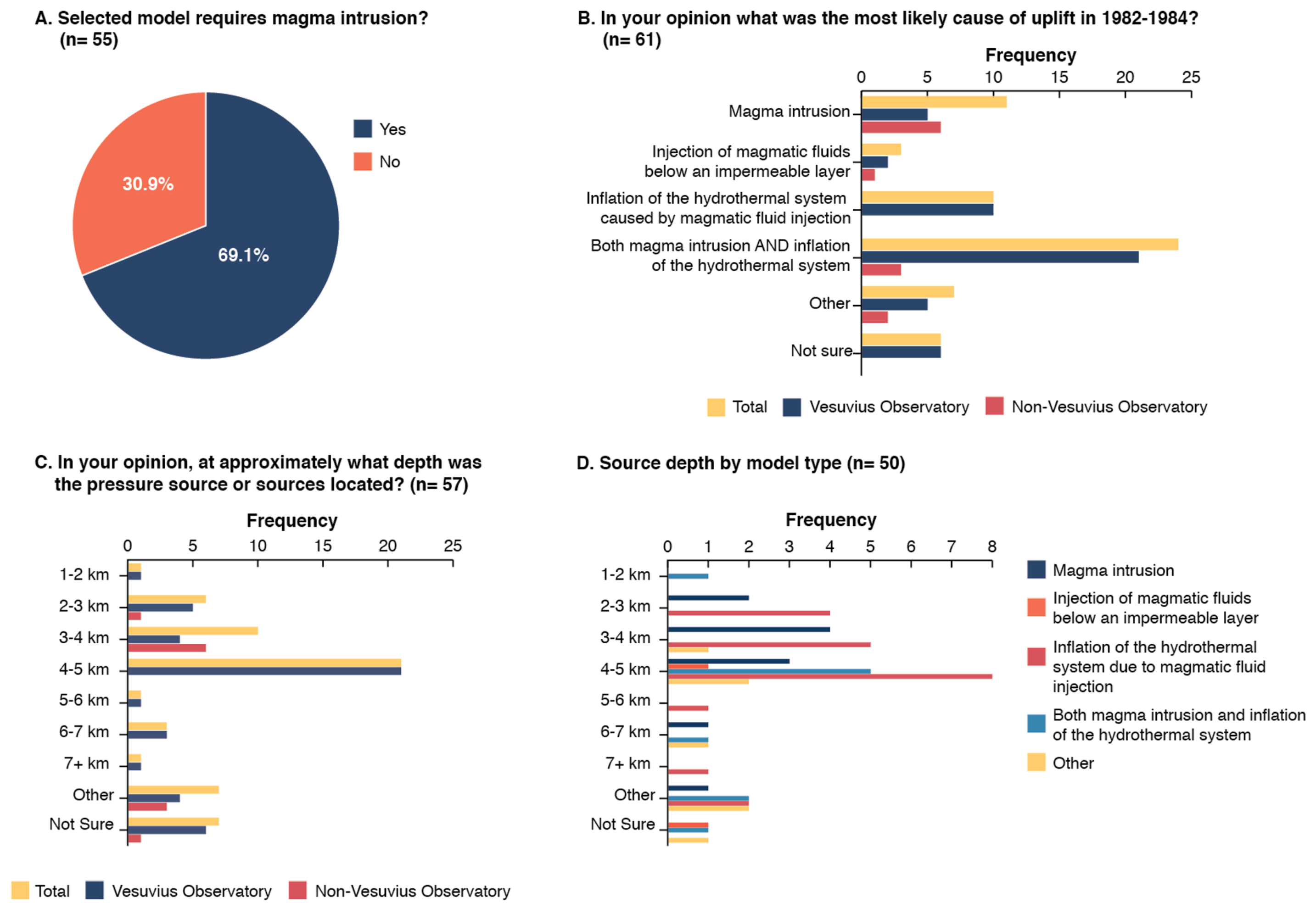


Figure 7.4: Survey participants preferred scenarios and source depths for the 1982-1984 uplift. A. Proportion of participants who consider the uplift to be partially or totally the result of a magma intrusion. B. Participants preferred scenarios for the cause of uplift. C. Preferred source depths for uplift. Responses that selected multiple depths ranges are categorised as other. D. Preferred source depths by uplift scenario.

Table 7.4: Reasons given for participants response to the question in your opinion, what was the most likely cause of the uplift in 1982-1984? The scenario column indicates the participants preferred causal mechanism of the deformation. (Mi = magma intrusion, Inf = inflation of the hydrothermal system due to an injection of magmatic fluids, Mi + Inf = both magma intrusion and inflation of the hydrothermal system, Ot = Other).

Model	Reason for Selection
Mi	Cannot be explained by fluid injection due to low permeability.
	Low permeability - quickly returned to previous state.
	Rapid uplift and geochemical modelling.
	…a sill can most realistically explain the ground deformation, seismicity and hydrothermal activity. It is also the most consistent with the history of the caldera, which is characterised by episodes of intrusion that, in most cases, do not lead to eruptions.
	Deformation pattern is not consistent with a simple hydrothermal source
Inf	The episode of 82-84 seems to show the existence of cyclicity in the geodynamics of the Phlegraean area.
	Most accredited hypothesis.
	Evidence given by significant changes in geochemical parameters in relation to seismicity and deformation. Increased CO_2/H_2O-CO_2/CH_4 and He/CH_4 ratios.
Mi + Inf	The causes are the same as those invoked for the crises 1969-1972. In this episode, a higher seismic energy release was observed, probably related to the higher rate of ground deformation.
	The uplift is permanent, i.e. it has not returned to the initial levels.
	An intrusion accounts for the localised deformation, gravimetric anomalies, distribution of seismicity, and geochemical anomalies. Intrusion triggers convection in the hydrothermal system
	These two mechanisms could justify trend in gravimetric anomalies. Seismicity.
	Presence of magmatic gas in fumaroles. Numerical models compatible with magmatic source.
	The extent of the uplift and its conservation over time is difficult to explain with just the injection of fluids and/or inflation of the hydrothermal system
	Permanent deformation, subsidence represents the hydrothermal component which is greater than in the previous uplift.
	The multidisciplinary models are convincing.
	Proven by geodetic modelling (Amoruso et al., 2014).
	Results of recent studies. The subsidence after 1984 can be explained as a flow of fluids under pressure.
	This unrest can be compared to the previous one. The larger uplift, faster deformation rate, and larger seismogenic area suggest an increase in the overpressure source and more mechanical stress in the shallow crust. …more important geochemical signals detected, greater contribution from magmatic fluids compared to the previous crisis.
Ot	Seismic and phreatic activity
	The results of recent studies, and in particular geochemical data, indicate that there was magmatic intrusion. Subsidence after 1984 indicates a decompression of hydrothermal fluids.
	Similar to the previous case. The main difference is the subsidence phase [following uplift]. … Troise et al. (2019) …provides a good explanation. … An important point: the CO2/H2O peak is after the uplift peak! CO2 starts increasing after the uplift peak, i.e. during subsidence.
	…uplift after the 1950s has only been previously observed in the last 2000 years prior to the 1538 eruption… In addition, the hydrothermal system is highly developed, so it certainly has a role to play.

Table 7.5: ***To what extent do you think the following groups are in agreement about the cause of the 1982-1984 uplift?*** **1= no agreement, 7= total agreement. Values are in % except for the Mean and Standard Deviation (SD).**

	1	2	3	4	5	6	7	Mean	SD
Neapolitan Scientists (n= 36)	2.8	2.8	5.6	19.4	33.3	30.6	5.6	4.92	1.296
Italian Scientists (n= 34)	-	2.9	5.9	26.5	35.3	26.5	2.9	4.85	1.077
International Scientists (n= 31)	-	3.2	6.5	29.0	38.7	19.4	3.2	4.74	1.064

7.2.4 Subsidence 1984-2004

57 participants responded to the question *in your opinion what was the most likely cause of subsidence after 1984?* The majority (66.7%) attributed this ground movement to a loss of pore pressure and the overall preferred scenario for subsidence was a *decrease in pore pressure in the hydrothermal system* (47.4%, n= 57, Fig. 7.5a and b). Non-Vesuvius Observatory respondents were found to be more likely to expect the involvement of magma solidification (p= 0.010), whilst those likely to advise Civil Protection in a future crisis were most likely to select the decrease in pore pressure scenario. 13 participants gave responses categorised as *Other*; 5 chose both *solidification of a magma intrusion* and *decrease in pore pressure in the hydrothermal system* and 4 selected *escape of magmatic gas and brine from below an impermeable layer into the overlying crust* and *decrease in pore pressure in the hydrothermal system*. 2 considered subsidence to result from the exhaustion of the supply of magmatic fluids into the hydrothermal system after uplift, 1 related subsidence to "*gas/fluid leakage and cooling"*, and 1 attributed subsidence to the combined effect of the *solidification of a magma intrusion* and the *escape of magmatic gas and brine from below an impermeable layer into the overlying crust*. 16 participants gave responses when asked *why* they had selected a particular scenario, which are given in Table 7.6. There are no significantly recurring themes amongst the answers although multiple responses refer to geochemical changes in fumarole gases from this period and imply that the subsidence is a necessary consequence of the previous uplift. When asked *to what extent do you think the following groups are in agreement about the cause of the subsidence* on a seven-point scale where 1 represents *no agreement* and 7 *total agreement*, the modal answer was 4, lower than for uplifts in 1969-1972 and 1982-1984 (Table. 7.7).

55 respondents answered the question *at approximately what depth(s) in the crust do you expect that the pressure source or sources were located* (Fig. 7.5c and d)? 20% of the sample selected *3-4 km* and the same number chose *4-5 km* depth. A further 20% selected *Not Sure*. As for the previously discussed uplift episodes, respondents not currently working at the Vesuvius Observatory were more likely to select *3-4 km* ($p= 0.007$), as were those participants who have more than 10 years of experience working on Campi Flegrei ($p= 0.020$). 4 respondents considered subsidence to be caused by changes in more than one pressure source. 2 located these sources at *2-3* and *6-7 km* depth, whilst a further 2 respondents located them at *2-3* and *4-5 km* depth. 4 provided a range of depths (*1-3 km*, *2-4 km* and *3-6 km*).

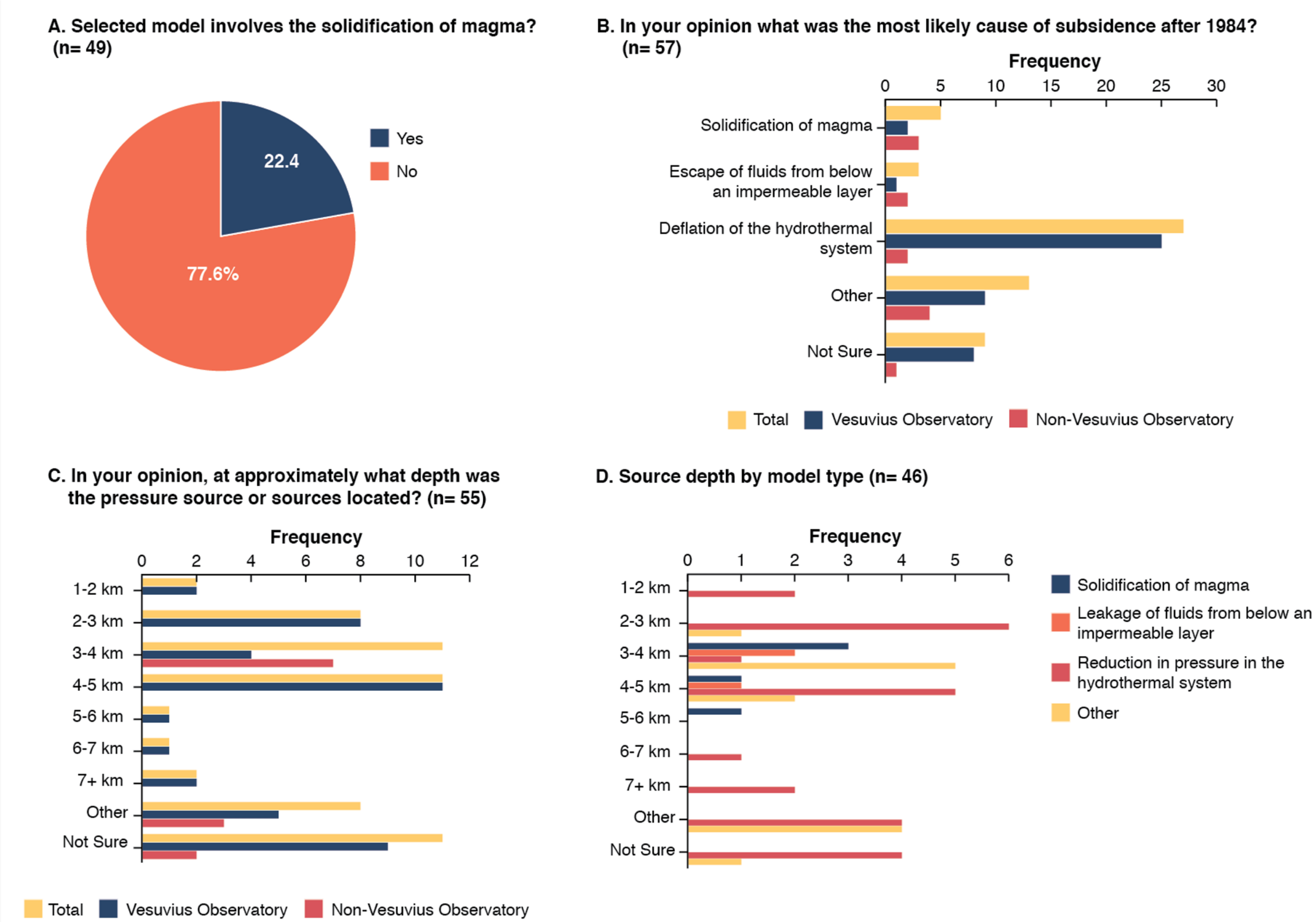


Figure 7.5: Survey participants preferred scenarios and source depths for the post-1984 subsidence. A. Proportion of participants who consider the subsidence to be partially or totally the result of the solidification of a magma intrusion. B. Participants preferred scenarios for the cause of subsidence. C. Preferred source depths for subsidence. Responses that selected multiple depths ranges are categorised as other. D. Preferred source depths by uplift scenario.

Table 7.6: Reasons given for participants response to the question in your opinion, what was the most likely cause of the subsidence after 1984? The scenario column indicates the participants preferred causal mechanism of the deformation. (Mg = solidification of a magma intrusion, Esc = escape of magmatic gas and brine from below an impermeable layer into the overlying crust, Def = decrease in pore pressure in the hydrothermal system, Ot = Other).

Scenario	Reason for Selection
Mg	…the sill spreads and thus produces a subsidence…
Esc	Geodetically proven (Amoruso et al., 2014).
Def	…there is a strong interconnection between pressurisation of the hydrothermal system/release of fluids, seismicity and ground deformation; the reduction of one is likely to result in the reduction of other parameters.
	Fluids released, system recharging.
	Loss of mass. Compaction. Large volumes of CO_2 and H_2O releases are comparable with the reduction in volume (subsidence).
	Chiodini et al. (2015).
	…significant variations in geochemical parameters in relation to seismicity and deformation. Increased ratios CO_2/H_2O - CO_2/CH_4 and He/CH_4.
	The most convincing modelling goes in that direction. Most likely the remaining deformation is due to the magmatic component.
Ot	The pressure variation produced by the magmatic intrusion is eliminated. The impulse produced by the intrusion is exhausted.
	Magma cooling + reduction in pressure in hydrothermal system.
	Aseismic deformation during subsidence is compatible with a deflation of the [hydrothermal] system due to the end of magmatic fluid supply connected with the end of intrusion and degassing of the source resulting in a reduction in pressure in the hydrothermal system.
	…subsidence begins when fracturing of the impermeable layer established a temporary connection between the lower layer characterized by lithostatic pressure and the upper one with hydrostatic pressure. The migration of fluids into the overlying porous layer causes a drop-in pressure, escape of fluids and subsidence.
	Because there was no eruption and consequent decrease in magmatic pressure.
	Subsidence (except for a very small part, max 10 cm, caused by magma degassing), cannot be explained without eruption, except by the leakage from the system and/or decrease in pressure of hydrothermal (or magmatic) fluids previously under pressure.
	See papers published by Moretti, De Natale and Troise. In addition, presence of mini uplifts (1989, 1994) without residual deformation.
	Intruded magma in the previous crisis likely spread out and cooled down. Some permanent deformation is recorded, the recovered amount of ground uplift may reflect both magma spreading, cooling and a decrease in pressure in the hydrothermal system…

Table 7.7: ***To what extent do you think the following groups are in agreement about the cause of the subsidence?*** **1= no agreement, 7= total agreement. Values are in % except for the Mean and Standard Deviation (SD).**

	1	2	3	4	5	6	7	Mean	SD
Neapolitan Scientists (n= 27)	-	3.7	3.7	44.4	33.3	11.1	3.7	4.56	1.013
Italian Scientists (n= 25)	-	4.0	8.0	48.0	28.0	12.0	-	4.36	0.952
International Scientists (n= 22)	-	4.5	13.6	36.4	36.4	9.1	-	4.32	0.995

7.2.5 Uplift Since 2004

Contrary to the previous uplifts, most of the sample consider the source of uplift since 2004 to be non-magmatic (66.7%, n= 54, Fig. 7.6a). The most commonly selected mechanism for uplift (Fig. 7.6b) was an inflation of the hydrothermal system due to an injection of magmatic fluids (36.7%, n= 60), followed by both magma intrusion and an inflation of the hydrothermal system (25%). 6 participants selected both an injection of magmatic gas and brine below an impermeable layer in the crust, and inflation of the hydrothermal system due to an injection of magmatic fluids, 1 person attributed the uplift to tectonic processes and a further 6 selected Not Sure. Justifications for the selection of a particular scenario are given in Table 7.8. Recurrent themes include the geochemical composition of fumarole gases and the slow rate of uplift relative to previous episodes. When asked *to what extent do you think the following groups are in agreement about the cause of the subsidence* on a seven-point scale where 1 represents no agreement and 7 total agreement, the modal answer was 4 (Table. 7.9).

When asked *at approximately what depth(s) in the crust do you expect that the pressure source or sources are located*, it was generally considered that the source location is shallower than for the uplift episodes in 1969-1972 and 1982-1984 (Fig. 7.6c and d). The most frequent response was 2-3 km depth (26.3%, n= 57), followed by 4-5 km (21.1%, n= 57). Those who do not work at the Vesuvius Observatory were found to be more likely to answer 3-4 km (p= 0.027), whilst those who expect to advise Civil Protection in the event of a future crisis were equally likely to select 2-3 or 3-4 km. 6 participants expect the uplift to result from the combined effect of two source, which were located at 0-1 and 3-4 km (n= 1), 2-3 and 4-5 km (n= 3), 2-3 and 6-7 km (n= 1) and 2-3 and >7 km (n= 1). 3 others gave a range of depths all of which were less than 5 km in depth.

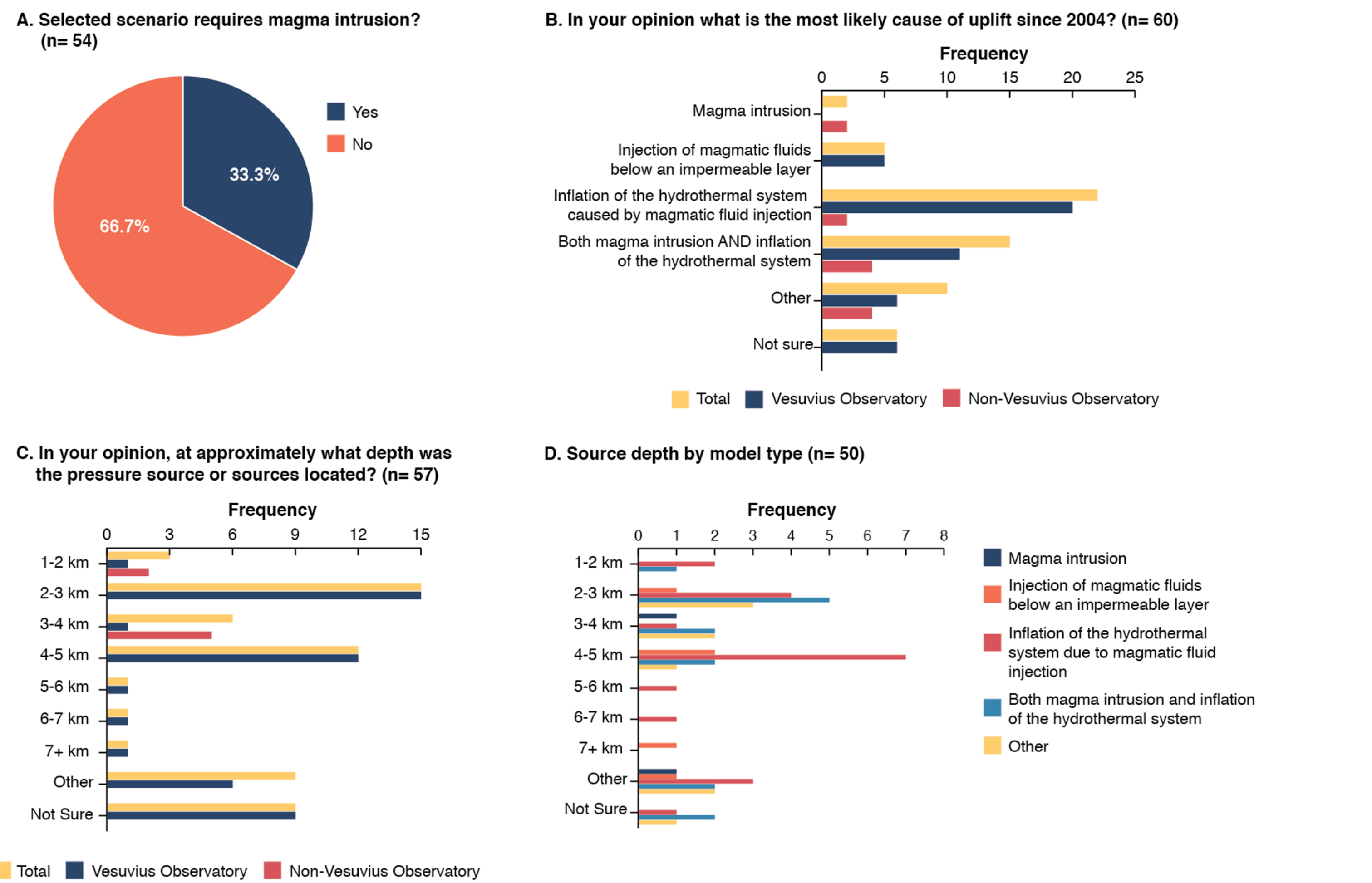


Figure 7.6: Survey participants preferred scenarios and source depths for uplift since 2004. A. Proportion of participants who consider the uplift to be partially or totally the result of a magma intrusion. B. Participants preferred scenarios for the cause of uplift. C. Preferred source depths for uplift. Responses that selected multiple depths ranges are categorised as other. D. Preferred source depths by uplift scenario.

Table 7.8: Reasons given for participants response to the question *in your opinion, what is the most likely cause of uplift since 2004?* The scenario column indicates the participants preferred causal mechanism of the deformation. (Mi = magma intrusion, Inf = inflation of the hydrothermal system due to an injection of magmatic fluids, Mi + Inf = both magma intrusion and inflation of the hydrothermal system, Ot = Other).

Scenario	Reason for Selection
Mi	Demonstrated by different geodetic studies
	Geochemistry indicates a magmatic contribution to fluid emissions. The deformation is consistent with a magmatic intrusion.
Inf	…part of the cyclicality of the geodynamics of Campi Flegrei.
	Slow rate of uplift, shallow earthquakes, gravimetric anomalies, alteration and fracturing of near surface environment.
	The involvement of magma and gas injection cannot be excluded - this hypothesis has not been sufficiently discussed
	Presence of magmatic gases in fumaroles, different characteristics of seismicity and ground deformation compared to 1982-1984
	Caliro et al. (2007; 2014), Chiodini et al. (2010; 2011; 2012; 2015; 2016)
	Evidence from the main geochemical parameters. CO_2 increase suggests an increase in the magmatic component, also increase in CO_2/H2O, CO2/CH4 and He/CH_4. Increase in CO suggests heating. Uplift is related to increase in frequency of magmatic fluid injection…
	…geochemical parameters and seismic phenomena
	Most convincing model in the literature
Mi + Inf	Uplift is almost an order of magnitude lower than the 1982-1984 crisis. …associated with the fracturing of the cap rock, which is greater than for previous crises, allowing for the emission of fluids and limiting the pressurisation of the hydrothermal system. Under these conditions the seismic energy release is low.
	Situation is more complex than for previous uplifts. Mixture of effects.
	I think there may have been a small magmatic intrusion in 2012, but in general I think there is a major disruption of the hydrothermal system.
	…the uplift rate is very different from those of past crises, this means that it is either not very viscous magma or a [different type of] fluid or a mix of these.
	Slower uplift rate, lower rates of seismicity and seismogenic volume are all consistent with a hydrothermal unrest. …the cumulative uplift is getting important, almost recovering 2/3 of the post-1984 subsidence, and starting to involve deeper seismogenic sources, as happened in 1982-84. Geochemical signals are quite straightforward, nothing similar has been observed at CF so far, suggesting a major role for magmatic fluids and magma driving the unrest. …
Ot	Slow and gradual variation of parameters, lack of deep earthquakes, absence of gravimetric anomalies. Distribution of deformation is almost unchanged.
	… local compression regime favours the formation of the lithostatic overpressure.
	… geochemical data exclude shallow magmatic intrusions. A possible alternative is a recharge of the deep magma reservoir at 8 km.
	Geochemical data exclude shallow magmatic intrusions. The only alternative is recharge of the main magma reservoir at 8 km.
	… shown that this is because of deep-derived fluids. Phenomenon likely amplified by presence of a shallow impermeable layer. …

Table 7.9: ***To what extent do you think the following groups are in agreement about the cause of uplift since 2004?*** **1= no agreement, 7= total agreement. Values are in % except for the Mean and Standard Deviation (SD).**

	1	2	3	4	5	6	7	Mean	SD
Neapolitan Scientists (n= 44)	2.3	9.1	31.8	34.1	11.4	9.1	2.3	3.8	1.250
Italian Scientists (n= 43)	-	7.0	27.9	37.2	18.6	7.0	2.3	3.98	1.123
International Scientists (n= 35)	-	8.6	31.4	31.4	11.4	14.3	2.9	4.0	1.283

7.2.6 Ground Movements Through Time

Figure 7.7 summarises the scenarios and source depths selected by respondents for each of the four deformation periods of interest. 76.7% consider the 1969-1972 and 1982-1984 uplifts to have a common source and 64.1% (n= 39) think the causative processes for uplift since 2004 are different to those for the previous episodes. Those that consider all episodes of uplift to result from the same processes (n= 15) are most likely to attribute uplifts to a non-magmatic processes. Most consider the same part of the system to be controlling subsidence and uplift after 2004 and in general participants consider the source depth for ground movements to vary through time (59.5%, n= 39). Overall, there is an agreement that the deformation source depth has shallowed over time.

Figure 7.7: Participants preferred scenarios and source locations for unrest. A. participants who assume that a ground movement is related to magma intrusion in the case of uplift and solidification of an intrusion in the case of subsidence. B. participants preferred scenarios for ground movements through time. C. participants preferred source locations for ground movements through time. Split boxes (those half coloured two ways) indicate that deformation was considered to result from the combined effect of two sources.

7.2.7 Hazard Related to Uplift Since 2004

65.6% (n= 32) of participants expect that uplift will end following an intensification of activity. 37.5% of responses mentioned explosive activity; 4 people consider a magmatic eruption to be possible, whilst 9 think that uplift could end in a phreatic eruption or hydrothermal explosion (Fig. 7.8a). 7 participants (21.9%) expect that the rates of uplift and seismicity will increase and that the characteristics of unrest will change to become like that observed in 1982-1984. 11 expect that the ground level will stabilise and either remain stationary or begin to subside, two of whom consider the uplift to result from recharge of a magma reservoir at 8 km depth and as such suggest an ignimbrite eruption is possible on a time scale of "*decades to centuries*". 1 person expects that "*Uplift will continue indefinitely. Only a significant tectonic phenomenon will change the current state*".

The most frequent responses to the question *what do you consider to be the main hazards associated with the current uplift*, were related to seismicity (40.5%, n= 37), followed by a magmatic eruption (27.0%, Fig. 7.8b). Those who expect to provide advice to the Civil Protection in the event of a future emergency were less likely to mention a magmatic eruption but more likely to provide a response that mentioned phreatic activity. Of the 6 responses categorised as *Other*, 5 were not related to physical hazards and included phrases such as; "*...panic among citizens…*", "*Inappropriate alerts or alerts based on presumptuous knowledge or beliefs*" and "*... challenge of observing cryptic volcanic unrest which results in ambiguous scientific opinion*".

The final question asked participants to locate zones of magma storage in the crust at Present (Fig. 7.8c). The most frequent depth selected was *7+ km* (30.9%, n= 55) but the majority of the sample expect the presence of magma at shallow depths, in particular at either *3-4 km* (27.3%) or *4-5 km* (20.0%). 8 participants expect that there are two main regions of magma storage; a shallow reservoir located at depths less than 5 km, and a deeper magma body at depths greater than 7 km. Those who expect to provide advice to the Civil Protection in the event of a future emergency were most likely to select *3-4 km* or *3-4 km and 7+ km*.

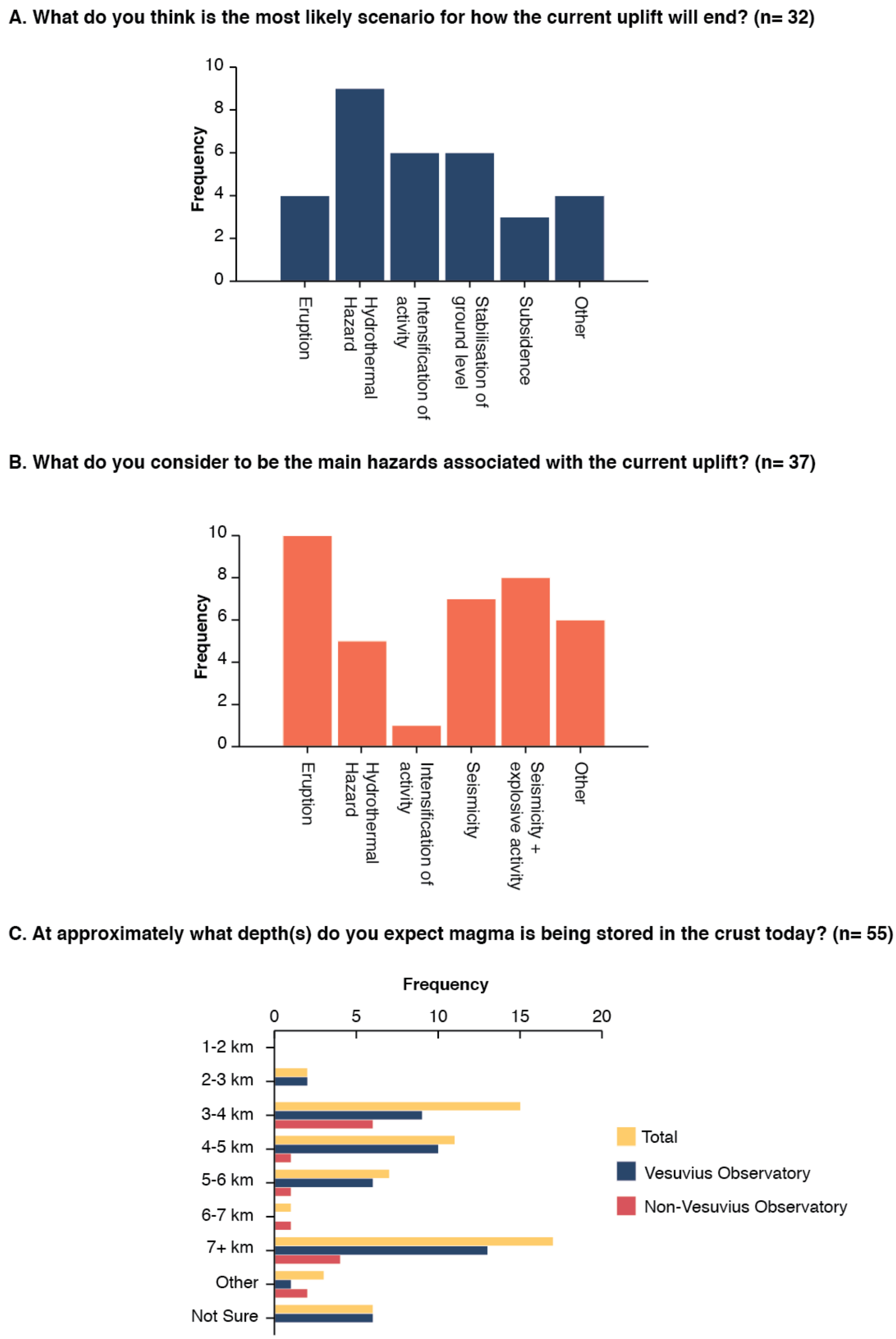


Figure 7.8: Hazard during uplift since 2004. A. scenarios for how uplift will end. B. main hazards of concern during uplift. C. participants preferred locations for magma storage in the crust. The category Hydrothermal Hazards refers to phreatic eruptions and hydrothermal explosions.

7.2.8 Indicators of Magma Intrusion

In response to the question *what changes in monitoring parameters would you expect to be characteristic of a magma intrusion at Campi Flegrei*, respondents typically stated a type of monitoring parameter rather than specific changes in that parameter. Of those that provided a response, the majority suggested a change in seismicity (83.0%, n= 47) and commonly suggested a change in some combination of the rate, magnitude and depth of events, whilst 4 respondents specified the appearance of Long-Period (LP) and/or Very Long Period (VLP) events. In total 47 participants gave a response related to deformation; most of which were related to the rate, geometry and occurrence of localised uplift. 1 participant explicitly stated an uplift rate of 0.1-0.15 m per 3 months as an indicator of an intrusion. 28 respondents (59.6%) stated that in the event of an intrusion there would be geochemical changes in fluids at the surface, 7 gave more specific responses each of which was related to changes in the concentration of CO_2 or other magmatic gases discharged from fumaroles. Responses categorised as *Other* most commonly referred to gravimetric changes, variations in gas flux or temperature changes (ground or fluid temperatures). When asked *what do you consider to be the key criteria for determining whether unrest is magmatic or non-magmatic in origin*, most stated changes in geochemistry (70.8%, n= 24). *Other* responses included variations in gravity and most frequently, “all parameters should be monitored”. 7 respondents made suggestions for aiding the differentiation between a magmatic and non-magmatic source in the event of a future unrest episode. They included; a revision of all data from the 1982-1984 unrest, InSAR and subsurface imaging using magneto-telluric, gravimetric and magnetic surveys, as well as seismic tomography.

7.2.9 Perception of Other Stakeholders

Participants were asked to rate on a seven-point scale (1= *very poor*, 7= *very good*) how well different stakeholder groups understand the behaviour of Campi Flegrei (Table 10, Fig. 7.9a). Answers for the *Civil Protection* ranged from 2 to 7, with a mean of 4.79 (n= 61, SD= 1.614) and modal score of 6, which is 1 point higher than how the sample rated the scientific understanding of the volcano. Understanding of other groups was rated between 1-7 and was lower than that for the *Civil Protection* with modal scores of 3, 2 and 2 for the *Local Authorities*, the *Public* and the *Media* respectively. When asked *in the event of a future volcanic emergency at Campi Flegrei,* *how well do you expect the following groups to understand information they receive about the volcano (1= very poor, 7= very good)*, all stakeholders were rated higher, with the understanding of the Civil Protection once again being rated significantly higher than other groups (Table 7.11, Fig. 7.9b). The modal scores for the *Civil Protection*, *Local Authorities*, and the *Public* were 6, 4, and 4 respectively. Equal numbers of respondents selected 1 and 4 for the Media.

Table 7.10: *How well would you rate the understanding of the behaviour of Campi Flegrei by the following groups*? 1 is very poor and 7 is very good. Values are in % except for the Mean and Standard Deviation (SD).

	1	2	3	4	5	6	7	Mean	SD
Civil Protection (n= 47)	-	17.0	2.1	19.1	21.3	27.7	12.8	4.79	1.614
Local Authorities (n= 47)	14.9	19.1	31.9	12.8	12.8	4.3	4.3	3.19	1.583
Public (n= 48)	29.2	35.4	14.6	10.4	4.2	4.2	2.1	2.46	1.501
Media (n= 47)	34.0	38.3	12.8	6.4	4.3	2.1	2.1	2.23	1.402

Table 7.11*: In the event of a future volcanic emergency at Campi Flegrei, how well do you expect the following groups to understand information they receive about the volcano?* 1 is very poor and 7 is very good.

	1	2	3	4	5	6	7	Mean	SD
Civil Protection (n= 53)	1.9	3.8	5.7	7.5	17.0	35.8	28.3	5.55	1.475
Local Authorities (n= 53)	9.4	7.5	5.7	24.5	22.6	17.0	13.2	4.47	1.772
Public (n= 53)	17.0	18.9	17.0	20.8	11.3	9.4	5.7	3.42	1.781
Media (n= 52)	19.2	15.4	17.3	19.2	13.5	7.7	7.7	3.46	1.852

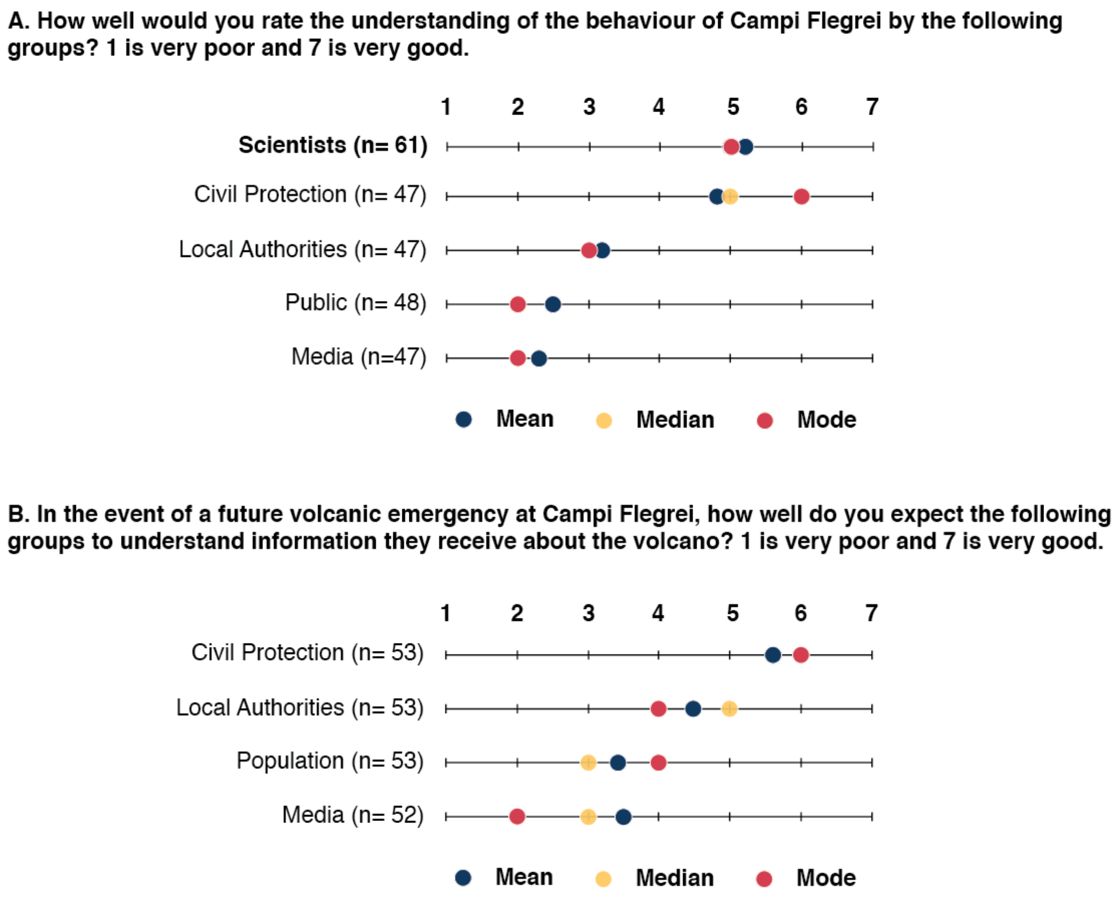


Figure 7.9: Scientists perceptions of other stakeholders understanding of Campi Flegrei. A. Comparison of how participants rated the scientific understanding of the behaviour of Campi Flegrei (Survey Sample) and that of other stakeholders. The values for scientists are the results from item 7 of the survey. The lower modal score for the scientific understanding relative to that for the Civil Protection most likely is an indicator that the scientific understanding was rated on a different mental scale to that for the other stakeholder groups. B. How well participants expect stakeholders to understand information they receive about the volcano in a future volcanic emergency.

The majority of respondents selected *Not Sure* in response to the question *what do you think the following groups think is the source of the current uplift* (uplift since 2004). Of those that gave a response there was a slight preference to select *Magma intrusion* over *Other* for each of the stakeholder groups (Fig. 7.10a). When asked *what hazards do you think the following groups are most concerned about during the current uplift*, the most frequently listed hazard was *seismicity*, although the *Civil Protection* were more generally thought to consider a magmatic eruption to be the primary hazard of concern (Fig. 7.10b). Hydrothermal hazards were significantly less likely to be referred to compared to the responses for question 18 (*What do you consider to be the main hazards associated with the current uplift?*) and were not mentioned at all in responses for the *Public* and *Media*. Participants were also more likely to give a response related to unrest management, in particular evacuation and public order, and the communication of the volcano status (categorised as *Other*) than a physical hazard. This was especially the case amongst responses related to the *Local Authorities*, the *Public* and the *Media*.

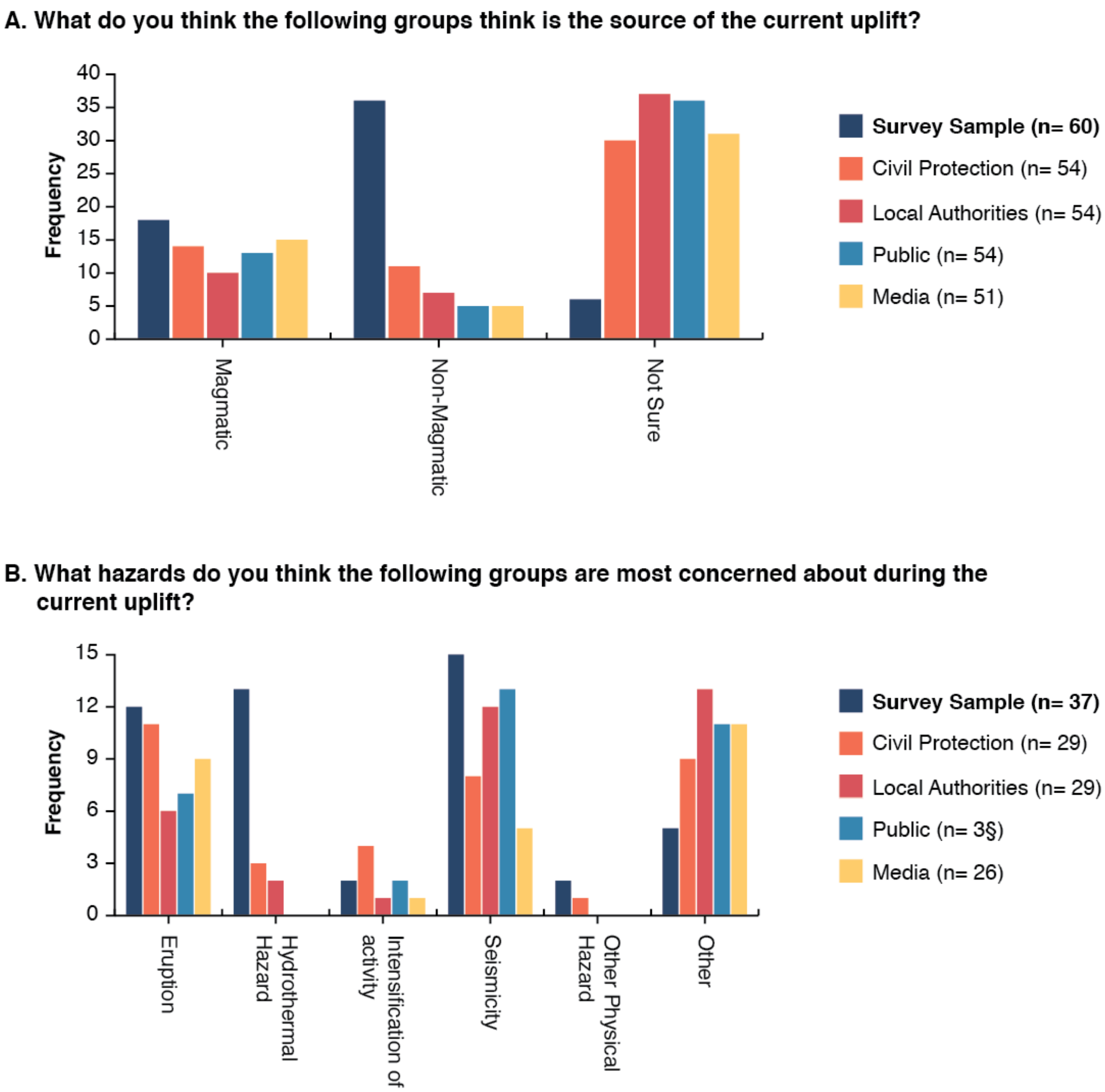


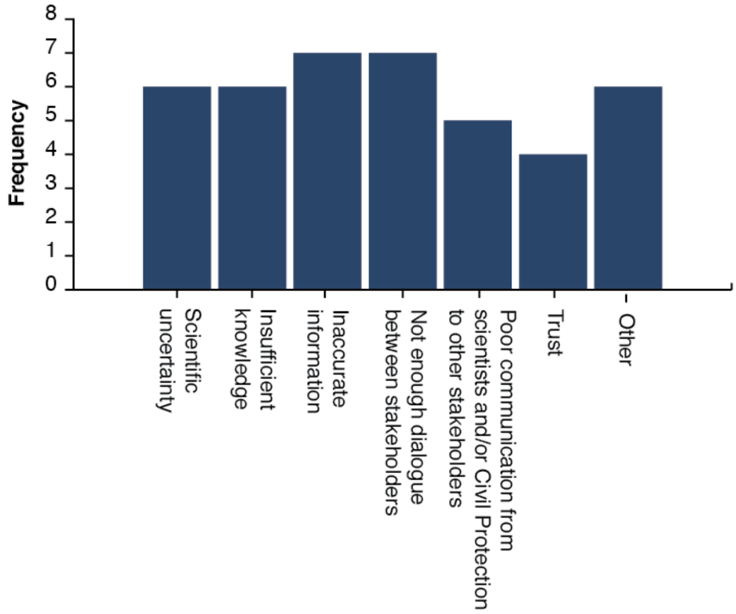
Figure 7.10: Perceptions of other stakeholder groups beliefs regarding the cause of uplift since 2004 (A) and associated hazards (B). The Survey Sample responses are those from items 15 and 18, which asked the scientists for their preferred scenario for the uplift and their greatest hazard concern during this ground movement.

7.2.10 Communication of Volcano Status

The majority of the sample think that there are currently difficulties in communicating the behaviour of the volcano between stakeholders (63.6%, n= 55). Those respondents who have experience working as a monitoring scientist during a volcanic crisis were found to be more likely to select *Yes* than those who did not (p= 0.018). 27 respondents provided examples of communications difficulties (Fig. 7.11a). Common themes that emerged included; not enough dialogue between stakeholders, poor communication between scientists and/or the Civil Protection with the media and the public, as well as a prevalence of inaccurate information, especially in the media. 1 respondent stated that the public rely on "*unreliable sources of information without clear arguments"*, whilst another said that the public and media "*interpret data themselves*". 2 others suggested that there is a lack of public trust in scientists in their responses, stating; "*In Italy in recent years there has been a strong anti-scientific feeling*" and *"… [the public] often believe that the truth is being concealed from them*". 6 participants gave responses that are related to a lack of scientific agreement regarding the behaviour of the volcano (categorised as *Scientific uncertainty*). These included phrases such as "*…there is no agreement as to the phenomena taking place in the scientific community*", "*…too many discordant voices that only fuel confusion amongst citizens*" and "*…too much contradictory science...*".

Less than half of the sample (47.3%, n= 55) expect that there will be difficulties communicating the cause of unrest and volcano status in the event of a future volcanic emergency (Fig. 7.11b), but only 5.5% were certain that there would be no difficulties. 26 respondents provided examples and whilst no one problem emerged, responses were most commonly related to scientific uncertainty (34.6%, n= 55). Conversely, 1 respondent thought that whilst communication difficulties are inevitable, there would be a consensus amongst those responsible for unrest management, and thus the public and media, that unrest is caused by magma movement. Reponses also referred to difficulties related to the quality and accuracy of information received by stakeholders. For example, the spread of misinformation amongst the local authorities and especially the public, a lack of reliable information, as well as inaccurate reporting by the media. Responses categorised as *Other* include themes such as; conflicts of interest, not being able to meet demands for information and a lack of public trust in scientists. 1 respondent suggested that public trust in scientists is lower than in the Local authorities and Civil Protection.

A. Difficulties regarding communication of the volcano's current behaviour between scientists, the Civil Protection, the local authorities, the media and the public. (n= 27)



B. Difficulties regarding communication of volcanic unrest and the volcano's status during a future volcanic emergency. (n= 26)

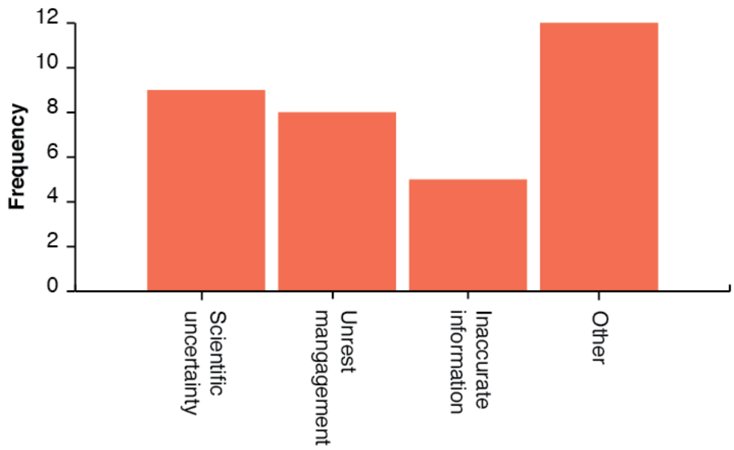


Figure 7.11: Perceived communication difficulties at Present (A) and during a future volcanic crisis at Campi Flegrei (B).

7.2 Discussion

7.3.1 Ground Movements at Campi Flegrei

The behaviour of Campi Flegrei is generally regarded by the sample to be the most well understood of the monitored Campanian volcanoes (i.e. Campi Flegrei, Ischia and Vesuvius) and, whilst individual responses were highly variable, the scientific understanding of this volcano is considered comparable to the persistently active volcanoes of Etna and Stromboli. This is perhaps surprising given that previously it has been found that the behaviour of frequently erupting volcanoes is perceived to be easier to understand (e.g. Donovan et al., 2014), and the well-established academic debate surrounding the causes of ground movement. It is considered most likely that this reflects the long-term availability of data regarding the volcano and the extensive monitoring network, as there is no association between the participants responses and the length of time they have worked on Campi Flegrei, or whether they are affiliated with the Vesuvius Observatory that may indicate a familiarity bias. The majority of the sample were in agreement as to whether the source of a ground movement was magmatic or non-magmatic for each of the four phases of ground movement addressed by the survey, and the proportion of the sample was consistent throughout (66.7-70.6%). There was no uniform view on the causative processes of each deformation period and there was no consistent association between participants responses and the indicators defined in section 7.1. Heterogeneity amongst answers relating to the preferred sources of deformation is to be expected as the sample likely contains participants who have developed conceptual models for unrest. However, the sample is large enough for preferred scenarios of unrest amongst the scientists to emerge.

7.3.1.1 Rapid Uplift in 1969-1972 and 1982-1984

There was a high degree of variability as to participants' preferred scenarios for the rapid uplift episodes in 1969-1972 and 1982-1984, with a proportion of the sample expressing a preference for each of the given mechanisms for ground deformation in the survey. Overall, however, there was a general agreement that these uplift episodes shared a common source mechanism, and most expected the uplifts to have been triggered by the intrusion of magma in the region of 3-5 km depth, and that a component of deformation was related to a pressurisation of the hydrothermal system in response to an injection of magmatic fluids during magma intrusion. Such a scenario for unrest is consistent with the group of models discussed in Chapter 3, section. 3.1.1.

Justifications for a two-component source included the observation of a minor subsidence following uplift and the results of modelling of the evolution of the 1982-1984 uplift. Similarly, justifications for a two-component source for the 1982-1984 episode included net-uplift and the results of modelling. In both cases then, the preferred scenario for the causative processes of uplift appears to have emerged over time based on observations of the ground level following uplift, rather than distinct changes in monitoring parameters during the unrest episodes.

The models described in Chapter 3, section 3.1.1, consider subsidence to result from a pore pressure loss in the hydrothermal system following its pressurisation during the preceding uplift. In this context, the minor subsidence following the 1969-1972 uplift compared to that post-1984, suggests that the hydrothermal system was pressurised to a lesser extent. A limitation of the survey is that it cannot be concluded as to whether the implied relative pore pressure increase during each of the uplifts is related to differences in the volatile composition of the magma intrusion in each case, the relative locations of magma intrusions during each episode, or a change in the hydraulic properties of the crust.

7.3.1.2 Ground Movements after 1984

Overall, there is a good agreement amongst the sample that the end of uplift in 1984 marks a change in the controls on ground movements. The preferred scenario for subsidence is a decompression of the hydrothermal system, whilst uplift since 2004 is considered by most of the sample to result from renewed pressurisation caused by injection of magmatic fluids. Despite this general agreement, and slightly less variation in participants responses as to the causes of these ground movements compared to those for deformation prior to 1984, participants perception of the level of scientific agreement as to the processes controlling post-1984 ground movements was slightly lower. This perhaps reflects a greater uncertainty as to the origins of the current uplift, and salience of the academic debate as to its cause in recently published articles in both scientific journals and the press.

Contrary to the preferred mechanism for the uplifts in 1969-1972 and 1982-1984, there is a general agreement that magma has not been intruded into the crust since 2004, so that in the preferred scenario for the current uplift, the magmatic fluids entering the hydrothermal system must instead originate from a pre-existing zone of magma storage (Fig. 7.12). When asked where magma is currently located in the crust approximately two thirds of the sample located a magma body at shallow depths, with preferred storage locations at 3-4 km and 4-5 km depth. Those in the sample that expect to advise the Civil Protection in the event of a future volcanic emergency, were most likely to consider magma to be present at 3-4 km depth. The general assumption of the presence of a shallow body and a pressurisation of the hydrothermal system resulting from an input of magmatic fluids, suggests that there may be an emerging preference for a conceptual model for the current uplift like that of Chiodini et al. (2015a; 2016). In this scenario, uplift is a consequence of an acceleration in degassing of a magma body located between 3-4 km depth that was intruded in 1982-1984 as it decompresses (Chiodini et al., 2016). As discussed in Chapter 3, the presence of a magma body at such depths is based upon changes in fumarole gas chemistry that can be alternatively interpreted without requiring the presence of shallow magma at Present (e.g. Caliro et al., 2007; Moretti et al., 2017; Chapter 6). Of note is that multiple participants provided responses stating that the cause of the current uplift has not yet been "*adequately debated*".

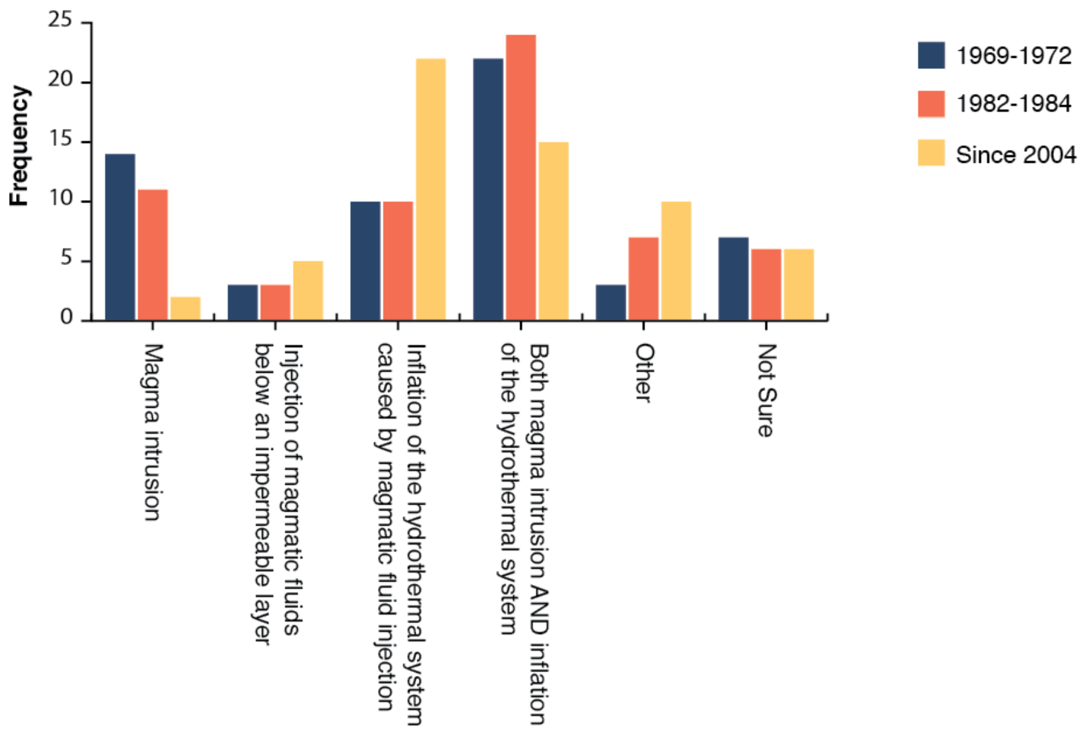


Figure 7.12: Preferred mechanisms for uplift episodes since 1969. The most frequent selections for each episode require an increase in pressure in the hydrothermal system.

Amongst the sample a range of mechanisms for the current uplift were given, each of which has differing implications for hazard. In the extreme cases, one participant stated that the deformation results from tectonic, rather than volcanic or hydrothermal, processes, whilst a minority consider it to be an effect of recharge of a magma reservoir at 8 km depth, and a precursor to an ignimbrite eruption on a timescale of "*decades or centuries*". In the Chiodini et al. (2015; 2016) model, a consequence of the processes controlling uplift is the probability of an eruption occurring increases as the deformation proceeds. Amongst the sample however, only four participants think that the uplift may end in an eruption. Instead there is a preference for an intensification of activity resulting in a phreatic eruption, or alternatively, either a stabilisation of the ground level or return to subsidence conditions without any significant escalation in unrest. Some also consider the uplift to be a precursor to more intense unrest "*like in 1982-1984*". In all cases those that mentioned phreatic activity consider the hydrothermal system to be becoming pressurised, suggesting a belief that the shallow system is also pressurising, in line with the conclusions of Chiodini et al. (2015; 2016) and contrary to those of Moretti et al. (2017). Previously, it has been suggested that phreatic eruptions at Campi Flegrei require the intrusion of magma into the upper few hundred metres of the crust (e.g. Italiano et al., 1984) and there are a no confirmed historical reports of phreatic eruptions, or significant explosive hydrothermal activity, except for the mud fountaining event following the 1930 Irpinia earthquake that only impacted the immediate surroundings (Signore, 1935).

It was found that the understanding of the behaviour of Campi Flegrei by the Civil Protection was generally perceived to be good, whilst that for the other stakeholder groups of interest was considerably lower (i.e. the Local Authorities, the Public and the Media). There was also a perception amongst the scientists that these groups views on the cause of the current uplift and their main hazards of concern differed from their own. The Civil Protection were considered to be most concerned about an eruption, most likely reflecting their operational and response planning priorities. The Local Authorities and the Public were felt to be most concerned about the seismic risk associated with unrest. Experience of past events influences risk perception (e.g. Slovic, 2000; Paton, 2000) and the scientists' perception of these groups perceptions are in agreement with a risk perception study of the population of Campi Flegrei conducted in 2006 that found hazard knowledge and salience to be low, but that respondents were more likely to consider bradyseismicity (i.e. unrest as in the 1969-1972 and 1982-1984 episodes) and seismicity to present a risk than other hazards (Ricci et al., 2013b).

A common hazard concern amongst the scientists was the potential for a phreatic eruption or hydrothermal explosion, but only a minority of responses (5 where n= 29) for the stakeholders referred to such phenomena, and those that did were related to the Civil Protection or the Local authorities. Whilst the impacts from such hazards tend to be low, the most likely location of such activity is Solfatara-Pisciarelli, a volcano-tourism site located in a populated area. Given the general belief that the hydrothermal system is pressurising, implicit in which is that the hydrothermal hazard is increasing, then this may indicate that work is required to increase the general awareness of these hazards.

7.3.1.3 Future Unrest

Overall, it was felt that the characteristics of seismicity and uplift are the most useful parameters for identifying magma on the move, but that changes in fumarole gas chemistry were the most useful for differentiating the involvement of magma during an unrest episode. This is perhaps surprising given that central to the current debate as to the cause of uplift since 2004 are alternative interpretations of the same geochemical trends. Inherent in any interpretation of future unrest will be uncertainty and ambiguity. As such, in the event of an intensification of unrest, effective communication will be a critical factor in determining whether the stakeholder response is realistic.

Significantly, most of the sample thought that there were currently communication difficulties regarding the behaviour of the volcano between stakeholders and only 5.5% (n= 55) were confident that there would not be major communication challenges in the event of a future emergency. Approximately a fifth of participants consider current difficulties to stem from scientific uncertainty regarding the cause of uplift since 2004. One respondent specifically stated that a lack of consistent message from the scientific community was “*fuelling confusion*” amongst the public. This may affect people's perception of the relationship between unrest and risk, and potentially their response to a future intensification of activity. Additionally, there is a risk that the perceived competence of scientists will be negatively impacted. This is important as it affects peoples trust in those responsible for making expert judgements, and in turn how they evaluate risk information that they receive (Poortinga and Pidgeon, 2003; Haynes et al., 2008; Wachinger et al., 2013). Whilst the survey was not designed to measure perceived levels of trust between stakeholders, it is notable that one respondent suggested that the public trust scientists and the Civil Protection (i.e. those most likely to be involved in informing and making decisions regarding unrest management) less than the local authorities, contrary to the results of Ricci et al. (2013). Identifying if this is an accurate perception and whether a decline in the trust of these groups has occurred since this study, is something that should be a priority for future work. Responses from other participants included "*people often believe that the truth is being concealed from them*", something that had been directly addressed through statements from the Vesuvius Observatory Director on their website, and that in Italy a "*strong anti-scientific feeling has developed*", so that trust may also be being impacted by the wider social context.

A relationship is apparent amongst the responses as to the perceived understanding of Campi Flegrei and stakeholders identified in responses regarding difficulties in communicating the status of the volcano. Only one participant mentioned communication between scientists and the Civil Protection, reflecting the general confidence in this groups understanding of the caldera. Most often responses referred to the public and the media, both of whom are considered to have a poor knowledge of the volcano. Salient amongst responses include difficulties related to the communication of information from scientists and the Civil Protection to these groups, the perceived prevalence of misinformation amongst the media and public, as well as a public reliance on unofficial sources. A key agent in the circulation of misinformation was identified as the media, who were generally considered to sensationalise scientific research and to be likely to amplify risk through sensationalism of information in the event of a future emergency. The influence the media can have in shaping public risk perception, both positive and negative, is evident from numerous volcanic crises, including past unrest episodes at Campi Flegrei (Barberi et al., 1984). The effect of sensationalist reporting of the volcano's current behaviour and possible evolution, as well as the academic debate surrounding it, has the potential to undermine trust in scientists and the Civil Protection, as well as influence the public's response to a future emergency.

7.4 Conclusions

The majority of monitoring scientists and those that may advise the Civil Protection in the event of a future volcanic crisis at Campi Flegrei are in general agreement that (i) unrest in 1969-1972 and 1982-1984 was magmatic, (ii) that post-1984 ground movements require changes in the supply of magmatic fluids into the hydrothermal system, and that (iii) uplifts require the transport of fluids from the magmatic system to shallower depths. No uniformity exists amongst the sample as to causative processes of ground movements, but there exists a preference amongst the sample for a combined magma-hydrothermal source of rapid uplifts before 1984, where the hydrothermal system amplifies deformation due to magma. Uplift since 2004 is generally considered to be related to a pressurisation of the hydrothermal system, and that magma currently exists at shallow depths in the crust. Despite this the potential eruption hazard associated with the current uplift was low, and the seismic and hydrothermal hazards are considered to be of greater concern.

Finally, it was found that there was a perception that communication difficulties between different stakeholder groups may have emerged as a result of the academic debate surrounding the current uplift and that scientists consider there to be a prevalence of misinformation amongst the public. There was also a clear perception that the media actively seek controversy and to amplify risk. As such, it is suggested that the greatest implications of scientific uncertainty during the current uplift may be for public perception of risk and trust in scientists, rather than for operational processes. This may be of critical importance in a future emergency.

Chapter 8

Conclusions

This chapter concludes the thesis by discussing the findings of the research in relation to the initial research question and aims outlined in Chapter 1. It addresses the specific research questions that guided the thesis and then suggests avenues for future research.

Chapter 1 of the thesis provided background information relating to caldera magmatic-hydrothermal systems and decadal unrest. Chapter 2 established the geological context from which the rest of the thesis followed. In Chapter 3 conventional models of post-1984 ground movements and their limitations were reviewed, then in Chapter 4 the methodology used to develop a new conceptual model for deformation after 1984 was outlined. The known recent (20^{th} Century onwards) behaviour of the caldera was reviewed in Chapter 5 and then integrated with the results of Chapter 2 in order to develop the conceptual model presented in Chapter 6. Chapter 7 is a self-contained chapter that examines scientists' perceptions of the causes of ground movements at Campi Flegrei.

8.1 Addressing the Research Aims

8.1.1 Research Aim 1

There is no agreement as to the structure of the crust below the deforming area amongst conventional models of unrest. As such the first research aim was to *'Establish a model of the structure of the magmatic-hydrothermal system and the location of hydrothermal reservoirs based on the current knowledge of the Campi Flegrei subsurface and hydrothermal activity'*. This aim was addressed by reviewing geological, geophysical and geochemical studies of Campi Flegrei, which were summarised in Chapter 2 and used to derive a schematic model of the crust. This model was adopted throughout the rest of the thesis. The review was guided by the research questions:

a. Where in the crust are magma bodies, magmatic fluids and hydrothermal fluids known to be located in the crust?

b. Are the positions of hydrothermal reservoirs stable through time?

Following classical models of magmatic-hydrothermal systems (e.g. Burnham, 1979; Fournier, 1999) it was found through the review that the Campi Flegrei system can be divided into a magmatic regime, where fluids are dominantly magmatic and pore pressures approach lithostatic, and an overlying hydrothermal regime where principally meteoric fluids circulate at close to hydrostatic pressures. The magmatic regime extends from 3 km depth, contains a mid-crustal reservoir of magma at c. 7.5 km depth and an overlying storage zone of magmatic fluids (Zollo et al., 2008). No evidence was located in the literature for significant volumes of magma storage at shallower depths. In the hydrothermal regime (0-3 km depth) two principal areas containing hydrothermal reservoirs were distinguished: at Mofete in the west, and below Pozzuoli-Agnano in the main deforming part of the caldera. The locations of the reservoirs and outflows at the surface are controlled by the presence of fractures, which act as fluid pathways. The essential features of the hydrothermal system below Pozzuoli were defined as (i) a reservoir zone where magmatic and meteoric fluids mix located between c. 1.5-3 km depth, (ii) a caprock of hydrothermally altered clays that limits flow to the surface, and (iii) a near vertical, pervasively fractured volume that connects the main reservoir zone to the surface, which is generally referred to as the Solfatara-Pisciarelli Diffuse Degassing Structure (DDS). Based on the comparable distribution of Roman thermal baths and present-day hydrothermal features it was concluded that the position of the hydrothermal reservoirs feeding activity is stable over millennial timescales.

8.2.2 Research Aim 2

To establish constraints for a new model of unrest the second research aim was to *'Determine whether long term trends in monitoring parameters and observations of hydrothermal activity at the surface indicate any changes in conditions in the hydrothermal system'*. The research questions guiding this aim were:

c. What is the relationship between the deforming area though time and the location of hydrothermal reservoirs?
d. Does the relationship between deformation and other monitoring parameters change significantly through time?
e. Did hydrothermal activity at the surface change significantly during and between uplift episodes?

These questions were addressed in the review of the behaviour of the caldera in Chapter 2. This found that the deforming area has been centred on the region of the crust that the assumed model expects to contain hydrothermal reservoirs below Pozzuoli throughout time, regardless of the direction of ground movement. Deformation is therefore likely to be accommodated in part by structures that influence fluid flow in the hydrothermal reservoirs and provides a mechanism for generating permeability in the hydrothermal system over time. Through the review it was determined that the relationship between deformation and rates of Volcano-Tectonic (VT) seismicity changed significantly between uplift episodes. In particular VT seismicity rates increased exponentially during the 1969-1972 and 1982-1984 rapid uplift episodes. Given that VT seismicity is a proxy for the amount of brittle deformation (i.e. fracturing and faulting, Kilburn, 2012), the crust at the end of uplift in 1984 must have been more fractured than at any point previously since the onset of unrest in 1950. A second notable relationship identified is that between geodetic signal and enrichment of CO_2 in Solfatara fumarole gases. During the 1982-1984 uplift an increase in CO_2 concentrations lagged the change in ground level, suggesting it was caused by the geophysical changes in the crust (e.g. Chiodini et al., 2003). In contrast the CO_2 enrichment preceded the start of the most recent uplift since 2004. From this it was inferred that the cause of the geochemical changes is driving deformation and that the processes controlling this uplift cannot be the same as those for previous episodes.

It was found from the review that the distribution of surface activity has not changed significantly over successive uplifts, so that any changes in fluid flow paths have occurred without a major hydrological reorganisation. It was not possible to compare the intensity of changes in surface activity between phases of ground movement as visual observations have not been regularly reported through time. However, it was found that geochemical changes in thermal waters and fumaroles, together with increased degassing and enlargement of mud pools at Solfatara are compatible with increased gas transport through the Solfatara-Pisciarelli DDS during both the 1969-1972 and 1982-1984 uplifts. Increased gas transport to the surface at Solfatara has also been measured during the most recent episode, suggesting that uplifts are associated with enhanced gas flux. Whilst it cannot be established otherwise due to the lack of observations, no evidence was found to suggest that the hydrothermal system was pressurised to a greater extent during the 1982-1984 uplift than the 1969-1972 episode. This is an inherent assumption of conventional models of post-1984 ground movements that consider subsidence after 1984 to be caused by a decrease in pore pressure in this part of the crust (e.g. De Natale et al., 1991; Gaeta et al., 1998; Orsi et al., 1999a; Todesco et al., 2003; De Natale et al., 2006; Battaglia et al., 2006; Gottsman et al., 2006; Troiano et al., 2011; Chiodini et al., 2015a; Moretti et al., 2017) but not therefore a constraint.

8.2.3 Research Aim 3

The third research aim was to *'Re-interpret long-term trends in monitoring data and observations of surface activity to produce a model that can account for the full sequence of ground movements since 1950'*. This was addressed through integrating the answers to previous research questions with the following.

f. What is the control on pore pressure in the hydrothermal system?
g. Are there any long-term changes in the pore pressure control and if so, where in the crust are these changes occurring?

It was found through review of the literature that permeability is the primary control on pore pressure distribution in the crust and that typically in caldera hydrothermal systems the bulk permeability is controlled by the density and characteristics of fractures. In the absence of direct observations in the reservoir zone below Pozzuoli this parameter was assumed to be fracture controlled. This was based on the dependence of the location of surface activity on fractures, the correspondence of this part of the caldera with a high density of faults and fractures, published interpretations of a seismic anomaly interpreted as a pervasively fractured volume saturated in fluids (e.g. Aster and Meyer, 1988), and observations from boreholes that intersect hydrothermal reservoirs in the second main area of hydrothermal activity at Mofete, which were found to be located in fracture zones (Rosi and Sbrana, 1987; AGIP). Furthermore, it was found in the literature that geochemical modelling of fluids that feed fluids at Solfatara circulate in the crust at temperatures of c.200-400 °C (Cliro et al., 2007). Hydrothermal alteration at these temperatures is in the calc-aluminium silicate and thermometamorphic zones (Browne, 1978). As such the formation would be expected to be sufficiently brittle to maintain fractures.

It is known from Chapter 2 that the long-term trend in VT-deformation is compatible with an increase in fracturing in the crust over successive uplift and it was found through the review that conditions in the crust in 1984 were favourable for widespread fracturing large enough to alter the bulk permeability of the crust. A spatial analysis of the distribution of seismicity through the crust relative to the position of a seismic anomaly used to define the position of the hydrothermal reservoirs below Pozzuoli (after Aster and Meyer, 1988) and the distribution of seismic b-values, determined that fracturing was most likely concentrated in the hydrothermal reservoir zone. Focal mechanisms located in the literature confirmed deformation in extension. As such it was concluded that in 1982-1984 there was an increase in the permeability of the main reservoir zone. Changes in fluid flow and a redistribution of pore pressures would necessarily follow, resulting in deformation.

8.2.4 Research Aim 4

The final research aim was to *'Ascertain if a preferred mechanism for ground movements exists amongst scientists who may be involved in a response to a future intensification of unrest, their expected evolution of uplift since 2004 and perceived challenges in communicating unrest'*. To meet this aim the following questions were answered:

h. Is there a preferred scenario for the controls on ground movements at Campi Flegrei amongst scientists?
i. Is there an agreement as to where magma is currently located in the crust and the expected evolution of ongoing uplift since 2004?
j. What, if any are the perceived challenges of communicating unrest between stakeholders at Present (2019)?

Through a survey of a sample of scientists based at the Vesuvius Observatory and known from the literature to have worked on Campi Flegrei, it was found that whilst there was no uniform view of the controls on ground movements exists, a preferred scenario for deformation did emerge amongst the sample for each period addressed by questions (uplifts in 1969-1972 and 1982-1984, post-1984 subsidence and uplift since 2004). There was a general agreement that uplifts in 1969-1972 and 1982-1984 were magmatic in origin and that deformation after 1984 was controlled by non-magmatic processes. For uplifts in 1969-1972 and 1982-1984 deformation was most commonly considered to be triggered by magma movements and to be related to the pressurisation of both magma and the hydrothermal system. Post-1984 ground movements were most commonly considered to represent a depressurisation and a re-pressurisation of the hydrothermal system. The re-pressurisation was most commonly considered to be due to an increase in the transport of magmatic fluids into the hydrothermal system, compatible with the model proposed for uplift since 2004 by Chiodini et al. (2015a; 2016). Most participants were also found to be in agreement that there is a zone of magma storage at shallow depth (<5 km) at Present (2019), which contrasts with the assumed model of the crust adopted here. The preference for a shallow magma body was not found to increase the likelihood that a respondent expected uplift since 2004 to be precursory to an eruption. Instead most consider that this ground movement will end with explosive hydrothermal activity as the hydrothermal system is increasingly pressurised or for the ground level to stabilise without any significant intensification of activity. It was found that there was a general perception that communication of the volcano's status during current unrest could be improved. The most salient challenge to communicating about unrest was the prevalence of inaccurate information available to the public. Some participants also highlighted scientific uncertainty and a lack of agreement as to the cause of uplift since 2004 as an impediment to effective communication between stakeholders.

8.2 Addressing the Initial Research Question

The initial research question guiding this thesis was ' *Is the long-term deformation sequence at Campi Flegrei since 1950 compatible with a single evolutionary sequence where the contribution of the hydrothermal system changes over time, and what is its perceived role in ground movements amongst those who may be involved in the scientific response to a future intensification of unrest?*'.

Changes in the long-term deformation profile since 1950 are concluded to be consistent with a single evolutionary sequence where the contribution of the hydrothermal system is controlled by dynamic changes in permeability. It was recognised that the occurrence of VT seismicity during the 1969-1972 and 1982-1984 uplifts was an indicator of permeability generation in the crust by fracturing and faulting and that the pore pressure signal after 1984 emerged after conditions in the crust were met that were favourable for an increase in bulk permeability in hydrothermal reservoirs below Pozzuoli by widespread fracturing. The timing of the inferred bulk permeability change is compatible with the observed increase in brittle deformation as a result of the repeated stretching of the crust over successive uplifts between 1950-1984. By considering the location of the expected permeability change and potential impacts on flow paths, it was determined that a likely impact of fracturing was to increase hydraulic connectivity in the hydrothermal reservoir, as well as between it, the surrounding caldera and the surface. Increased transport of fluids out of the reservoir zone was then inferred, resulting in its depressurisation and subsidence. It was also recognised that a natural consequence of a pore pressure loss over time, in combination with processes of permeability destruction known to operate over the relevant timescales (e.g. crack closure due to pore pressure loss and fracture sealing by mineral precipitation and hydrothermal alteration) would result in a net-loss of permeability over the duration of subsidence. On the basis that there is no evidence that the background supply of meteoric and background fluids to the reservoir zone was permanently disturbed by uplift in 1982-1984, the permeability loss would favour an eventual recovery of pore pressures. The presence of a shallow body of magma or an increase in magmatic fluid flux into the hydrothermal system is therefore not required for the current uplift to proceed, and in the context of the proposed model the eruption hazard is at its lowest since 1982-1984. This conclusion is in direct contrast with the those of existing models. As such, a recommendation of the thesis for improving assessment and communication of unrest is the formalised definition and evaluation of end member scenarios.

The second part of the research question was addressed through surveys of scientists. It was found that whilst there was no uniform agreement as to the causes of unrest, there was a general view that pressurisation of the hydrothermal system amplified deformation due to magma intrusions in 1969-1972 and 1982-1984. Overall, a hydrothermal control was favoured for ground movements post-1984 and the deformation during this period was most commonly considered to reflect changes in pore pressure in this part of the crust. Subsidence was considered to reflect pore pressure loss following pressurisation of the hydrothermal system during the preceding uplift, whilst pressurisation after 2004 was attributed to an increase in magmatic fluid flux. There is therefore a general perception that the involvement of the hydrothermal system in ground movements is dependent on prior changes in the magmatic system.

8.3 Future Work

On the basis of the results of the thesis, the following objectives have been identified for further research:

i. Quantification of the model proposed in Chapter 6 through evaluation of the magnitudes of permeability changes required to achieve the observed deformation using computer modelling.

ii. Identification of the existence of analogous long-term unrest at other calderas through a review of other systems to validate the model and establish if it may be applied elsewhere.

iii. Improved characterisation of fluid paths and therefore the permeability of the caldera over time through monitoring of hydrothermal features outside of the main degassing area at Solfatara-Pisciarelli and mapping of soil CO_2 degassing.

iv. Investigation of the potential of developing low-cost sensors from commercially available off the shelf components for monitoring changes in hydrothermal activity over large areas of the caldera.

v. Application of social science methods to investigate the perception and communication of unrest between different stakeholder groups.

vi. Application of social science methods to develop communication products to aid the communication of the hazards associated with long-term unrest and uncertainty.

Chapter 9

Appendices

(A) Thermal water chemistry analysis

(B) Distribution of seismicity in relation to the hydrothermal system

(C) University College London (UCL) Research Ethics Committee application and approval to conduct low risk research.

(D) Surveys and participant information sheets for data collection used in Chapter 7

(A) Temporal Trends in Thermal Water Chemistry

A dataset of temperatures and chemical compositions of thermal waters at Campi Flegrei was collected from published data, in an attempt to identify relationships between changes in thermal water characteristics and ground movements that may indicate long-term changes in the hydrothermal system.

The following questions were defined:

i. Do the physico-chemical characteristics of thermal waters change in response to ground deformation and seismicity? Do these changes precede or lag the changes in physical parameters?
ii. Are the changes in waters at a location the same between different uplift episodes?
iii. Can long-term changes in the feeder pathways be identified?
iv. Did samples from the 1969-1972 unrest have a higher enthalpy than in 1982-1984, as stated in the literature?

The dataset consists of major, minor, and trace element concentrations, and isotopic compositions from surface waters across Campi Flegrei (Fig. 9.1). Time series are highly discontinuous, with only a few data points available from before 1983, for a few locations. The data was first used to classify waters (Fig. 9.1 and 9.2) according to the dominant anion present (Cl, SO_4 or HCO_3) over time and were in general agreement with classifications in Valentino and Stanzione (2003) and Aiuppa et al., (2006). Locations without significant multi-year time series were then discarded. Temporal trends in all geochemical parameters in the dataset were then compared with changes in the ground level and rates of Volcano-Tectonic (VT) seismicity between 1970-1999 (the dates of the earliest and most recent samples). The paucity of the dataset and discontinuous nature limited analysis and interpretation. However, the following was established:

i. All analysed locations showed changes in the characteristics of waters in response to changes in geophysical parameters during the 1982-1984 unrest. The changes were not consistent between locations and were of varying intensities.

ii. Data was insufficient to determine if systematic variations were repeated between the 1969-1972 and 1982-1984 unrest episodes.

iii. Temporal trends at Terme Puteolane indicate an input of magmatic gas rich fluids in the 1969-1972 unrest. At this location the Na Cl rich component of the waters also became more important from the 1982-1984 unrest. Permanent long-term changes could not be identified elsewhere. The greatest variations during the 1982-1984 unrest were observed at the locations closest to Solfatara at Pisciarelli and Hotel Tennis. These waters are influenced by the Solfatara plume and show an enrichment in steam and magmatic gases during the unrest, followed by a depletion during subsidence after 1984, as observed in fumarole gases. Only waters at Terme Puteolane and Hotel Tennis showed any significant changes in anion concentrations overtime. In both cases the changes suggest an input of magmatic gas-rich steam during the 1982-1984 unrest.

These observations confirm observations made in the references given in Chapter 4, which are discussed in Chapter 5. It was not possible to establish if the enthalpy of fluids was higher during the 1969-1972 unrest relative to during the 1982-1984 unrest. The temporal trends in anion concentrations and the physical characteristics of waters for the wells discussed in Chapter 5 (Hotel Tennis, Stufe di Nerone and Terme Puteolane), which showed the largest variations, are given in the following pages.

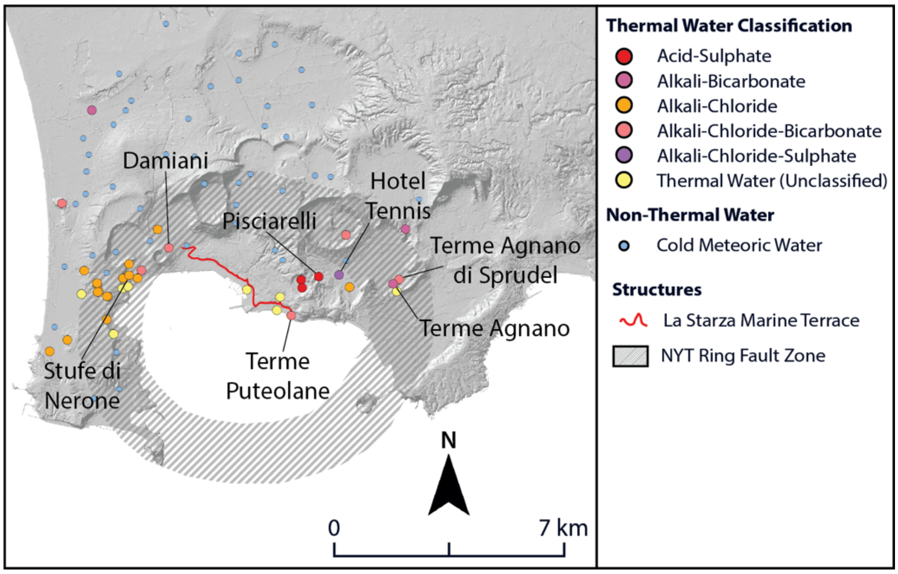
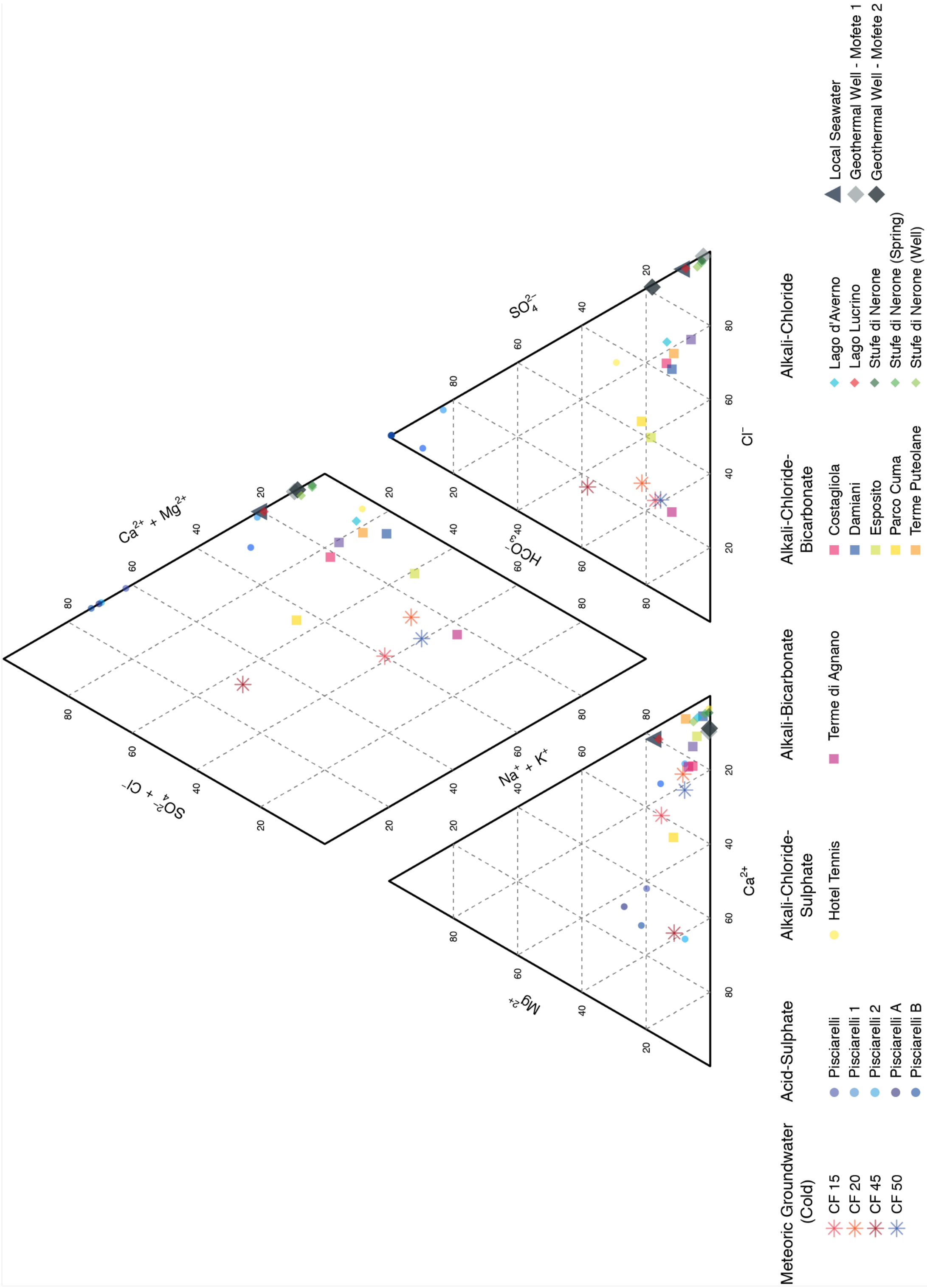


Figure 9.1: Water tyoes across Campi Flegrei as categorised from the dataset collated from the literature. Labelled locations are those where data is available over multiple years. Cold Meteoric water locations from Aiuppa et al. (2006).


Ca2+ + Mg2+
SO42- + Cl-
Na+ + K+
HCO3-
SO42-
Cl-
Mg2+
Ca2+
Meteoric Groundwater (Cold)
CF 15
CF 20
CF 45
CF 50
Acid-Sulphate
Pisciarelli
Pisciarelli 1
Pisciarelli 2
Pisciarelli A
Pisciarelli B
Alkali-Chloride-Sulphate
Hotel Tennis
Alkali-Bicarbonate
Terme di Agnano
Alkali-Chloride-Bicarbonate
Costagliola
Damiani
Esposito
Parco Cuma
Terme Puteolane
Alkali-Chloride
Lago d'Averno
Lago Lucrino
Stufe di Nerone
Stufe di Nerone (Spring)
Stufe di Nerone (Well)
Local Seawater
Geothermal Well - Mofete 1
Geothermal Well - Mofete 2

Hotel Tennis

Acid-Chloride-Sulphate Water

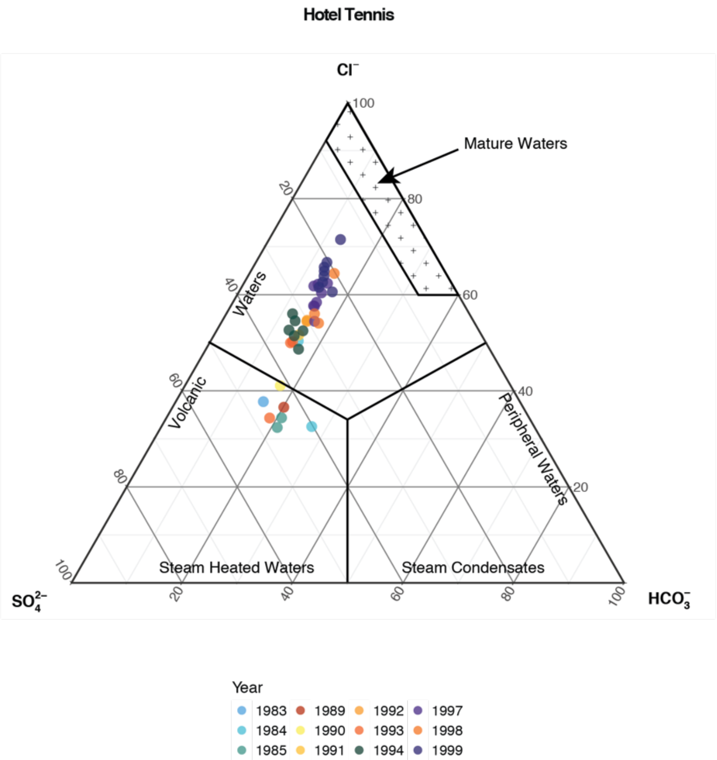


Figure 9.3: Ternary diagram after Giggenbach, (1980) showing the relative concentrations of major anions over time at Hotel Tennis.

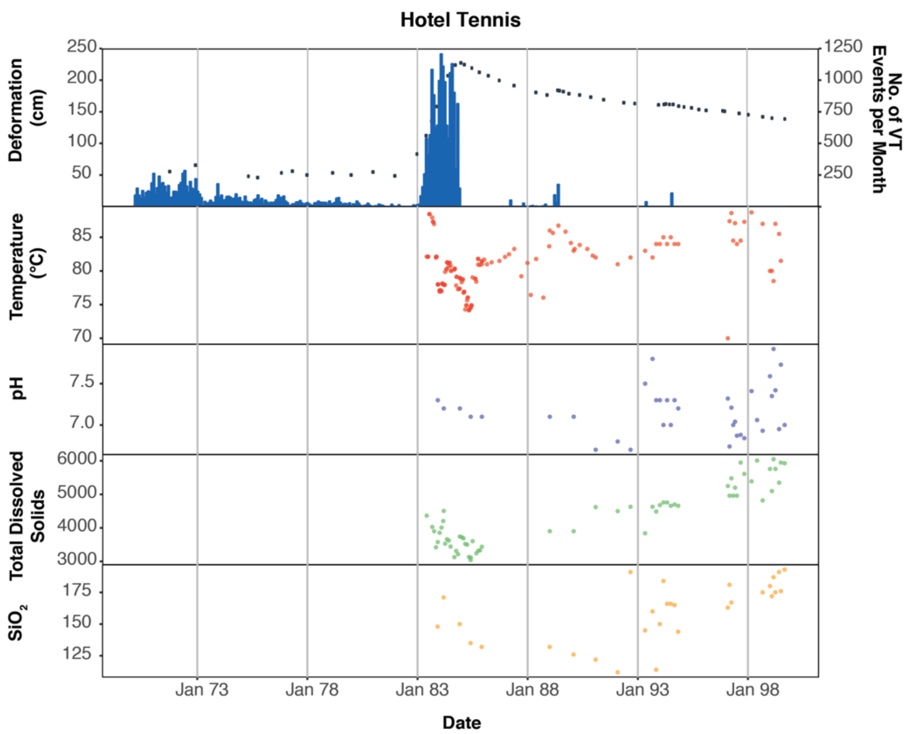


Figure 9.4: Temporal trends in geophysical parameters and water characteristics at Hotel Tennis. Top panel: deformation (dotted line) against rates of Volcano-Tectonic (VT) seismicity (blue columns).

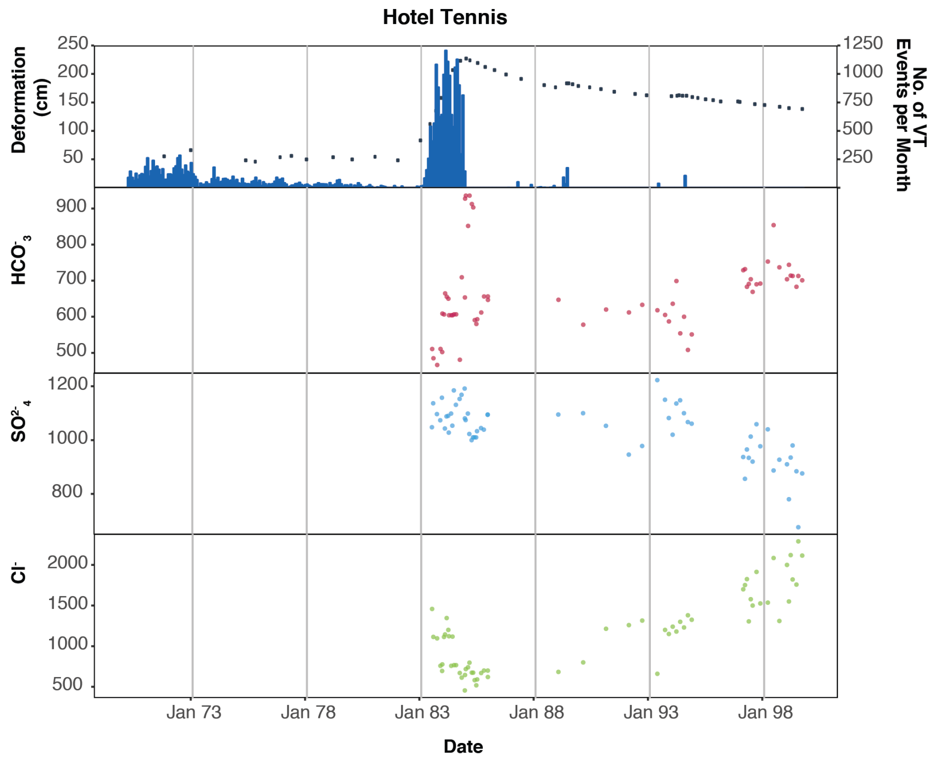


Figure 9.5: Temporal trends in geophysical parameters and major anion concentrations at Hotel Tennis. Top panel: deformation (dotted line) against rates of Volcano-Tectonic (VT) seismicity (blue columns).

Stufe di Nerone (Spring)

Alkali-Chloride Water

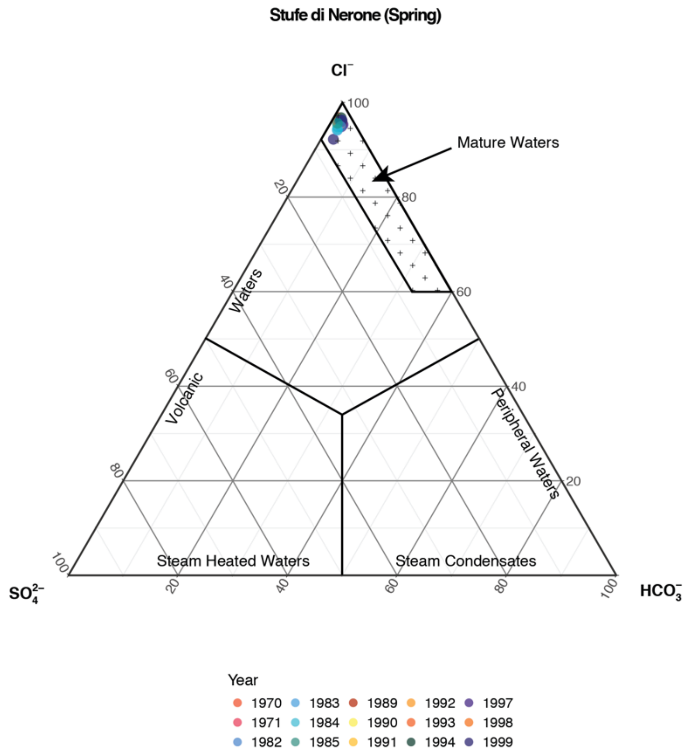


Figure 9.6: Temporal trends in geophysical parameters and water characteristics at Stufe di Nerone (Spring). Top panel: deformation (dotted line) against rates of Volcano-Tectonic (VT) seismicity (blue columns).

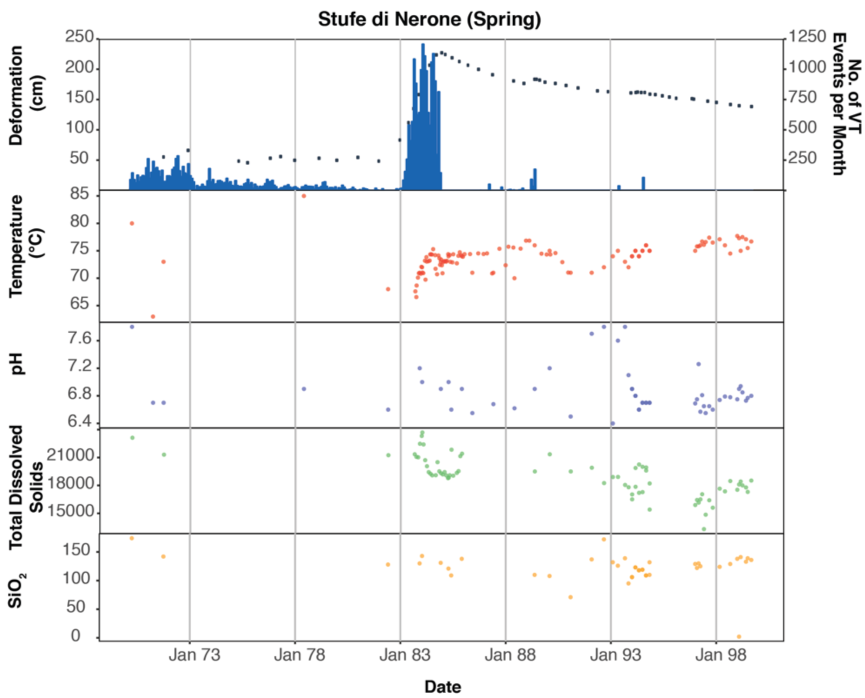


Figure 9.7: Temporal trends in geophysical parameters and water characteristics at Stufe di Nerone (Spring). Top panel: deformation (dotted line) against rates of Volcano-Tectonic (VT) seismicity (blue columns).

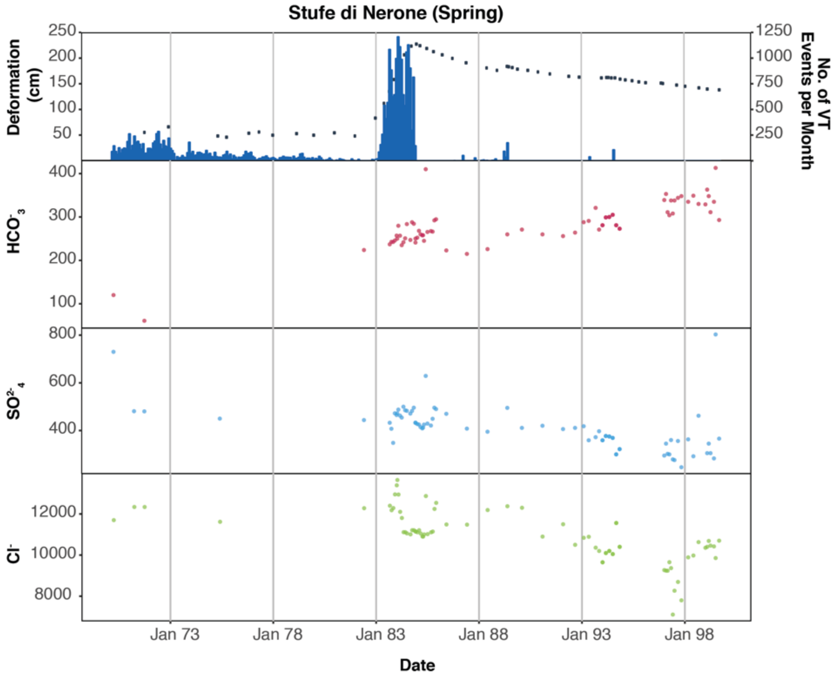


Figure 9.8: Temporal trends in geophysical parameters and major anion concentrations at Stufe di Nerone (Spring). Top panel: deformation (dotted line) against rates of Volcano-Tectonic (VT) seismicity (blue columns).

Terme Puteolane

Alkali-Chloride-Bicarbonate Water

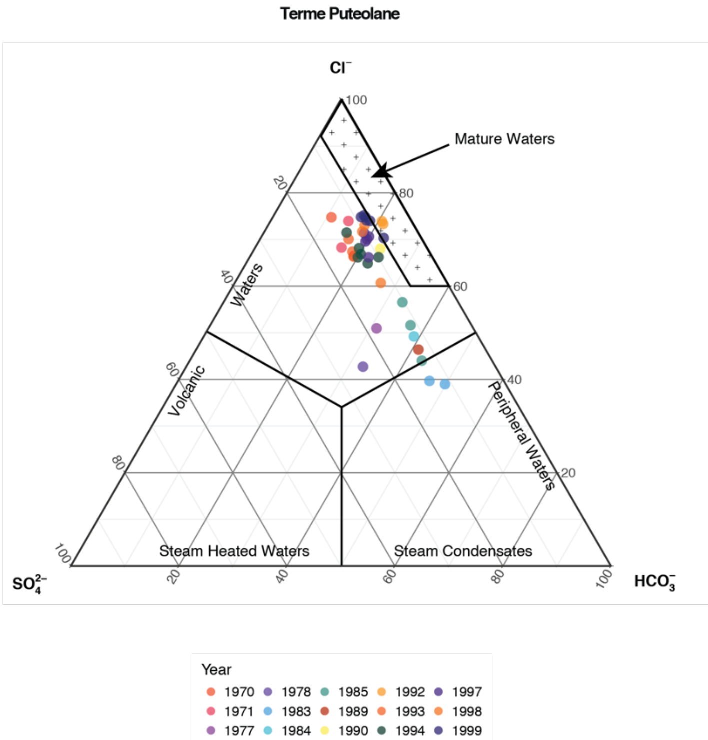


Figure 9.9: Ternary diagram after Giggenbach, (1980) showing the relative concentrations of major anions over time at Terme Puteolane.

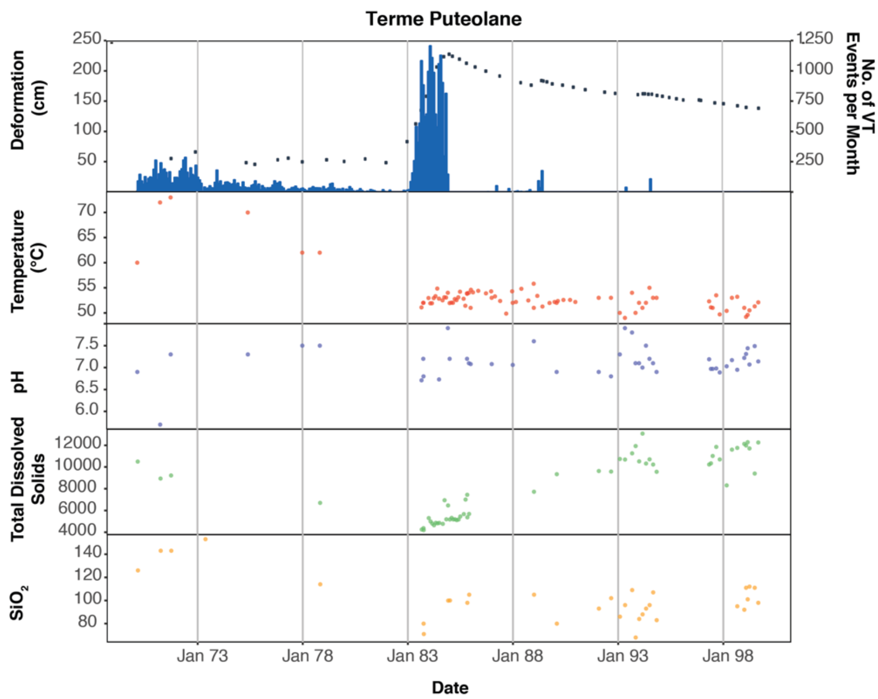


Figure 9.10: Temporal trends in geophysical parameters and water characteristics at Terme Puteolane. Top panel: deformation (dotted line) against rates of Volcano-Tectonic (VT) seismicity (blue columns).

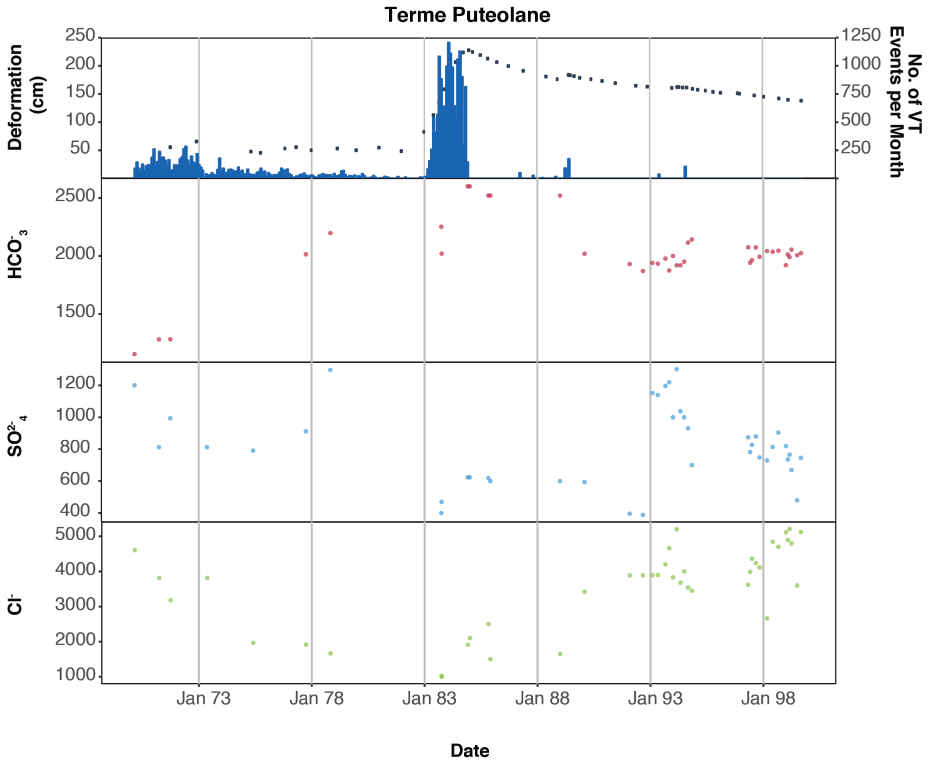


Figure 9.11: Temporal trends in geophysical parameters and major anion concentrations at Terme Puteolane. Top panel: deformation (dotted line) against rates of Volcano-Tectonic (VT) seismicity (blue columns).

(B) Distribution of Seismicity at Campi Flegrei

The following figures were constructed using ArcScene® by Esri. They show the spatial distribution of Volcano-Tectonic (VT) seismicity during uplift between 1983-1984 and 2005-2017, in relation to the seismic anomaly below Pozzuoli that was inferred by Aster and Meyer (1988) to indicate the location of hydrothermal reservoirs below Pozzuoli. Figures for seismicity in 1983-1984 were created using 2791 located events (Figs. 9.12 and 9.13). Those for seismicity in 2005-2017 used a catalogue of 492 located earthquakes (Figs. 9.14 and 9.15). All earthquake data are from the Vesuvius Observatory. The seismic anomaly was digitised from Aster and Meyer (1988).

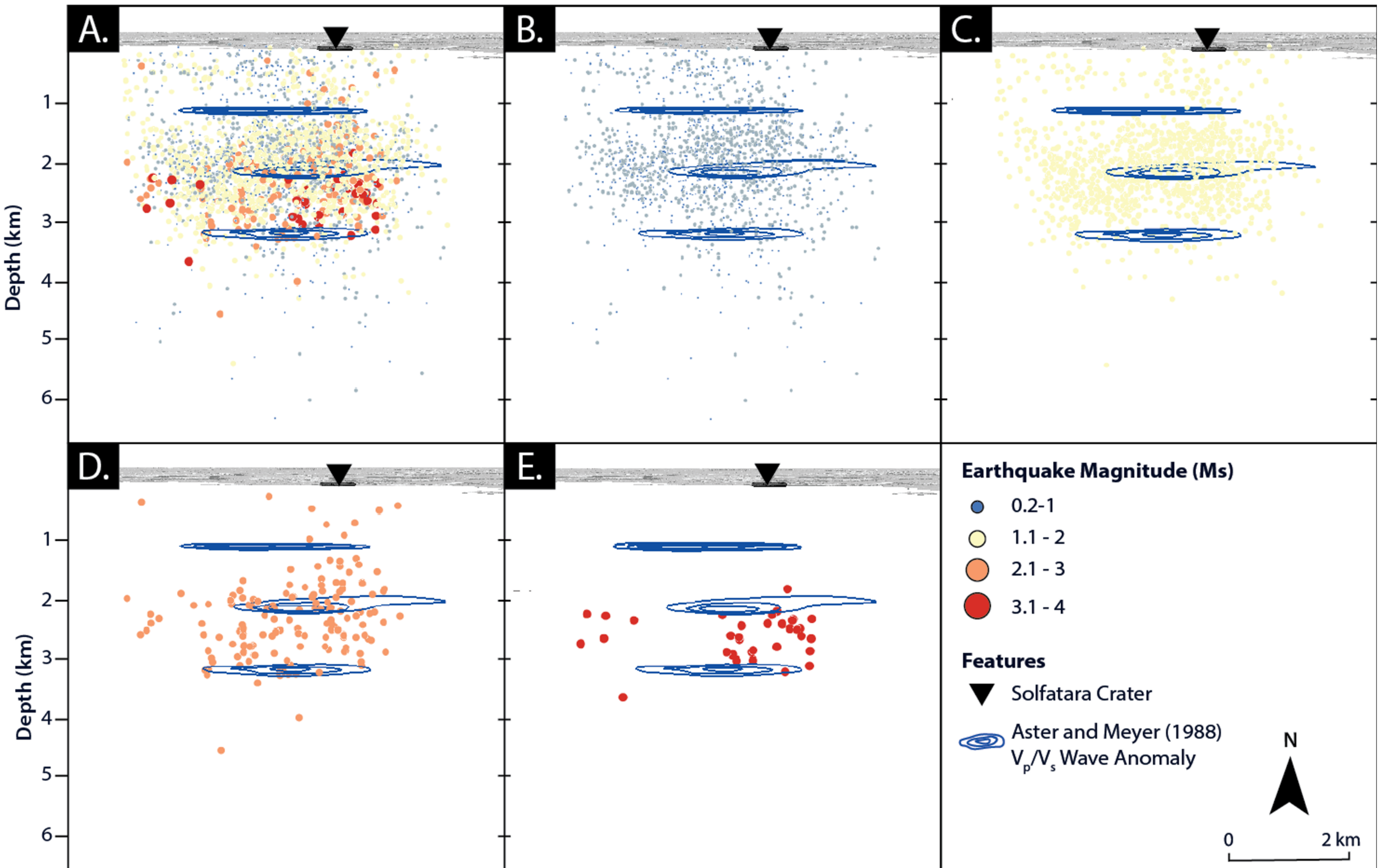


Figure 9.12: Distribution of hypocentres between 1983-1984 looking N from the Gulf of Pozzuoli. The seismogenic volume broadly corresponds with the seismic anomaly identified by Aster and Meyer (1988) as a fractured volume saturated in fluids. The largest events are concentrated below the caprock (c. 1.5 km) within the region expected to contain the hydrothermal reservoirs. Events are coloured by magnitude.

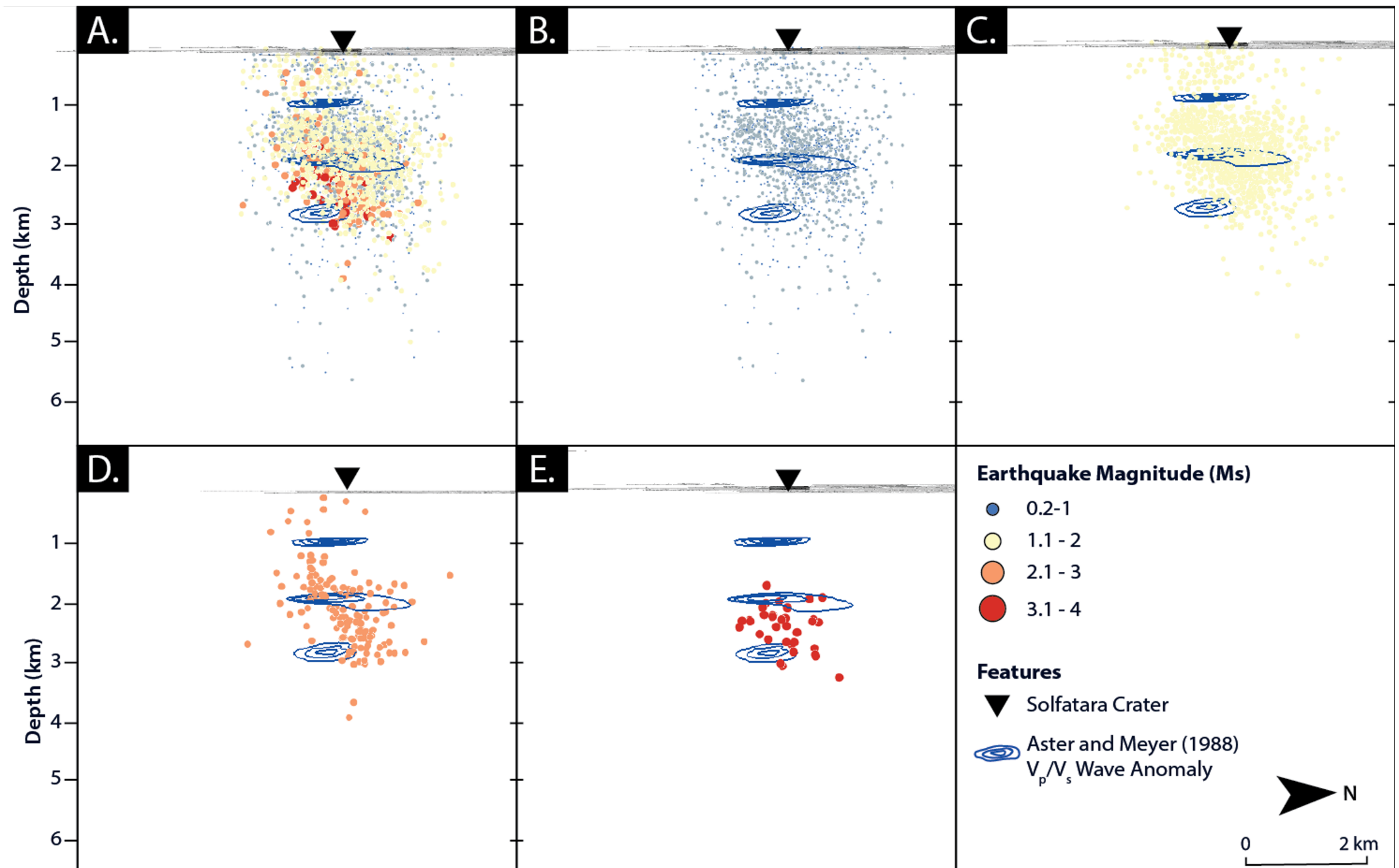


Figure 9.13: Distribution of hypocentres between 1983-1984 looking from the E to the W from the Gulf of Pozzuoli. The seismogenic volume broadly corresponds with the seismic anomaly identified by Aster and Meyer (1988) as a fractured volume saturated in fluids. The largest events are concentrated below the caprock (c. 1.5 km) within the region expected to contain the hydrothermal reservoirs. Events are coloured by magnitude.

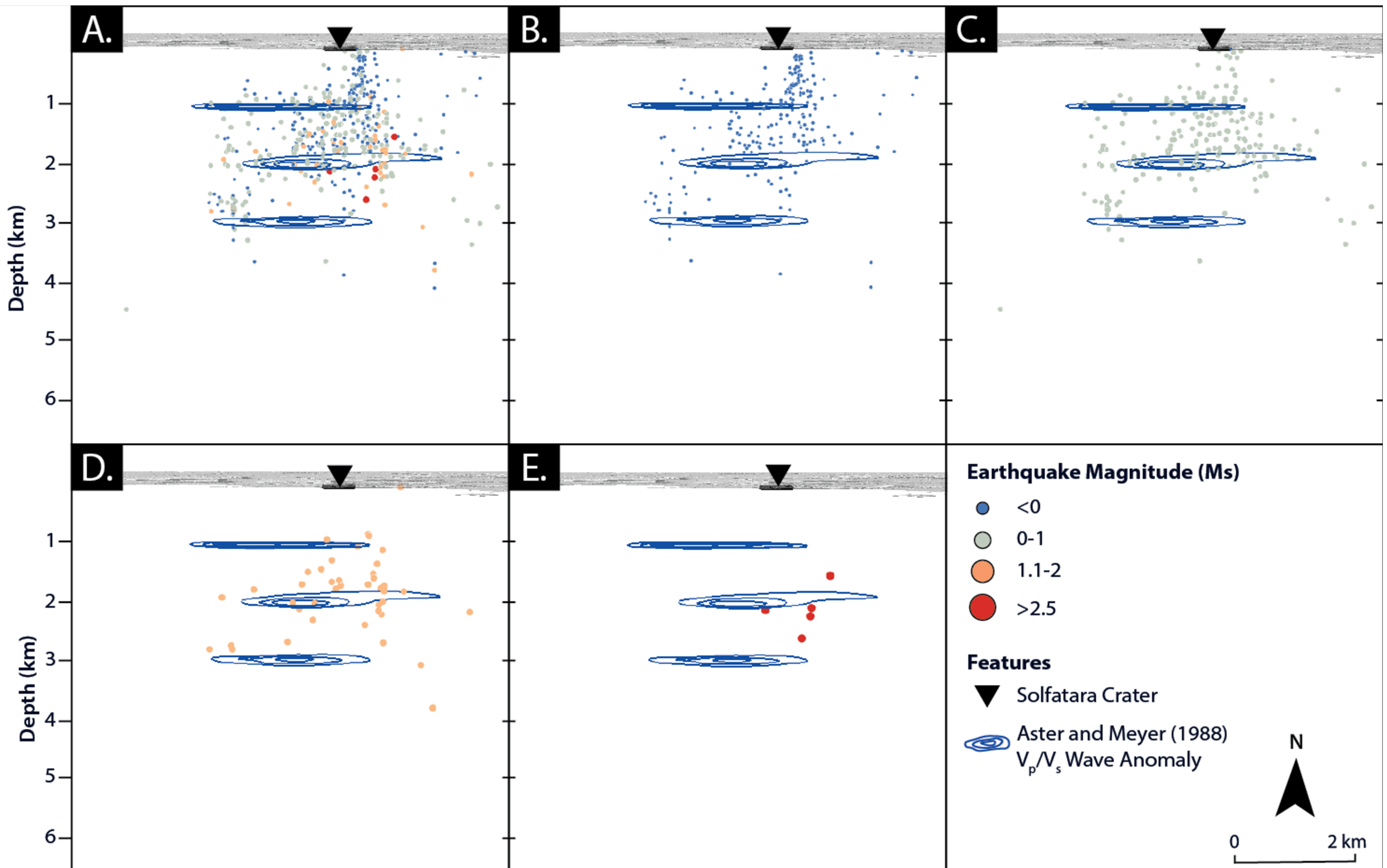


Figure 9.14: Distribution of hypocentres between 2005-2017 looking N from the Gulf of Pozzuoli. The seismogenic volume broadly corresponds with the seismic anomaly identified by Aster and Meyer (1988) as a fractured volume saturated in fluids. Events are coloured by magnitude,

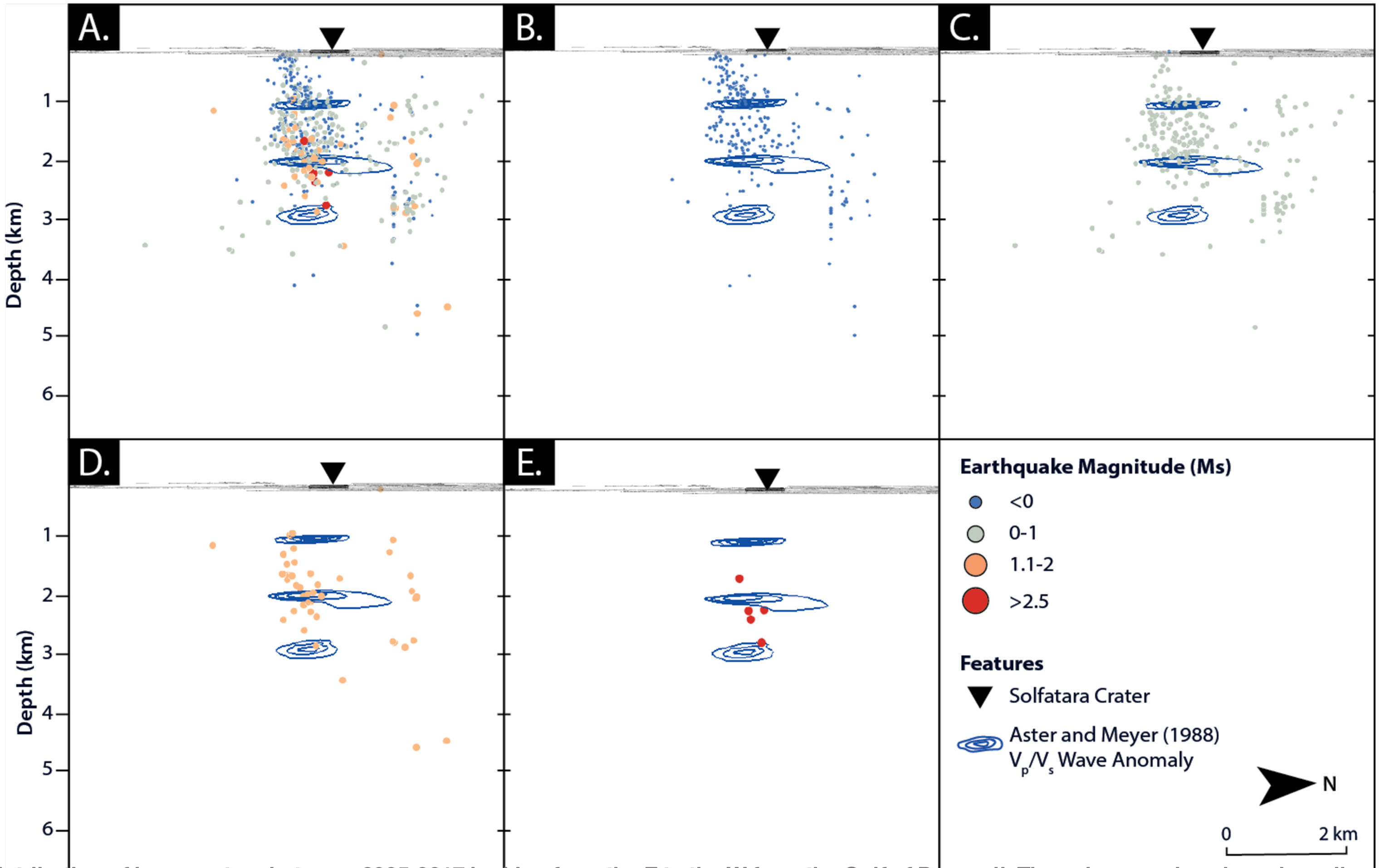


Figure 9.15: Distribution of hypocentres between 2005-2017 looking from the E to the W from the Gulf of Pozzuoli. The seismogenic volume broadly corresponds with the seismic anomaly identified by Aster and Meyer (1988) as a fractured volume saturated in fluids. Events are coloured by magnitude.

(C) University College London (UCL) Research Ethics Committee application and approval to conduct low risk research.

This Appendix contains the application to conduct Low-Risk research for Chapter 6 of the thesis submitted to the University College London (UCL) Research Committee and approval letter. An application to amend the original research plan and an email approving the amendment is also included.

UCL Research Ethics Committee Low-Risk Research Application Form

LONDON'S GLOBAL UNIVERSITY

UCL

UCL Research Ethics Committee

Note to Applicants: It is important for you to include all relevant information about your research in this application form as your ethical approval will be based on this form. Therefore anything not included will not be part of any ethical approval.

You are advised to read the Guidance for Applicants when completing this form.

Application For Ethical Review: Low Risk

Are you applying for an urgent accelerated review? Yes ☐ No ☒

If yes, please state your reasons below. Note: Accelerated reviews are for exceptional circumstances only and need to be justified in detail.

Is this application for a continuation of a research project that already has ethical approval? *For example, a preliminary/pilot study has been completed and is this an application for a follow-up project?*	Yes ☐ No ☒

If yes, provide brief details (see guidelines) including the title and ethics id number for the previous study:

Section A: Application details

1	**Title of Project**	Communication of volcano status at Campi Flegrei caldera, Italy
2	**Proposed data collection start date**	01/03/2019
3	**Proposed data collection end date**	24/09/2019
4	**Project Ethics Identification Number**	8601/001
5	**Principal Investigator**	Lara Smale
6	**Position held** (Staff/Student)	Student
7	**Faculty/Department**	Earth Sciences
8	**Course Title** *(if student)*	PhD
9	**Contact Details** Email: Telephone:	 lara.smale.13@ucl.ac.uk [redacted]
10	**Provide details of other Co-Investigators/Partners/Collaborators who will work on the project.** ***Note:*** *This includes those with access to the data such as transcribers.*	

Name: Dr. Christopher Kilburn Position held: Staff	Name: Dr. Stephen Edwards Position held: Staff

Faculty/Department: Earth Sciences Location (UCL/overseas/other UK institution): UCL Email: c.kilburn@ucl.ac.uk	Faculty/Department: Earth Sciences Location (UCL/overseas/other UK institution): UCL Email: s.edwards@ucl.ac.uk
If you **do not know** the names of all collaborators, please write their roles in the research. The Civil Protection Liaison Officer at the Vesuvius Observatory will help to distribute questionnaires to scientists' offices at the Observatory and with translation of questionnaire questions, information sheets and consent forms from English into Italian.	

11	**If the project is funded *(this includes non-monetary awards such as laboratory facilities)***
Name of Funder	Natural Environment Research Council (NERC)
Is the funding confirmed?	Yes. Funds are from an existing NERC DTP studentship

12	**Name of Sponsor**
The Sponsor is the organisation taking responsibility for the project, which will usually be UCL. If the Sponsor is not UCL, please state the name of the sponsor.	

13	**If this is a student project**
Supervisor Name	Dr. Christopher Kilburn
Position held	Director, UCL Hazard Centre
Faculty/Department	Earth Sciences
Contact details	c.kilburn@ucl.ac.uk

Section B: Project details

The following questions relate to the objectives, methods, methodology and location of the study. Please ensure that you answer each question in lay language.

14	**Provide a *brief* (300 words max) background to the project, including its intended aims.**
	Campi Flegrei, a caldera-type volcano in Italy, has been in a state of unrest since 1950 that is similar to behaviour before it last erupted in 1538. Three episodes of rapid ground uplift (c.1 m yr^{-1}) occurred between 1950-1984 that elevated the caldera by c. 4 m. Since 2005 uplift has resumed but at a rate that is 17 times slower. No eruption has occurred yet, however evacuations of up to 40000 people during previous uplifts have had significant long-term impacts. Essential to effective caldera unrest management is clear, consistent and reliable information transfer between Scientists, Decision Makers (those responsible for emergency management), the Media and the Public. Inadequate communication of the volcano status and associated uncertainty can lead to false alarms, breakdown of trust between stakeholders and, in the event of an eruption, loss of life. Campi Flegrei poses a particular challenge due to competing interpretations of the source of the unrest (e.g. magmatic vs. non-magmatic) and thus the hazard. Furthermore, accounts of past evacuations suggest there has been a high degree of mistrust between stakeholders, whilst the longevity of uplift episodes has the potential to promote the proliferation of mis-

	information, especially given the ease with which information can be transformed on social media. The principal aims of this study are i) to establish the extent to which differing conceptual models of the volcano's behaviour exist, ii) to assess the extent to which existing communications protocols meet end-user requirements, and iii) to assess the perceptions and levels of trust between different stakeholders. This novel project will collect data with the view of informing recommendations for improving existing warning plans in case of a future volcano emergency at Campi Flegrei. Target participants include volcano scientists, Decision Markers (i.e. Civil Protection and local authorities) and members of the public.

15 Methodology & Methods (tick all that apply)

☒ Interviews*

☐ Focus groups*

☒ Questionnaires (including oral questions)*

☐ Action Research

☐ Observation

Participant Observation

☐ Documentary analysis (including use of personal records)

☒ Audio/visual recordings (including photographs)

**Attach copies to application (see below).*

☐ Collection/use of sensor or locational data

☐ Controlled Trial

☐ Intervention study (including changing environments)

☐ Systematic review

☐ Secondary data analysis – ***(See Section D)***

☐ Advisory/consultation groups

☐ Other, give details:

16a Provide – <u>in lay person's language</u> - an overview of the project; focusing on your methodology and including information on what data/samples will be taken (including a description of the topics/questions to be asked), how data collection will occur and what (if relevant) participants will be asked to do. This should include a justification for the methods chosen. **(500 words max)**

Please **do not** attach or copy and paste a research proposal or case for support.

Data will be collected from scientists, Decision Makers and members of the public likely to be involved in a future volcano emergency at Campi Flegrei. Four versions of a questionnaire have been created for different phases of data collection. Each is constructed of Campi Flegrei centric questions that will collect qualitative and quantitative data. Questions are organised into seven themes as follows; i) demography, ii) past experience of volcano emergencies, iii to iv) beliefs regarding causes of past and current unrest respectively, v) beliefs about future activity, vi) communication of volcano status, and vii) perceptions of other stakeholders. The questionnaires will be self-administered and answered on a voluntary basis in the participants own time. Semi-structured interviews will be conducted with volunteers from the Vesuvius Observatory only. The interview schedule follows the same question themes.

Data collection will be carried out in four phases. In phase 1 an online survey created using Opinio will be shared with scientists and Decision Makers globally using the Volcano Listserv and Twitter. This survey will be used in lieu of a pilot study due to funding and time restrictions to check the formatting of questions. Phase 2 will consist of data collection via hard-copy questionnaires delivered to scientists' offices at the Vesuvius Observatory, the institution responsible for monitoring Campi Flegrei. Interviews will be carried out with volunteers from this group, at the Observatory or via Skype depending on the interviewee's preference. The purpose of the interviews is to

	elucidate information from the questionnaires. Written and informed consent will be obtained prior to interview as per the ESRC Framework for Research Ethics. Phase 3 will consist of an anonymous online survey shared by a link sent to scientists external to the Observatory who have previously published on the volcano and/or contributed to emergency planning. A link will also be sent to relevant responsible individuals at each section of the Istituto Nazionale di Geofisica e Vulcanologia (INGV) to be shared internally. A link to the questionnaire for Decision Makers will be sent to an email distribution list and to relevant responsible individuals at target organisations (e.g. Civil Protection, voluntary groups, local authorities) to be shared by internally. Phase 4 will consist of an online questionnaire for members of the public living in the Campi Flegrei region shared via social media. Data collection will be in Italian, except during the interviews, which will be in English due to language limitations. Results from questionnaires will be coded according to a numerical coding frame for analysis using SPSS, whilst interviews will be transcribed by the data collector then uploaded into the QSR NVivo® software for thematic analysis. The study has been designed to maximise the response rate amongst monitoring scientists, who are the primary study target due to the critical nature of communication between scientists and Decision Makers in volcanic emergencies. An online survey has been adopted for the other stakeholders to access groups not based locally to Campi Flegrei and to keep within time and financial constraints.
16b	**Attachments** If applicable, please attach a copy of any interview questions/workshop topic guides/questionnaires/test (such as psychometric), etc and state whether they are in final or draft form. Drafts of the four questionnaires and the interview schedule are attached. Once finalised they will be translated by a native speaker from the Vesuvius Observatory into Italian.

17	**Please state which code of ethics (see Guidelines) will be adhered to for this research (for example, BERA, BPS, etc).**
ESRC	

Location of Research	
18	**Please indicate where this research is taking place.** ☐ UK only (Skip to 'location of fieldwork') ☐ Overseas only ☒ UK & overseas
19	**If the research includes work outside the UK, is ethical approval in the host country (local ethical approval) required?** ***(See Guidelines.)*** Yes ☐ No ☒ **If no,** please explain why local ethical approval is not necessary. **If yes,** provide details below including whether the ethical approval has been received. **Note:** Full UCL ethical approval will not be granted until local ethical approval (if required) has been evidenced. No local ethical approval is required as the research is of minimal risk and there is no relevant local ethics approval process. This has been queried and confirmed by knowledgeable persons at both the Vesuvius Observatory and the Department of Social Sciences at the Università degli Studi di Napoli Federico II.

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20	**If you (or any members of your research team) are travelling overseas in person are there any concerns based on governmental travel advice (*www.fco.gov.uk*) for the region of travel?** Yes ☐ No ☒ **Note:** C*heck www.fco.gov.uk and submit a travel insurance form to UCL Finance (see application guidelines for more details). This can be accessed here:* https://www.ucl.ac.uk/finance/secure/fin_acc/insurance.htm (You will need your UCL login details.)

21	**State the location(s) where the research will be conducted and data collected. For example public spaces, schools, private company, using online methods, postal mail or telephone communications.** Data collection will take place using questionnaires and interviews at the Vesuvius Observatory, which is a section of the Istituto Nazionale di Geofisica e Vulcanologia (INGV), a research institute for geophysics and volcanology in Italy that is funded by the Italian Ministry of Education, Universities and Research. Data will also be collected using online methods by sharing of a link to an online survey shared via email and social media.
22	**Does the research location require any additional permissions (e.g. obtaining access to schools, hospitals, private property, non-disclosure agreements, access to biodiversity permits (CBD), etc.)?** Yes ☒ No ☐ **If yes,** please state the permissions required. Permission is required and has been obtained to access the Vesuvius Observatory, as well as to carry out data collection from Dott.ssa Francesca Bianco, the Observatory Director. Approach letters will be sent to the appropriate responsible individuals at selected relevant scientific/decision maker organisations asking them for permission to collect data via an online questionnaire. On receipt of permission an email containing a link to the information sheet and questionnaire will be sent to the contact at the organisation for them to distribute internally via email.
23	**Have the above approvals been obtained?** Yes ☐ No ☒ **If yes,** please attach a copy of the approval correspondence. Only approval for access and data collection at the Vesuvius Observatory has been obtained thus far. Other approvals will be obtained after completion of the ethics approval process and finalisation of the online questionnaires. **If not,** confirm they will be obtained prior to data collection. Yes ☒ No ☐

Section C: Details of Participants

In this form 'participants' means human participants and their data (including sensor/locational data, observational notes/images, tissue and blood samples, as well as DNA).

24	**Does the project involve the recruitment of participants?**
Yes ☒	Complete all parts of this Section.
No ☐	Move to Section D.

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Participant Details	
25	Approximate maximum number of participants required: N/A Approximate upper age limit: N/A Lower age limit: 18 Justification for the age range and sample size: There is no maximum number of participants required. A reasonable estimate of the number of participants based at the Vesuvius Observatory is between 20-30. This is based on the response rates of other studies using similar methods elsewhere. This dataset will form the core of the study. If the sample size is too small for quantitative analysis, then the data can be used qualitatively. No maximum sample size is required for data collected using online methods as these results will be used for comparison with the Vesuvius Observatory results. If the response rate from these participants is insufficient for quantitative analysis, then the results will be used qualitatively in discussion with the core results. The study will only include adults over the age of 18 and there is no upper limit to the age. This is to be representative of the population most likely to have to make response decisions in a future emergency, either as an individual or as part of an institution.
Recruitment/Sampling	
26	Describe how potential participants will be recruited into the study. **Note:** This should include reference to how you will identify and approach participants. For example, will participants self-identify themselves by responding to an advert for the study or will you approach them directly (such as in person or via email)? All aspects of data collection are reliant upon voluntary participation in the research. Participants will choose to opt-in to the study. • Global survey targeting volcano scientists and decision makers. A link to an online survey accompanied by an information sheet will be shared using the volcano listserv and social media accounts belonging to the Principal Investigator and the UCL Hazard Centre. • Vesuvius Observatory Scientists. Hard copy questionnaires, together with information sheets will be delivered to offices in the Observatory with permission of the Observatory Director (Dott.ssa Francesca Bianco). The questionnaire will also have a link to an online version, should the scientists prefer to complete the questionnaire in an electronic format. The questionnaire will ask if scientists would like to take further part in the study by being interviewed and to contact the data collector using provided details if that is the case to arrange an interview. • Non-Observatory Scientists and Decision Makers. An email explaining the study, together with a link to the information sheet and online questionnaire will be sent to an email distribution list of individuals directly inviting them to participate in the study. Approach letters will also be sent to the appropriate responsible individuals at selected relevant scientific/decision maker organisations asking them for permission to collect data. On receipt of permission an email containing a link to the information sheet and questionnaire will be sent to the contact at the organisation for them to distribute internally via email. • Members of the Public. A link to the information sheet and questionnaire will be made available on Social Media (Twitter and Facebook) for individuals to discover.
Informed Consent	
27a	Describe the process you will use when seeking to obtain consent.

	Note: This should include reference to what participants are being asked to consent to, such as whether their contribution will be identifiable/anonymous, limits to confidentiality and whether their data can be withdrawn at a later date. *(An annotated template information sheet and consent form have been provided for your use.)* Information sheets will be provided with questionnaires, for both the hard-copy and online versions. The questionnaires will then be completed anonymously and voluntarily, in the respondent's own time. No consent forms will therefore be required as consent is implicit in the completion and submission of the questionnaire, as according to the UCL Research Ethics Committee guidelines. The information sheet will explain the purpose of the study and that by completing the questionnaire the participant is consenting to their responses being used in a chapter of a PhD thesis, which may be published in a peer reviewed journal in the future. It also explains that their responses will be stored on a password protected device until completion of the PhD. It explains that questions may be skipped, that they may terminate the questionnaire at any time, without giving an explanation and that once submitted their answers may not be withdrawn. The information sheet makes it clear that the researchers may be contacted for more information and that a copy of the findings may be requested. Written and informed consent will be collected from scientists at the Vesuvius Observatory who volunteer to take part in individual semi-structured interviews. Interviewees will be informed via the information sheet that the interview will be in English, that they can terminate the interview at any time, that any data they have provided to that point will be destroyed and that they may withdraw their answers from the study at any time in the four-week period following the interview. They will also be given an opportunity to discuss the project and the information sheet before being given the consent form. The consent form will ask them to confirm the following: • That they have read and understood the Information Sheet, that they have had time to consider the information and has an opportunity to ask questions • That they understand that their participation is voluntary, that they are free to skip any question without giving a reason and to withdraw from the study at any time. Also, that they are able to withdraw their data up to 4 weeks after interview. • That they consent to the interview being audio recorded and understand that the recordings will be deleted following transcription. • That they consent to hand written notes being made that will be destroyed after being digitised. • That they understand that the transcripts and/or notes will be fully anonymised and stored securely. I also understand that the signed consent form will be digitised, and that the researchers paper copy will be destroyed. • That they understand that their answers will be anonymous and used in a chapter of a PhD thesis, that the study may be published in a peer reviewed journal and that they will not be identifiable in the PhD thesis or in any publications. No interviews will proceed without obtaining the required consent.
27b	**Attachments** Please list them below: *Ensure that a copy of all recruitment documentation (recruitment emails/posters, information sheet/s, consent form/s) have been attached to the application.* The following items are attached. They have been provided in English but will be translated into Italian before use. • Information sheet for global online survey (for phase 1 of project) • Information sheet for questionnaire for scientists • Information sheet for questionnaire for Decision Makers • Information sheet for questionnaire for Members of the Public

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	• Information sheet for individual semi-structured interviews with Vesuvius Observatory scientists • Consent form for individual interviews with Vesuvius Observatory scientists • Draft invitations to participate in online questionnaire • Draft Approach Letter
27c	If you are ***not*** intending to seek consent from participants, clarify why below: Consent will not be sought from those completing the questionnaire as it is a self-completion survey that is completed on a voluntary basis. Consent is therefore implicit in its completion, as according to the UCL Research Ethics Committee guidelines.

28	How will the results be disseminated (including communication of results with participants)? The data is being collected for a chapter of a PhD thesis and with the view to publishing a paper in a peer reviewed journal. The information sheets provided to participants provide the researcher contact details and ask them to contact the researchers if they would like a copy of the PhD chapter on its completion, which is expected to be at the end of 2019. Either a digital or hard copy of the chapter and an accompanying executive summary will be sent depending on the recipient's preference. Copies of the PhD chapter and an accompanying executive summary will be sent to individuals who provided access for data collection at an organization.

Section D: Accessing/Using Pre-collected Data

Access to data

29	If you are using data or information held by third party, please explain how you will obtain this. You should confirm that the information has been obtained in accordance with the General Data Protection Regulation 2018.

Accessing pre-collected data

30	**Does your study involve the use of previously collected data?** **No** ☒ Move to Section E. **Yes** ☐ Complete all parts of this Section. **Note:** If you ticked any boxes with an asterisk (*),ensure further details are provided in Section E: Ethical Issues.

31	**Name of dataset/s:**		
32	**Owner of dataset/s (if applicable):**		
33	**Is the data in the public domain?**	Yes ☐	No ☐
	If not, do you have the owner's permission/license?	Yes ☐	No* ☐
33	**Is the data anonymised?**	Yes ☐	No ☐
	If not:		
	i. Do you plan to anonymise the data?	Yes ☐	No* ☐
	ii. Do you plan to use individual level data?	Yes* ☐	No ☐

	iii. Will you be linking data to individuals?	Yes* ☐ No ☐
34	Is the data sensitive?	Yes* ☐ No ☐
35	Will you be conducting analysis within the remit it was originally collected for?	Yes ☐ No* ☐
36	If not, was consent gained from participants for subsequent/future analysis?	Yes ☐ No* ☐

Section E: Ethical Issues

Ethical Issues

37 Please address clearly any ethical issues that may arise in the course of this research and how they will be addressed. Further information and advice can be found in the guidelines.

Note: All ethical issues should be addressed - **do not leave this section blank**. All projects give rise to ethical issues. If you think there are no ethical issues, you need to provide an explanation as to why.

Identified ethical issues and how they will be addressed are listed below.

- Ensuring information sheets and consent forms are understood. All information sheets and consent forms will be provided in the native language of the study participants (Italian). In the case of the global survey in phase one of the data collection they will also be provided in English. Italian translations of the original English versions will be carried out by a native speaker. They will then be reverse translated into English prior to distribution to ensure that the original meaning of the text has been maintained and any corrections will be made accordingly. The researchers contact details are also clearly provided on the information sheets, so that they may be contacted to answer any questions and to provide further information about the study. Scientists who volunteer to be interviewed will be given information sheets and consent forms in advance of the interview, so that they have the opportunity to ask questions about the study or to request further information.
- Ensuring participants understand the meaning of questions in questionnaires. Caldera unrest is unlikely to be a familiar topic to some of the intended study participants and those not in the scientist group are likely to be non-experts. There is also a possibility that non-experts may develop an inappropriately heightened concern for future activity at Campi Flegrei. To mitigate this the translation of the questionnaire materials will be undertaken by a native speaker from the Vesuvius Observatory, who is also involved in outreach activities and is therefore familiar with communicating to different audiences. As for the information sheets and consent forms they will also be reverse translated as a check before data collection. Different versions of the questionnaire have been produced for the different participant groups to maximise the likelihood of understanding by ensuring appropriate use of language. To ensure that all participants have the same base level information regarding the volcano, descriptions of past activity, maps, graphs with accompanying explanations and

likelihood translation tables have been provided. The language used and information regarding the activity of the volcano is consistent with that used in publicly available materials published by the Vesuvius Observatory in order to prevent false perceptions or unnecessary concerns regarding the volcano developing. No jargon has been included and potentially ambiguous terms have been defined. Information directing participants to official sources of volcano information (the Vesuvius Observatory and Civil Protection) will also be included on the final questionnaires for reference. The information sheets make it clear that questions may be skipped.

- Ensuring that participants feel free to answer questionnaires honestly, even if they think their answers differs from others in the organisation. All questionnaires will be self-administered, completed in the participants own time and in a location of their choosing. Any question may be skipped if the participant would prefer not to answer, without giving a reason. The questionnaires do not ask for any directly identifiable information and the questions have been structured to minimise the risk of indirect identification of individuals. No individuals will be identifiable in the results. This will be made clear in the information sheets.
- Ensuring the meaning of questionnaire responses are maintained when they are translated into English for data analysis. The questionnaires largely consist of closed questions where there is no possibility for ambiguity. Answers to open questions will be translated by the Principal investigator and checked with the PhD project supervisor (a fluent Italian speaker). Reverse translation checks will be carried out to ensure the original meanings are maintained.
- Ensuring interview participants understand the meaning of questions during interviews. The interviews with Vesuvius Observatory scientists will be conducted in English out of necessity as the Principal researcher does not speak Italian and funds do not currently allow for an interpreter. Participation in the interviews is voluntary and interviewees will be made fully aware that it will be conducted in English before volunteering to participate. They will therefore have self-identified a certain confidence and proficiency in English. All information sheets and consent forms will be in Italian. In the event it becomes clear that a question has not been understood, the answer will not be included in the results. Participants will also be made aware that they can skip any question or terminate the interview at any time without giving a reason why on the information sheets.
- Ensuring that participants feel free to answer interview questions honestly, even if they think their answers differs from others in the organisation and maintaining confidentiality during interviews. Only the researchers and the participant will be present, to avoid any perceived pressure from other members of the Observatory, it will also take place at a location of the participants choosing within the Observatory, or over Skype to ensure they are comfortable with the setting. Only the principal investigators will know the identities of the interviewees. Transcripts will be anonymised, and audio recordings deleted after transcription. No individuals will be identifiable in the results. This will be made clear in the information sheets. There is no previously established relationship between the principal investigator, who will be conducting the interviews, and potential participants.
- Online surveys. It is important that participants are aware of the identities of the researchers, the purpose of the study and its legitimacy. The researcher contact details are provided on the information sheets and at the end of the questionnaire. The researchers, research group and department are also fully discoverable online, should participants wish to establish their identity. Hyperlinks to researcher profiles on the UCL Department of Earth Sciences website will be added to the information sheet.

	• Ensuring confidentiality. The primary ethical issue of the study is ensuring that participants cannot be identified. No directly identifiable information will be solicited and where free text answers to questionnaire items are given, the results will be generalised into specific themes or assigned to an "other" category. This is to aid quantitative analysis and to avoid identification of individuals who may hold a unique opinion. The results of the interviews will be used only to elucidate answers in the questionnaires completed by Vesuvius Observatory scientists. No views that are distinct from those collected in the questionnaires will be included in the results to exclude potentially identifiable opinions. Hard copy materials (e.g. questionnaires, interview consent forms) and audio files will be destroyed as soon as practicable after digitisation or transcription respectively. • Permission to collect data. Data collection at the Vesuvius Observatory, at other sections of the INGV and Decision Maker organisations presents a potential ethical issue if permission from responsible individuals at those organisations is not given. Permission has been sought and obtained from the director of the Vesuvius Observatory (Dott.ssa Francesca Bianco). Permission will be sought from relevant responsible individuals at other organisations via email. Only once permission has been obtained will a link to an online questionnaire be sent. • Sampling strategy. An opt-in approach has been adopted, whereby the participants must take an active step to take part in the study by completing the questionnaire or completing a consent form in the case of the interviews at the Vesuvius Observatory. Questionnaires will be made freely available so that participants may choose to take part independently. They will be self-administered and completed in the participants own time. Interviews will take place with volunteers at a time and location of their choosing. • Ensuring informed consent. All information sheets will be provided in Italian, explain the purpose of the study and how the researchers may be contacted for more information. The questionnaire information sheets also make it clear that participation is voluntary and what it is participants are consenting to by completing a questionnaire. It is clearly stated what the results will be used for and how participants may get a copy of the results should they wish. In the case of the interviews, those that volunteer to take part in an individual interview will be provided with an information sheet and will be required to sign a consent form should they wish to proceed. The consent forms clearly state what the participant is consenting to by taking part. Those that volunteer to be interviewed will be given information sheets and consent forms in advance to ensure they fully understand what they are consenting to and to give them opportunity to contact the researchers for more information if necessary. Both information sheets and the interview informed consent form have been designed according to UCL Research Ethics Committee guidelines and ESRC Ethics Principles. • Unintended association with the Istituto Nazionale di Geofisica e Vulcanologia (INGV) activities. The Istituto Nazionale di Geofisica e Vulcanologia (INGV) are responsible for monitoring volcanoes in Italy and the Vesuvius Observatory is a section of this organisation. To avoid unintended association a clear statement expressing that this study is not connected to the INGV in anyway is included in all the information sheets. In the event of unintentional or unforeseen consequences resulting from the data collection, the UCL Research Ethics Committee will be contacted for guidance on how to proceed.

Risks & Benefits	
38	Please state any *benefits* to participants in taking part in the study (this includes feedback, access to services or incentives),

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	There are no specific benefits to participants. They will be informed that they are contributing to current research and that they may request a copy of the results in the form of the completed PhD chapter.
39	Do you intend to offer incentives or compensation, including access to free services)? Yes ☐ No ☒ **If yes,** specify the amount to be paid and/or service to be offered **as well as a justification for this**.
40	Please state any *risks* to participants and how these risks will be managed. We believe that there are no foreseeable physical, psychological, social, economic or legal risks to the participants involved in the study. Risks to individual privacy are minimal. In the case of the paper questionnaires distributed at the Vesuvius Observatory there is a minimal risk that a familiar person could recognise the handwriting on a completed questionnaire or identify a person through a combination of a series of variables. It is unlikely that the content of the questionnaire could be detrimental to the respondent, however to minimise the risk of a loss of confidentiality participants will be asked to return questionnaires in a sealed envelope that will be provided. The researchers will not know the identity of anyone that completes a questionnaire. The questionnaires will remain sealed until they can be digitised by the Principal Investigator, which will happen as soon as practicable. The paper versions will be destroyed, and the electronic copies will be stored on an encrypted MacBook, which will be backed up on an encrypted USB. On return to the UK the files will be transferred onto a password protected computer in the UCL Department of Earth Sciences on return to the UK and removed from the laptop. A backup will be stored on the encrypted USB. A minimal loss of confidentiality risk has also been identified regarding the semi-structured interviews that will take place with Vesuvius Observatory scientists. To mitigate this risk the interviews will take place in a private location selected by the interviewee within the Observatory or over Skype and the identity of the interviewees will be known only by the Principal investigator. Interview consent forms will be digitised, as will any hand-written notes. No personal data other than the interviewee name and signature on the consent form will be collected. Hard-copies will be destroyed after digitisation. Audio files will be transcribed as soon as practicable and the transcripts will be fully anonymised. Audio files will be removed from the recording device and transferred onto an encrypted laptop on interview completion. All electronic files from the interviews will be stored on the laptop with the questionnaire data. The laptop will be backed up onto an encrypted USB. As for the questionnaires, all data will be transferred onto a password protected computer in the UCL Department of Earth Sciences on return to the UK and removed from the laptop. A backup will be stored on the encrypted USB. We believe there are no foreseeable risks to online survey participants. No names, addresses, post codes, email addresses, IP addresses or other information from which a participant could be directly identified will be collected. Data will be coded as soon as practicable and stored securely.
41	Please state any *risks* to you or your research team and how these risks will be managed. The risks to the research team whilst at the Vesuvius Observatory are minimal and are those associated with normal everyday activity in a workplace. A risk assessment has been completed. To minimise risk to the researchers when conducting individual interviews, they will take place in the Vesuvius Observatory during normal working hours or via Skype.

Section F: Data Storage & Security

Please ensure that you answer each question and include all hard and electronic data.

42	**Will the research involve the collection and/or use of personal data?** Yes ☒ No ☐ ***Personal data*** *is data which relates to a living individual who can be identified from that data OR from the data and other information that is either currently held, or will be held by the data controller (the researcher).* *This includes:* – *any expression of opinion about the individual and any intentions of the data controller or any other person toward the individual.* – *sensor, location or visual data which may reveal information that enables the identification of a face, address, etc (some postcodes cover only one property).* – *combinations of data which may reveal identifiable data, such as names, email/postal addresses, date of birth, ethnicity, descriptions of health diagnosis or conditions, computer IP address (if relating to a device with a single user).* If you do not have a registration number from Legal Services, please clarify why not:
43	**Is the research collecting or using** – sensitive personal data as defined by the General Data Protection Regulation (racial or ethnic origin / political opinions / religious beliefs / trade union membership / physical or mental health / sexual life / commission of offences or alleged offences), and/or – data which might be considered sensitive in some countries, cultures or contexts. **If yes,** state whether explicit consent will be sought for its use and what data management measures are in place to adequate manage and protect the data. No

44	**All research projects using personal data must be registered with Legal Services before the data is collected, please provide the Data Protection Registration Number:** **Z6364106/2019/01/125** If you do not have a registration number from Legal Services, please clarify why not:

During the project *(including the write up and dissemination period)*	
45	**State what types of data will be generated from this project** (i.e. transcripts, videos, photos, audio tapes, field notes, etc). • Data (quantitative and qualitative) from completed hard-copy questionnaires • Data (quantitative and qualitative) from completed online questionnaires

	• Audio recording of interviews (to be deleted once transcribed) • Transcripts of interviews • Hand written notes during interviews (to be destroyed once digitised) **How will data be stored, including where and for how long?** This includes all hard copy and electronic data on laptops, share drives, usb/mobile devices. During data collection at the Vesuvius Observatory hard-copies of completed questionnaires will be stored in a locked drawer at the Observatory or in a locked laptop bag to which only the Principal Investigator will have access. The questionnaires will be digitised, and the hard copies destroyed as soon as possible. Interview consent forms will be digitised as soon as practicable, as will any hand-written notes at which point hard copies will be destroyed. Audio files will be removed from the recording device and transferred onto an encrypted laptop on interview completion. The interviews will then be fully anonymised during transcription, which will take place as soon as possible after the interview so that the audio flies may be deleted. All electronic files created during onsite data collection in Italy will be stored on an encrypted laptop backed up onto an encrypted USB. On return to the UK all electronic files will be transferred to a password protected computer in the Department of Earth Sciences at UCL and removed from the laptop. Data from the online survey will also be downloaded from Opinio and stored on this machine. A data backup will be created on an encrypted USB. Uncoded data, such as the digitised questionnaires and interview transcripts will be deleted on completion of the PhD, which is scheduled for the end of 2019. **Who will have access to the data, including advisory groups and during transcription?** The Principal Investigator (Lara Smale). The PhD Project supervisor (Christopher Kilburn) will check the Italian to English translation of any open question questionnaire items.
46	**Do you confirm that all personal data will be stored and processed in compliance with the General Data Protection Regulation (GDPR 2018).** Yes ☒ No ☐ **If not,** please clarify why.
47	**Will personal data be processed or be sent outside of the European Economic Area (EEA)?*** Yes ☐ No ☒ **If yes,** please confirm that there are adequate levels of protection in compliance with the GDPR 2018 and state what the arrangements are below. ***Please note** that if you store your research data containing identifiable data on UCL systems or equipment (including by using your UCL email account to transfer data), or otherwise carry out work on your research in the UK, the processing will take place within the EEA and will be captured by Data Protection legislation.

After the project	
48	**What data will be stored and how will you keep it secure?**

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	Uncoded data, such as the digitised questionnaires and interview transcripts will be deleted on completion of the PhD, which is scheduled for the end of 2019. Coded questionnaire answers and sections of text from interviews used in the results will be stored on an encrypted USB held by the Principal Investigator. **Where will the data be stored and who will have access?** Coded data will only be accessible to the Principal Investigator (Lara Smale). **Will the data be securely deleted?** Yes ☒ No ☐ **If yes,** please state when this will occur: On completion of the PhD, which is scheduled for the end of 2019.
49	**Will the data be archived for use by other researchers?** Yes ☐ No ☒ **If yes,** please provide further details including whether researchers outside the European Economic Area will be given access.

Section G: Declaration **I confirm that the information in this form is accurate to the best of my knowledge.**	
Signature	[redacted]
Date	29/01/2019
If student: **I have met with and advised the student on the ethical aspects of this project design.** Supervisor Name: Dr. Christopher Kilburn	
Supervisor Signature:	[redacted]
Date:	29/01/2019

Signature of Head of Department (or Chair of the Departmental Ethics Committee)
Part A I have read the 'criteria of minimal risk' as defined on page 3 of the Guidelines (http://ethics.grad.ucl.ac.uk/forms/guidelines.pdf) and I recommend that this application be considered by the Chair of the UCL REC. Yes ☒ No ☐
Part B **I have discussed this project with the principal researcher who is suitably qualified to carry out this research and I approve it. I am satisfied that** (highlight as appropriate):**

1. **Data Protection registration:**
   - has been satisfactorily completed
   - has been initiated
   - is not required
2. **A risk assessment:**
   - has been satisfactorily completed
   - has been initiated
3. **Appropriate insurance arrangements are in place and appropriate sponsorship [funding] has been approved and is in place to complete the study.**

Yes ☒ No ☐

4. **A Disclosure and Barring Service check(s):**
   - has been satisfactorily completed
   - has been initiated
   - is not required

Note: Links to details of UCL's policies on the above can be found at: http://ethics.grad.ucl.ac.uk/procedures.php

****If any of the above checks are not required please clarify why below.**

A Disclosure and Barring Service check is not required as the research does not include working in 'Regulated' activity with vulnerable groups as defined by the Safeguarding Vulnerable Groups Act 2006 or in a position of trust as defined by the Rehabilitation of Offenders Act Exception Order 1975.

Name:	Professor Paul Upchurch
Signature:	[redacted]
Date:	11/02/2019

Updated 19.10.2017

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UCL Research Ethics Committee Low-Risk Research Approval

UCL RESEARCH ETHICS COMMITTEE
OFFICE FOR THE VICE PROVOST RESEARCH

11th March 2019

Dr Christopher Kilburn
Department of Earth Sciences
UCL

Dear Dr Kilburn,

Notification of Ethics Approval with Provisos
Project ID/Title: 8601/001: Communication of volcano status at Campi Flegrei caldera, Italy.

Further to your satisfactory responses to the Committee's comments, I am pleased to confirm in my capacity as Chair of the UCL Research Ethics Committee (REC) that your study has been ethically approved by the UCL REC until **11th March 2020.**

Ethical approval is subject to the following conditions:

Notification of Amendments to the Research
You must seek Chair's approval for proposed amendments (to include extensions to the duration of the project) to the research for which this approval has been given. Each research project is reviewed separately and if there are significant changes to the research protocol you should seek confirmation of continued ethical approval by completing an 'Amendment Approval Request Form' http://ethics.grad.ucl.ac.uk/responsibilities.php

Adverse Event Reporting – Serious and Non-Serious
It is your responsibility to report to the Committee any unanticipated problems or adverse events involving risks to participants or others. The Ethics Committee should be notified of all serious adverse events via the Ethics Committee Administrator (ethics@ucl.ac.uk) immediately the incident occurs. Where the adverse incident is unexpected and serious, the Joint Chairs will decide whether the study should be terminated pending the opinion of an independent expert. For non-serious adverse events the Joint Chairs of the Ethics Committee should again be notified via the Ethics Committee Administrator within ten days of the incident occurring and provide a full written report that should include any amendments to the participant information sheet and study protocol. The Joint Chairs will confirm that the incident is non-serious and report to the Committee at the next meeting. The final view of the Committee will be communicated to you.

Final Report
At the end of the data collection element of your research we ask that you submit a very brief report (1-2 paragraphs will suffice) which includes in particular issues relating to the ethical implications of the research i.e. issues obtaining consent, participants withdrawing from the research, confidentiality, protection of participants from physical and mental harm etc.

In addition, please:

Office of the Vice Provost Research, 2 Taviton Street
University College London
Tel: +44 (0)20 7679 8717
Email: ethics@ucl.ac.uk
http://ethics.grad.ucl.ac.uk/

- ensure that you follow all relevant guidance as laid out in UCL's Code of Conduct for Research: http://www.ucl.ac.uk/srs/governance-and-committees/resgov/code-of-conduct-research
- note that you are required to adhere to all research data/records management and storage procedures agreed as part of your application. This will be expected even after completion of the study.

With best wishes for the research.

Yours sincerely

[redacted]

Dr Lynn Ang
Joint Chair, UCL Research Ethics Committee

CC: Lara Smale

UCL Research Ethics Committee Low-Risk Research Amendment Approval Request Form

Amendment Approval Request Form

<table>
<tr><td>1</td><td>Project ID Number:
8601/001</td><td>Name and Address of Principal Investigator:
Lara Smale, UCL Earth Sciences, 5 Gower Place, London, WC1E 6BS (lara.smale.13@ucl.ac.uk)</td></tr>
<tr><td>2</td><td colspan="2">Project Title: Communication of volcano status at Campi Flegrei caldera, Italy.</td></tr>
<tr><td>3</td><td colspan="2">Type of Amendment/s (tick as appropriate)

Research procedure/protocol (including research instruments) ☐
Participant group ☒
Sponsorship/collaborators ☐
Extension to approval needed (extensions are given for one year) ☐
Information Sheet/s ☒
Consent form/s ☐
Other recruitment documents ☐
Principal researcher/medical supervisor* ☐
Other ☐

*Additions to the research team other than the principal researcher, student supervisor and medical supervisor do not need to be submitted as amendments but a complete list should be available upon request *</td></tr>
<tr><td>4</td><td colspan="2">Justification (give the reasons why the amendment/s are needed)

The research is part of a PhD thesis that must be submitted in October 2019. Due to time constraints and in order to meet this deadline it has been decided to reduce the number of participant groups. In the original study plan the following questionnaires about communicating the status of the Campi Flegrei volcano, Italy were intended to be distributed;

• A paper questionnaire at the Vesuvius Observatory
• An online questionnaire targeting scientists and decision makers globally that was to be shared via a LISTSERV and social media
• An online questionnaire sent to non-observatory scientists known to work on Campi Flegrei and/or those who may be involved in determining the volcano status during a future episode of unrest.
• An online questionnaire sent to Decision Makers (i.e. those responsible for civil protection decisions)
• An online questionnaire shared with members of the public via social media.

The study will now only distribute a paper questionnaire amongst scientists at the Vesuvius Observatory and send an online questionnaire to non-observatory scientists known to work on Campi Flegrei and/or those who may be involved in determining the volcano status during a future episode of unrest. These groups formed the core of the original study plan.

The information sheet has been minimally altered to reflect the change in the number of participant groups and also the removal of a section in the questionnaire regarding volcano alert levels to reduce the length of the questionnaire.</td></tr>
<tr><td>5</td><td colspan="2">Details of Amendments (provide full details of each amendment requested, state where the changes have been made and attach all amended and new documentation)

• The study will now no longer share an online survey targeting scientists and decision makers globally using LISTSERV and social media</td></tr>
</table>

	• The study will no longer distribute surveys to Decision Makers or Members of the Public. • Three changes have been made to the information sheets for scientist questionnaire. First, the second sentence of the second paragraph ("What is the purpose of the study") has been changed from "*We are interested in improving our understanding the challenges of interpreting the causes of such periods of unrest and communicating the status of the volcano between Scientists, Decision Makers (those responsible for emergency response decisions), the Media and the Public*" to "*We are interested in improving our understanding of the challenges of interpreting the causes of such periods of unrest and communicating the status of the volcano*". Second, in the third paragraph ("Taking part in the study") the fourth, fifth and sixth sentences have been changed from "*It then asks for your thoughts on Volcano Alert Level systems and how to communicate the status of the volcano. Finally, it asks for your perceptions of other stakeholders (e.g. Decision Makers, the Media and the Public). It will take up to 10 minutes to complete*" to "*It then asks for your thoughts on your perceptions of other stakeholder groups understanding of unrest. It will take 5 to 10 minutes to complete*". Third, in the Local Data Protection Privacy Notice the categories "job category" and "education level" have been removed as they are no longer included in the questionnaire. • The following change has been made to the information sheet for the interviews with the Vesuvius Observatory Scientists; the second sentence of the second paragraph has changed from "*We are interested in improving our understanding the challenges of interpreting the causes of such periods of unrest and communicating the status of the volcano between Scientists, Decision Makers (those responsible for emergency response decisions), the Media and the Public.*" To "*We are interested in improving our understanding the challenges of interpreting the causes of such periods of unrest and communicating the status of the volcano".*
6	**Ethical Considerations** (insert details of any ethical issues raised by the proposed amendment/s) None have been identified. No data collection has been started as yet and no additional data is to be requested.
7	**Other Information** (provide any other information which you believe should be taken into account during ethical review of the proposed changes) N/A

Declaration (to be signed by the Principal Researcher)

- I confirm that the information in this form is accurate to the best of my knowledge and I take full responsibility for it.
- I consider that it would be reasonable for the proposed amendments to be implemented.
- For student projects, I confirm that my supervisor has approved my proposed modifications.

Signature:
[redacted]

Date: 15th May 2019

UCL Research Ethics Committee Low-Risk Research Amendment Approval

VPRO.Ethics
30 May 2019 at 12:26
Details
APPROVED Amendment Approval Request for Project 8601/001
To: Smale, Lara, Cc: Kilburn, Christopher

Siri found new contact info in this email: Ethics Vpro. ethics@ucl.ac.uk add to Contacts...

Lara, I am pleased to confirm that your amendment request has been approved by the UCL REC Chair. Please take this email as confirmation of that approval.

IMPORTANT: For projects collecting personal data only
You should inform the Data Protection Team – data-protection@ucl.ac.uk of your proposed amendments to include a request to extend ethics approval for an additional period.
With best wishes for your ongoing research, Helen
Helen Dougal
UCL Research Ethics Administrator
Office of the Vice-Provost (Research)
University College London
2 Taviton Street, London, WC1H 0BT
Email: ethics@ucl.ac.uk

Please note that I work from home on Tuesdays. My contact details for each day are as follows:
Monday/Wednesday/Thursday/Friday: 020 7679 8717 | (Internal extension 28717)
Tuesday: [redacted]

(D) Surveys and participant information sheets for data collection used in Chapter 7

This Appendix includes the survey distributed at the Vesuvius Observatory in June 2019 as part of data collection used in Chapter 7 of the thesis, the online version of the survey and the information sheets that were given to participants detailing the nature of the research and the intended use of the results.

Participant Information Sheet for the Survey Conducted the Vesuvius Observatory, June 2019

LONDON'S GLOBAL UNIVERSITY

Foglio informativo per gli scienziati dell'Osservatorio Vesuviano partecipanti

Comunicazione dello stato delle caldere ai Campi Flegrei, Italia

UCL Research Ethics Committee (Comitato Etico di UCL) Numero identificativo del progetto: 8601/001

La invitiamo a prendere parte alla ricerca. Prima di decidere se vuole prenderne parte, tuttavia, vorremmo spiegarLe lo scopo di questa ricerca, e cosa può aspettarsi dalla stessa. La invitiamo a chiedere personalmente ai ricercatori oppure utilizzare i loro contatti indicati a fine pagina se desidera ricevere più informazioni o dovessero sorgere domande. Prendetevi tutto il tempo necessario per decidere se partecipare o meno. La ringraziamo di aver letto questa.

Qual è lo scopo della ricerca?
Grandi caldera vulcani, come i Campi Flegrei, spesso attraversano periodi di unrest vulcanica caratterizzati da un sollevamento del suolo che è solitamente accompagnato da sismicità. Ci focalizziamo sul miglioramento della nostra conoscenza delle difficoltà nell'interpretazione delle cause di questi periodi di unrest e nella comunicazione dello stato attuale del vulcano. Questa ricerca fa parte di uno studio relativo ai Campi Flegrei per una tesi di Dottorato presso il Dipartimento di Scienze della Terra, University College London (UCL) nel Regno Unito.

Partecipare alla ricerca
Il questionario sarà accessibile a tutti gli scienziati dell'Osservatorio Vesuviano, ma la partecipazione è volontaria e non è necessaria una preesistente conoscenza dei Campi Flegrei. Nel caso in cui decida di prenderne parte, allora Le verrà chiesto di completare il questionario. La prima parte contiene domande generali, relative alla Sua esperienza, dopodiché Le verrà chiesto di esprimere la Sua opinione riguardante la causa dell'unrest ai Campi Flegrei e la sua conoscenza della nozione di unrest. Infine Le chiederemo la Sua opinione riguardante la conoscenza della nozione di unrest da parte di soggetti interessati. Le basteranno dai 5 ai 10 minuti per completarlo. Se preferite non rispondere a certe domande, le potrete ignorare.

Il questionario non propone informazioni direttamente identificabili, come nomi o contatti, e i partecipanti non saranno identificabili nei risultati. Una volta consegnato il questionario completato, questo non potrà essere ritirato, ma Le è consentito di abbandonare il questionario in ogni momento senza dover dare spiegazioni.

Come verranno utilizzate le mie risposte?
Se decide di completare il questionario, le Sue risposte contribuiranno allo sviluppo della ricerca, parte di un capitolo di tesi di Dottorato, che sarà conclusa a termine dell'anno corrente 2019. I risultati saranno pubblicati in una rivista scientifica oggetto di esame inter pares. I dati raccolti saranno trasferiti in forma digitale e salvati all'interno di un dispositivo protetto presso il Dipartimento di Scienze della Terra, all'University College London (UCL). Le copie cartacee originali saranno eliminate una volta completato il Dottorato. Se desidera ricevere una copia dei risultati una volta completato il Dottorato, La invitiamo a contattare i ricercatori utilizzando i contatti riportati di seguito.

Attraverso il completamento e la presentazione del questionario, Lei conferma che comprende le informazioni sopra riportate e acconsente al utilizzo delle Sue risposte. Può conservare questo foglio informativo in caso di necessità futura.

University College London, Gower Street, London WC1E 6BT
www.ucl.ac.uk

Informativa sulla protezione dei dati locali sulla privacy

Nota sulla privacy:
University College London (UCL) è l'organo regolatore di questo progetto. Il Responsabile della Protezione dei Dati presso UCL opera come supervisore di ogni attività presso l'Università stessa, compresa la rielaborazione di dati personali, e può essere contattato all'indirizzo mail data-protection@ucl.ac.uk.

Questa nota sulla privacy "locale" illustra le informazioni relative a questa ricerca specifica. Ulteriori informazioni relative all'utilizzo delle informazioni dei partecipanti da parte di University College London possono essere trovate nella nostra nota sulla privacy "generale":

Per i partecipanti agli studi di ricerca, le informazioni possono essere trovate qui:
https://www.ucl.ac.uk/legal-services/privacy/ucl-general-research-participant-privacy-notice

Secondo il Codice di protezione dei dati (GDPR e DPA2018), tutte le informazioni che sono fornite ai partecipanti sono presenti nelle note di privacy sia "locali", sia "generali".

Le categorie di dati personali utilizzate saranno le seguenti:

Luogo di lavoro
Fascia di età
Area di specializzazione
Esperienze passate di crisi vulcaniche e simulazioni di crisi
Esperienza passata consigliando Protezione Civile

La base giuridica per il trattamento dei Suoi dati sarà lo svolgimento di un compito nell'interesse civile.

I Suoi dati personali saranno processati fino al completamento del PhD sopra menzionato. Ci impegneremo a rendere anonimi o usare pseudonimi per i Suoi dati personali, minimizzando la rielaborazione di dati personali ovunque sia possibile.

Se avesse dubbi sul processo di rielaborazione dei Suoi dati, o se desidera contattarci in relazione ai Suoi diritti, non esiti a contattare UCL in qualsiasi momento all'indirizzo data-protection@ucl.ac.uk.

Questa ricerca fa parte di una borsa di studio per studente di Dottorato presso il Dipartimento di Scienze della Terra, University College London (UCL), è finanziata dal Natural Environment Research Council (NERC) (Consiglio per la ricerca sull'ambiente naturale), e non connessa alle attivitá dell'Istituto Nazionale di Geofisica e Vulcanologia (INGV).

Contatti

In presenza di domande, se desidera ricevere più informazioni sulla ricerca o richiedere una copia dei risultati, La preghiamo gentilmente di contattare i ricercatori utilizzando i contatti riportati di seguito.

Ricercatore:
Lara Smale

Tel: [redacted]
Email: lara.smale.13@ucl.ac.uk

Ricercatore Principale:
Dr. Christopher Kilburn

Email: c.kilburn@ucl.ac.uk

La ringraziamo di prendere in considerazione una collaborazione a questa ricerca.

Survey for the Vesuvius Observatory

Sezione 1

1. Età
   (18-30) (31-40) (41-50) (51-60) (61+)

2. Dove lavori? (es. sezione INGV, università)___________________________________

3. Quale delle seguenti opzioni descrive meglio il suo settore di competenza?

☐ Geologia (Scienze della Terra)
☐ Geofisica (Geodesia)
☐ Geofisica (Sismologia)
☐ Geochimica
☐ Reti di Monitoraggio
☐ Educazione al rischio vulcanico
☐ Altro (specificare) ____

4. Quanti anni ha studiato i seguenti vulcani? Spazi vuoti per possibili aggiunte.

Vulcano	Nessuno	Meno di 1	Da 1 a 5	Da 6 a 10	Più di 10
Campi Flegrei					
Vesuvio					
Ischia					
Stromboli					
Etna					

5. Ha mai riscontrato una crisi vulcanica? Se sì, in quale ruolo?

Vulcano	Anno	Cittadino	Scienziato osservatore	Consulente per la protezione civile	Altro

1

6. Nel caso di future emergenze vulcaniche ai Campi Flegrei, si aspetta di essere contattato/a in qualità di consulente per la Protezione Civile riguardo lo stato attuale del vulcano e la potenziale evoluzione dell'unrest vulcanica?

☐ No
☐ Sì

7. Valuti il livello di conoscenza scientifica del comportamento dei seguenti vulcani italiani.
   1 = molto scarso, 7 = molto buono.

	1	2	3	4	5	6	7	Non saprei
Campi Flegrei								
Vesuvio								
Ischia								
Stromboli								
Etna								

8. Valuti il livello di conoscenza scientifica del comportamento delle seguenti vulcani.
   1 = molto scarso, 7 = molto buono.

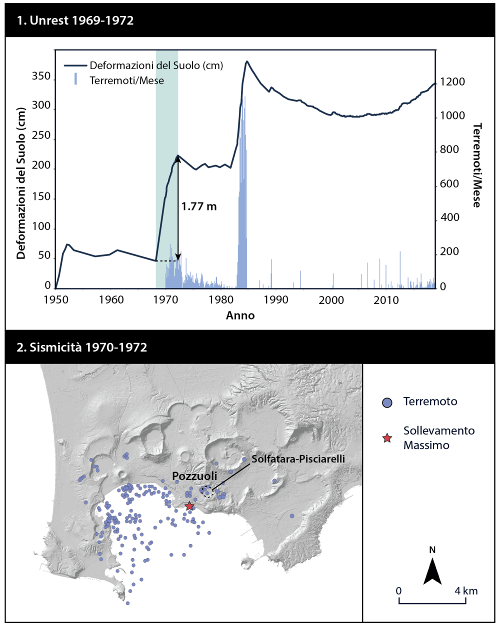
	1	2	3	4	5	6	7	Non saprei
Campi Flegrei								
Long Valley Caldera (USA)								
Nisyros (Grecia)								
Yellowstone (USA)								
Rabaul, Papua Nuova Guinea								

2

Sezione 2

Questa sezione propone domande sulla bradisismicità dei Campi Flegrei. Nel caso possano risultare utili, si inseriscono qui riassunti di episodi specifici di movimento del suolo.

La massima deformazione durante il sollevamento accaduto fra il 1969 e 1972 fu di 1.77 m. Successivamente all'installazione dei sismometri nel marzo 1970 si registrarono all'incirca 5000 terremoti tra 1 e 5 km di profondità con massima magnitudine di M 2.5. Un'intensificazione di attività fumarolica si osservò nell'area di Solfatara-Pisciarelli. Una volta terminato il processo di sollevamento ci fu un'immediata subsidenza di circa 0.2 m, dopo la quale il livello del terreno oscillò fino al 1982. La sismicità declinò lentamente di seguito al sollevamento.



3

9. Secondo Lei, quale fu la causa più probabile del sollevamento del 1969-1972?

- ☐ Intrusione del magma
- ☐ Iniezione di fluidi magmatici sotto ad uno strato impermeabile dei crosta
- ☐ Gonfiamento del sistema idrotermale causato da un'iniezione di fluidi magmatici
- ☐ Sia intrusione magmatica, sia gonfiamento del sistema idrotermale
- ☐ Altro____
- ☐ Non saprei

Perché?____

____

____

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____

Approssimativamente, secondo Lei, a quale(i) profondità nella crosta era localizzata la sorgente (o le sorgenti) di pressione? Selezioni.

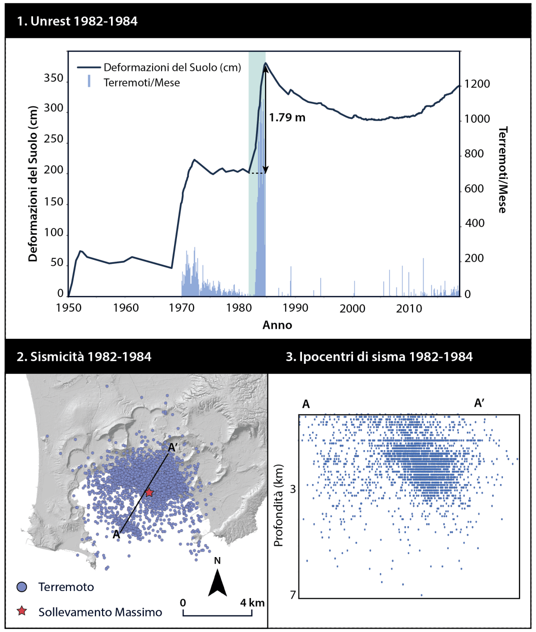
(0-1) (1-2) (2-3) (4-5) (5-6) (6-7) (7+) (Non Saprei)

10. Secondo Lei, quanto concordano i seguenti gruppi riguardo la causa del sollevamento del suolo in 1969-1972. 1 = totale disaccordo, 7 = totale accordo.

	1	2	3	4	5	6	7	Non saprei
Comunità scientifica di Napoli								
Comunità scientifica italiana								
Comunità scientifica internazionale								

4

Il massimo sollevamento del suolo nel 1982-1984 fu di 1.79 m. Più di 16000 terremoti con magnitudine tra 0.2 e 4.2 furono registrati fra 0 e 7 km di profondità, la maggioranza dei quali (circa 80%) avevano magnitudine minore di M 2. La più alta frequenza di eventi si verificò sotto Pozzuoli. Durante il sollevamento ci fu un'intensificazione di attività fumarolica a Solfatara e un incremento iniziale nel contenuto di H_2S, CH_4 e H_2O nei gas fumarolici, seguito da un incremento nel rapporto CO_2/H_2O. La sismicità terminò improvvisamente una volta concluso il sollevamento.



5

11. Secondo Lei, quale fu la causa più probabile del sollevamento del 1982-1984?

☐ Intrusione del magma
☐ Iniezione di fluidi magmatici sotto ad uno strato impermeabile dei crosta
☐ Gonfiamento del sistema idrotermale causato da un'iniezione di fluidi magmatici
☐ Sia intrusione magmatica, sia gonfiamento del sistema idrotermale
☐ Altro____
☐ Non saprei

Perché?____

Approssimativamente, secondo Lei, a quale(i) profondità nella crosta era localizzata la sorgente (o le sorgenti) di pressione? Selezioni.

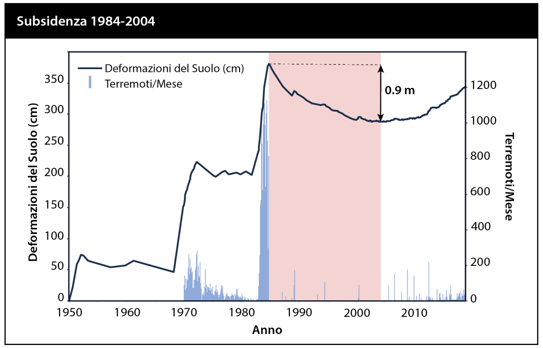
(0-1) (1-2) (2-3) (4-5) (5-6) (6-7) (7+) (Non Saprei)

12. Secondo Lei, quanto concordano i seguenti gruppi riguardo la causa del sollevamento del suolo in 1982-1984. 1 = totale disaccordo, 7 = totale accordo.

	1	2	3	4	5	6	7	Non saprei
Comunità scientifica di Napoli								
Comunità scientifica italiana								
Comunità scientifica internazionale								

6

Concluso il sollevamento degli anni 1982-1984, inizio una subsidenza senza sismicità che durò fino al 2004, abbassando il livello del suolo di 0.9 m. Durante questo periodo ci fu un generale declino nel rapporto CO_2/H_2O dei gas fumarolici di Solfatara.



- ☐ Solidificazione di un'intrusione del magma
- ☐ Fuoriuscita di fluidi magmatici da sotto uno strato impermeabile nella crosta
- ☐ Riduzione della pressione nel sistema idrotermale
- ☐ Altro____
- ☐ Non saprei

Perché?____

Approssimativamente, secondo Lei, a quale(i) profondità nella crosta era localizzata la sorgente (o le sorgenti) di pressione? Selezioni.

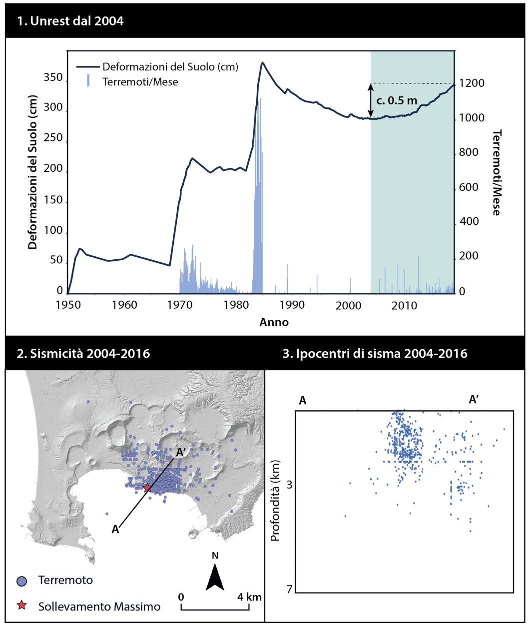
(0-1) (1-2) (2-3) (4-5) (5-6) (6-7) (7+) (Non Saprei)

13. Secondo Lei, quanto concordano i seguenti gruppi riguardo la causa della subsidenza? 1 = totale disaccordo, 7 = totale accordo.

	1	2	3	4	5	6	7	Non saprei
Comunità scientifica di Napoli								
Comunità scientifica italiana								
Comunità scientifica internazionale								

(Continua alla pagina successiva)

Dal 2004, il sollevamento del suolo ha portato ad una massima deformazione di circa +0.5 m fino ad oggi. Circa 2000 terremoti sono stati registrati sotto Pozzuoli, principalmente tra 1 e 2.5 km di profondità, con magnitudine tra M -2 a 2.5. Durante il sollevamento c'è stato un incremento nel degassaggio a Solfatara-Pisciarelli e un continuo incremento nel rapporto CO_2/H_2O dei gas fumarolici.



9

14. Secondo Lei, quale fu la causa più probabile del sollevamento dal 2004?

☐ Intrusione del magma
☐ Iniezione di fluidi magmatici sotto ad uno strato impermeabile dei crosta
☐ Gonfiamento del sistema idrotermale causato da un'iniezione di fluidi magmatici
☐ Sia intrusione magmatica, sia gonfiamento del sistema idrotermale
☐ Altro______________________________
☐ Non saprei

Perché?______________________________

Approssimativamente, secondo Lei, a quale(i) profondità nella crosta era localizzata la sorgente (o le sorgenti) di pressione? Selezioni.

(0-1) (1-2) (2-3) (4-5) (5-6) (6-7) (7+) (Non Saprei)

15. Secondo Lei, quanto concordano i seguenti gruppi riguardo la causa del sollevamento. 1 = totale disaccordo, 7 = totale accordo.

	1	2	3	4	5	6	7	Non saprei
Comunità scientifica di Napoli								
Comunità scientifica italiana								
Comunità scientifica internazionale								

10

16. Quale ritiene sia lo scenario finale più probabile del sollevamento attuale?

____
____
____
____
____
____

17. Secondo Lei, qual è il rischio maggiore che si possa associare al sollevamento attuale?

____
____
____
____
____
____
____

18. A quale profondità approssimativa, secondo Lei, si potrebbe trovare del magma nella crosta dei Campi Flegrei al giorno d'oggi? Selezioni.

☐ 1-2 km
☐ 2-3 km
☐ 3-4 km
☐ 4-5 km
☐ 5-6 km
☐ >7 km
☐ Non saprei

19. Quali cambiamenti dei parametri di monitoraggio vi aspettereste in caso d'un'intrusione di magma ai Campi Flegrei?

____
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20. Qual è secondo Lei il criterio più utili che determina se l'unrest sia magmatica o non magmatica in origine? Ci sono raccolte dati o strumenti non utilizzati al momento che potrebbero perfezionare le interpretazioni future?

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Sezione 3

21. Valuti il livello di conoscenza del comportamento dei Campi Flegrei da parte dei seguenti gruppi. 1 = molto scarso, 7 = molto buono.

	1	2	3	4	5	6	7	Non saprei
Protezione Civile								
Autorità Locali								
Popolazione dei Campi Flegrei								
I media								

22. A Suo avviso, qual è la causa dell'attuale sollevamento secondo i gruppi seguenti?

	Magma Intrusione	Altrei	Non saprei
Protezione Civile			
Autorità Locali			
Popolazione dei Campi Flegrei			
I media			

23. Secondo Lei, quali sono i rischi più preoccupanti durante il sollevamento attuale secondo i gruppi seguenti?

Protezione Civile____

Autorità Locali____

Popolazione dei Campi
Flegrei____

I Media____

24. Ci sono difficoltà nella comunicazione dell'attuale comportamento del vulcano tra gli scienziati al giorno d'oggi, la Protezione Civile, le autorità locali, i media e i civili? Se si, quali?

□ No
□ Sì
□ Non saprei

____
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____
____

25. Secondo Lei, nel caso di una futura emergenza vulcanica ai Campi Flegrei, quanto bene i seguenti gruppi capiranno le informazioni ricevute riguardanti il vulcano? 1 = molto male, 7 = molto bene

	1	2	3	4	5	6	7	Non saprei
Protezione Civile								
Autorità Locali								
Popolazione dei Campi Flegrei								
I media								

26. Nel caso di future emergenze vulcaniche ai Campi Flegrei, quali potrebbero essere secondo Lei le difficoltà relative alla comunicazione della causa di unrest e dello stato attuale del vulcano tra gli scienziati, la Protezione Civile, le autorità locali, i media e i civili? Se si, quali?

□ No
□ Sì
□ Non saprei

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La ringrazio molto per la Sua partecipazione.

Se vuole saperne di più sul risultato, La invito a contattare Lara Smale all'indirizzo lara.smale.13@ucl.ac.uk.

Participant Information Sheet for the Online Survey Conducted in July 2019

LONDON'S GLOBAL UNIVERSITY

Foglio informativo per partecipanti

Comunicazione dello stato delle caldere ai Campi Flegrei, Italia

UCL Research Ethics Committee (Comitato Etico di UCL) Numero identificativo del progetto: 8601/001

La invitiamo a prendere parte alla ricerca. Prima di decidere se vuole prenderne parte, tuttavia, vorremmo spiegarLe lo scopo di questa ricerca, e cosa può aspettarsi dalla stessa. La invitiamo a contattare i ricercatori utilizzando i contatti riportati di seguitose desidera ricevere più informazioni o dovessero sorgere domande. Prendetevi tutto il tempo necessario per decidere se partecipare o meno. La ringraziamo di aver letto questa.

Qual è lo scopo della ricerca?
Grandi caldera vulcani, come i Campi Flegrei, spesso attraversano periodi di unrest vulcanica caratterizzati da un sollevamento del suolo che è solitamente accompagnato da sismicità. Ci focalizziamo sul miglioramento della nostra conoscenza delle difficoltà nell'interpretazione delle cause di questi periodi di unrest e nella comunicazione dello stato attuale del vulcano. Questa ricerca fa parte di uno studio relativo ai Campi Flegrei per una tesi di Dottorato presso il Dipartimento di Scienze della Terra, University College London (UCL) nel Regno Unito.

Partecipare alla ricerca
La partecipazione è volontaria e non è necessaria una preesistente conoscenza dei Campi Flegrei. Se decide di prenderne parte, allora Le verrà chiesto di completare il questionario corrispondente. La prima parte contiene domande generali, relative alla Sua esperienza, dopodiché Le verrà chiesto di esprimere la Sua opinione riguardante la causa dell'unrest ai Campi Flegrei e la Sua conoscenza della nozione di unrest.
Infine Le chiederemo la Sua opinione riguardante la conoscenza di unrest da parte di gruppi azionistici sul territorio. Le basteranno dai 5 ai 10 minuti per completarlo. Se preferite non rispondere a certe domande, sentitevi liberi di ignorarle.

Il questionario non propone informazioni direttamente identificabili, come nomi o contatti, e i partecipanti non saranno identificabili nei risultati. Una volta consegnato il questionario completato, questo non potrà essere ritirato, ma Le è consentito di abbandonare il questionario in ogni momento senza dover dare spiegazioni.

Come verranno utilizzate le mie risposte?
Se decide di completare il questionario, le Sue risposte contribuiranno allo sviluppo della ricerca, parte di un capitolo di tesi di Dottorato, che sarà conclusa a termine dell'anno corrente 2019. I risultati saranno pubblicati in una rivista scientifica oggetto di esame inter pares. I dati raccolti saranno salvati all'interno di un dispositivo protetto presso il Dipartimento di Scienze della Terra, all'University College London (UCL). I dati saranno eliminate una volta completato il Dottorato. Se desidera ricevere una copia dei risultati una volta completato il Dottorato, La invitiamo a contattare i ricercatori utilizzando i contatti riportati di seguito.

Attraverso il completamento e la presentazione del questionario, Lei conferma che comprende le informazioni sopra riportate e acconsente al utilizzo delle Sue risposte. Può conservare questo foglio informativo in caso di necessità futura.

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Area di specializzazione
Esperienze passate di crisi vulcaniche e simulazioni di crisi
Esperienza passata consigliando Protezione Civile

La base giuridica per il trattamento dei Suoi dati sarà lo svolgimento di un compito nell'interesse civile.

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Questa ricerca fa parte di una borsa di studio per studente di Dottorato presso il Dipartimento di Scienze della Terra, University College London (UCL), è finanziata dal Natural Environment Research Council (NERC) (Consiglio per la ricerca sull'ambiente naturale), e non connessa alle attivitá dell'Istituto Nazionale di Geofisica e Vulcanologia (INGV).

Contatti

In presenza di domande, se desidera ricevere più informazioni sulla ricerca o richiedere una copia dei risultati, La preghiamo gentilmente di contattare i ricercatori utilizzando i contatti riportati di seguito.

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Ricercatore Principale:
Prof. Christopher Kilburn

Email: c.kilburn@ucl.ac.uk

La ringraziamo di prendere in considerazione una collaborazione a questa ricerca.

Participant Information Sheet for the Online Survey Conducted in July 2019

Grandi caldere vulcaniche, come i Campi Flegrei, spesso sperimentano periodi di unrest. Ci focalizziamo sul miglioramento della nostra conoscenza delle sfide che si possono riscontrare nell'interpretazione delle cause di questi periodi di unrest e nella comunicazione dello stato attuale del vulcano. Questa ricerca fa parte di uno studio condotto relativo ai Campi Flegrei per una tesi di Dottorato presso il University College London (UCL) nel Regno Unito.

Il questionario non propone informazioni direttamente identificabili, come nomi o contatti, e i partecipanti non saranno identificabili nei risultati. Una volta consegnato il questionario completato, questo non potrà essere ritirato, ma Le è consentito di abbandonare il questionario in ogni momento senza dover dare spiegazioni. Ci vorranno circa 5 minuti per essere completati (in italiano).

Una foglio informative che spiega il questionario e la nota sulla privacy è allegata all'e-mail di invito e si trova anche a questo link:

Foglio Informativo

Se decide di completare il questionario, le Sue risposte contribuiranno allo sviluppo della ricerca, parte di un capitolo di tesi di Dottorato, che sarà conclusa a termine dell'anno corrente 2019. I risultati saranno pubblicati in una rivista scientifica oggetto di esame inter pares.

Attraverso il completamento e la presentazione del questionario, Lei conferma che comprende le informazioni sopra riportate, la foglio informativo, e acconsente al riutilizzo delle Sue risposte.

In presenza di domande, se desidera ricevere più informazioni sulla ricerca o richiedere una copia dei risultati, La preghiamo gentilmente di contattare i ricercatori utilizzando Lara.smale.13@ucl.ac.uk.

1. **Età**

   ◯ 18-30 ◯ 31-40 ◯ 41-50 ◯ 51-60 ◯ 61+

2. **Dove lavori? (es. sezione INGV, università)**

3. **Quale delle seguenti opzioni descrive meglio il suo settore di competenza?**

   - ☐ Geologia (Scienze della Terra)
   - ☐ Geofisica (Geodesia)
   - ☐ Geofisica (Sismologia)
   - ☐ Geochimica
   - ☐ Reti di Monitoraggio
   - ☐ Educazione al rischio vulcanico
   - ☐ Altro (specificare)

9%

4. **Quanti anni ha studiato i seguenti vulcani? Spazi vuoti per possibili aggiunte.**

Vulcano	Nessuno	Meno di 1	Da 1 a 5	Da 6 a 10	Più di 10
Campi Flegrei	☐	☐	☐	☐	☐
Vesuvio	☐	☐	☐	☐	☐
Ischia	☐	☐	☐	☐	☐
Stromboli	☐	☐	☐	☐	☐
Etna	☐	☐	☐	☐	☐
	☐	☐	☐	☐	☐
	☐	☐	☐	☐	☐
	☐	☐	☐	☐	☐

5. **Ha mai riscontrato una crisi vulcanica? Se sì, in quale ruolo?**

Vulcano	Anno	Cittadino	Scienziato osservatore	Consulente per la protezione civile	Altro
		☐	☐	☐	☐
		☐	☐	☐	☐
		☐	☐	☐	☐
		☐	☐	☐	☐
		☐	☐	☐	☐

6. **Nel caso di future emergenze vulcaniche ai Campi Flegrei, si aspetta di essere contattato/a in qualità di consulente per la Protezione Civile riguardo lo stato attuale del vulcano e la potenziale evoluzione dell'unrest vulcanica?**

- ○ No
- ○ Si

19%

7. **Valuti il livello di conoscenza scientifica del comportamento dei seguenti vulcani italiani. 1 = molto scarso, 7 = molto buono.**

	1	2	3	4	5	6	7	Non Saprei
Campi Flegrei	○	○	○	○	○	○	○	○
Vesuvio	○	○	○	○	○	○	○	○
Ischia	○	○	○	○	○	○	○	○
Stromboli	○	○	○	○	○	○	○	○
Etna	○	○	○	○	○	○	○	○

8. **Valuti il livello di conoscenza scientifica del comportamento delle seguenti vulcani. 1 = molto scarso, 7 = molto buono.**

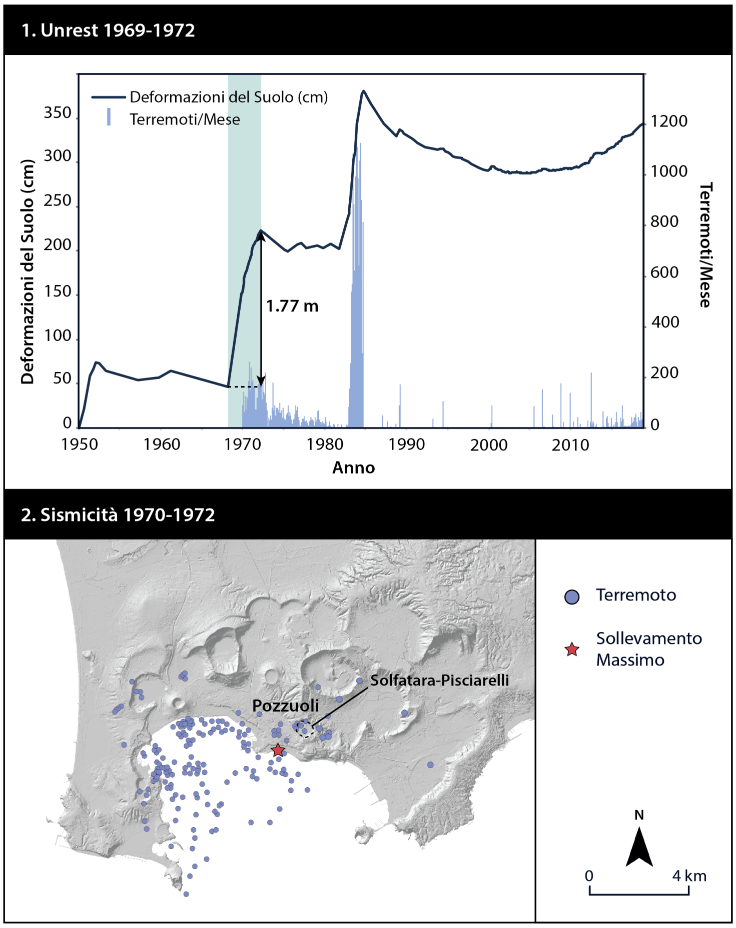
	1	2	3	4	5	6	7	Non Saprei
Long Valley Caldera (USA)	○	○	○	○	○	○	○	○
Nisyros (Grecia)	○	○	○	○	○	○	○	○
Yellowstone (USA)	○	○	○	○	○	○	○	○
Rabaul, Papua Nuova Guinea	○	○	○	○	○	○	○	○

25%

Sezione 2

Questa sezione propone domande sulla bradisismicità dei Campi Flegrei. Nel caso possano risultare utili, si inseriscono qui riassunti di episodi specifici di movimento del suolo.

La massima deformazione durante il sollevamento accaduto fra il 1969 e 1972 fu di 1.77 m. Successivamente all'installazione dei sismometri nel marzo 1970 si registrarono all'incirca 5000 terremoti tra 1 e 5 km di profondità con massima magnitudine di M 2.5. Un'intensificazione di attività fumarolica si osservò nell'area di Solfatara-Pisciarelli. Una volta terminato il processo di sollevamento ci fu un'immediata subsidenza di circa 0.2 m, dopo la quale il livello del terreno oscillò fino al 1982. La sismicità declinò lentamente di seguito al sollevamento.



9. **Secondo Lei, quale fu la causa più probabile del sollevamento del 1969-1972?**

- ☐ Intrusione del magma
- ☐ Iniezione di fluidi magmatici sotto ad uno strato impermeabile dei crosta
- ☐ Gonfiamento del sistema idrotermale causato da un'iniezione di fluidi magmatici
- ☐ Sia intrusione magmatica, sia gonfiamento del sistema idrotermale
- ☐ Non saprei
- ☐ Altro

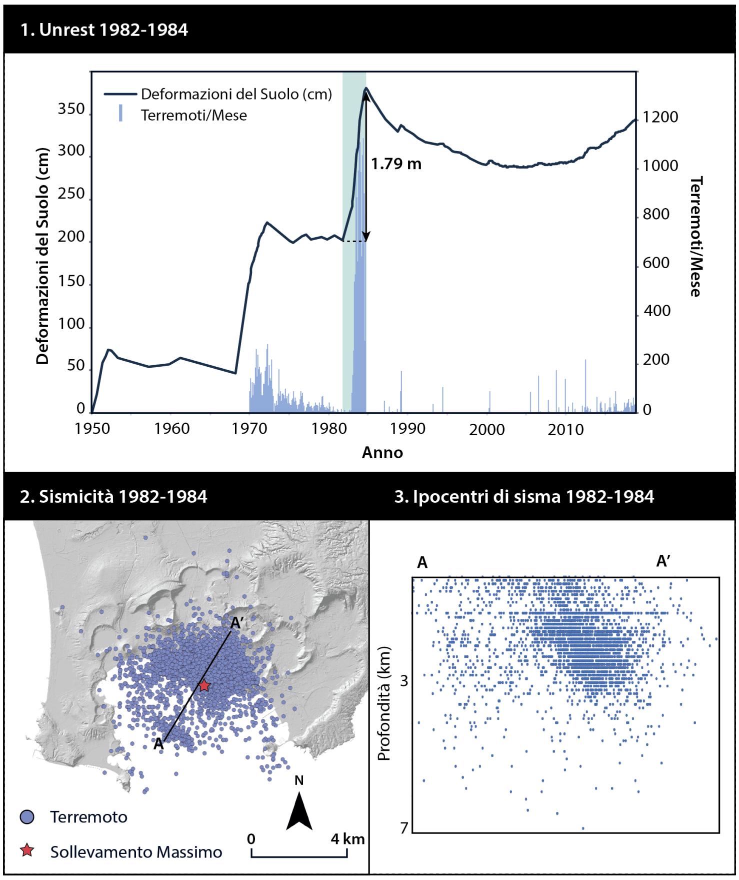
10. **Perché?**

11. **Approssimativamente, secondo Lei, a quale(i) profondità nella crosta era localizzata la sorgente (o le sorgenti) di pressione? Selezioni.**

☐ 0-1 km ☐ 1-2 km ☐ 2-3 km ☐ 3-4 km ☐ 4-5 km ☐ 5-6 km ☐ 6-7 km ☐ >7 km ☐ Non Saprei

35%

Il massimo sollevamento del suolo nel 1982-1984 fu di 1.79 m. Più di 16000 terremoti con magnitudine tra 0.2 e 4.2 furono registrati fra 0 e c. 7 km di profondità, la maggioranza dei quali (circa 80%) avevano magnitudine minore di M 2. La più alta frequenza di eventi si verificò sotto Pozzuoli. Durante il sollevamento ci fu un'intensificazione di attività fumarolica a Solfatara e un incremento iniziale nel contenuto di H_2S, CH_4 e H_2O nei gas fumarolici, seguito da un incremento nel rapporto CO_2/H_2O. La sismicità terminò improvvisamente una volta concluso il sollevamento.



12. **Secondo Lei, quale fu la causa più probabile del sollevamento del 1982-1984?**

- ☐ Intrusione del magma
- ☐ Iniezione di fluidi magmatici sotto ad uno strato impermeabile dei crosta

- ☐ Gonfiamento del sistema idrotermale causato da un'iniezione di fluidi magmatici
- ☐ Sia intrusione magmatica, sia gonfiamento del sistema idrotermale
- ☐ Non saprei
- ☐ Altro

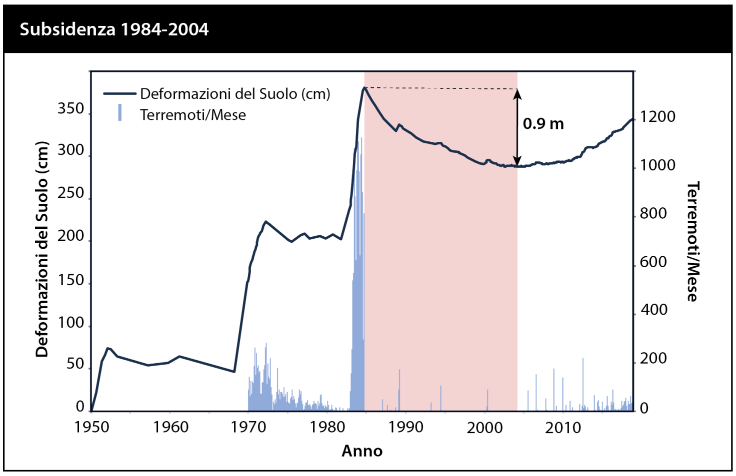
13. **Perché?**

14. **Approssimativamente, secondo Lei, a quale(i) profondità nella crosta era localizzata la sorgente (o le sorgenti) di pressione? Selezioni.**

☐ 0-1 km ☐ 1-2 km ☐ 2-3 km ☐ 3-4 km ☐ 4-5 km ☐ 5-6 km ☐ 6-7 km ☐ >7 km ☐ Non Saprei

45%

Concluso il sollevamento degli anni 1982-1984, inizio una subsidenza senza sismicità che durò fino al 2004, abbassando il livello del suolo di 0.9 m. Durante questo periodo ci fu un generale declino nel rapporto CO_2/H_2O dei gas fumarolici di Solfatara.



15. **Secondo Lei, quale fu la causa più probabile della subsidenza?**

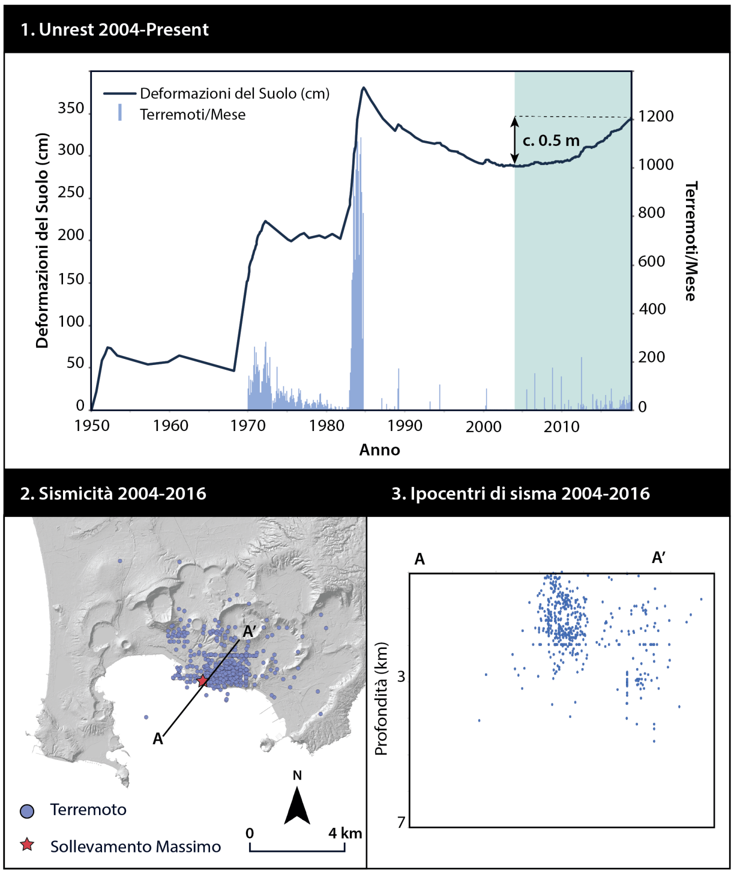
- ☐ Solidificazione di un'intrusione del magma
- ☐ Fuoriuscita di fluidi magmatici da sotto uno strato impermeabile nella crosta
- ☐ Riduzione della pressione nel sistema idrotermale
- ☐ Non saprei
- ☐ Altro

16. **Perché?**

17. **Approssimativamente, secondo Lei, a quale(i) profondità nella crosta era localizzata la sorgente (o le sorgenti) di pressione? Selezioni.**

☐ 0-1 km ☐ 1-2 km ☐ 2-3 km ☐ 3-4 km ☐ 4-5 km ☐ 5-6 km ☐ 6-7 km ☐ >7 km ☐ Non Saprei

Dal 2004, il sollevamento del suolo ha portato ad una massima deformazione di circa +0.5 m fino ad oggi. Circa 2000 terremoti sono stati registrati sotto Pozzuoli, principalmente tra 1 e 2.5 km di profondità, con magnitudine tra M -2 a 2.5. Durante il sollevamento c'è stato un incremento nel degassaggio a Solfatara-Pisciarelli e un continuo incremento nel rapporto CO_2/H_2O dei gas fumarolici.



18. **Secondo Lei, quale fu la causa più probabile del sollevamento dal 2004?**

- ☐ Intrusione del magma
- ☐ Iniezione di fluidi magmatici sotto ad uno strato impermeabile dei crosta
- ☐ Gonfiamento del sistema idrotermale causato da un'iniezione di fluidi magmatici
- ☐ Sia intrusione magmatica, sia gonfiamento del sistema idrotermale

- [ ] Non saprei
- [ ] Altro

19. Perché?

20. **Approssimativamente, secondo Lei, a quale(i) profondità nella crosta era localizzata la sorgente (o le sorgenti) di pressione? Selezioni.**

- [ ] 0-1 km
- [ ] 1-2 km
- [ ] 2-3 km
- [ ] 3-4 km
- [ ] 4-5 km
- [ ] 5-6 km
- [ ] 6-7 km
- [ ] >7 km
- [ ] Non Saprei

64%

21. **Quale ritiene sia lo scenario finale più probabile del sollevamento attuale?**

22. **Secondo Lei, quali sono i pericoli e rischi maggiore che si possa associare al sollevamento attuale?**

23. **Secondo Lei, quanto concordano i seguenti gruppi riguardo la causa del sollevamento. 1 = totale disaccordo, 7 = totale accordo.**

	1	2	3	4	5	6	7	Non Saprei
Comunità scientifica di Napoli	☐	☐	☐	☐	☐	☐	☐	☐
Comunità scientifica italiana	☐	☐	☐	☐	☐	☐	☐	☐
Comunità scientifica internazionale	☐	☐	☐	☐	☐	☐	☐	☐

74%

24. **A quale profondità approssimativa, secondo Lei, si potrebbe trovare del magma nella crosta dei Campi Flegrei al giorno d'oggi? Selezioni.**

- ☐ 1-2 km
- ☐ 2-3 km
- ☐ 3-4 km
- ☐ 4-5 km
- ☐ 5-6 km
- ☐ 6-7 km
- ☐ >7 km
- ☐ Non Saprei

25. **Quali cambiamenti dei parametri di monitoraggio vi aspettereste in caso del movimento di magma (intrusione) ai Campi Flegrei?**

80%

Sezione 3

26. **Valuti il livello di conoscenza del comportamento dei Campi Flegrei da parte dei seguenti gruppi. 1 = molto scarso, 7 = molto buono.**

	1	2	3	4	5	6	7	Non Saprei
Protezione Civile	☐	☐	☐	☐	☐	☐	☐	☐
Autorità Locali	☐	☐	☐	☐	☐	☐	☐	☐
Popolazione dei Campi Flegrei	☐	☐	☐	☐	☐	☐	☐	☐
I media	☐	☐	☐	☐	☐	☐	☐	☐

27. **A Suo avviso, qual è la causa dell'attuale sollevamento secondo i gruppi seguenti?**

	Magma Intrusione	Altro	Non Saprei
Protezione Civile	☐	☐	☐
Autorità Locali	☐	☐	☐
Popolazione dei Campi Flegrei	☐	☐	☐
I media	☐	☐	☐

28. **Secondo Lei, quali sono i pericoli e rischi più preoccupanti durante il sollevamento attuale secondo i gruppi seguenti?**

Protezione Civile	
Autorità Locali	
Popolazione dei Campi Flegrei	
I Media	

90%

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29. **Ci sono difficoltà nella comunicazione dell'attuale comportamento del vulcano tra gli scienziati al giorno d'oggi, la Protezione Civile, le autorità locali, i media e i civili? Se si, quali?**

- ◯ No
- ◯ Sì
- ◯ Non saprei

30. **Secondo Lei, nel caso di una futura emergenza vulcanica ai Campi Flegrei, quanto bene i seguenti gruppi capiranno le informazioni ricevute riguardanti il vulcano? 1 = molto male, 7 = molto bene**

	1	2	3	4	5	6	7	Non Saprei
Protezione Civile	☐	☐	☐	☐	☐	☐	☐	☐
Autorità Locali	☐	☐	☐	☐	☐	☐	☐	☐
Popolazione dei Campi Flegrei	☐	☐	☐	☐	☐	☐	☐	☐
I media	☐	☐	☐	☐	☐	☐	☐	☐

31. **Nel caso di future emergenze vulcaniche ai Campi Flegrei, quali potrebbero essere secondo Lei le difficoltà relative alla comunicazione della causa di unrest e dello stato attuale del vulcano tra gli scienziati, la Protezione Civile, le autorità locali, i media e i civili? Se si, quali?**

- ◯ No
- ◯ Sì
- ◯ Non saprei

100%

< Fine

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Chapter 10

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