Volcanism in the Campania Plain:
Vesuvius, Campi Flegrei and Ignimbrites
edited by B. De Vivo
© 2006 Elsevier B.V. All rights reserved.

Chapter 14

A hydrothermal model for ground movements (bradyseism) at Campi Flegrei, Italy

B. De Vivo and A. Lima*

Dipartimento di Scienze della Terra, Universita di Napoli Federico II, Via Mezzocannone 8 – 80134 Napoli, Italy

Abstract

Ground movements (bradyseism) at Campi Flegrei, Italy, have been explained by a classical model that involves the intrusion of new magma to shallow depth, or by models which emphasize both the magmatic and aquifer effects. The authors describe a model for the ground deformations that involves only hydrothermal fluids, of magmatic or meteoric/marine origin, with no direct involvement of the magma, other than as a heat source. They explain the bradyseism at Campi Flegrei by a hydrothermal model, using the porphyry systems (Henley and McNabb, 1978; Burnham, 1979; Fournier, 1999) as an analogue of the Campi Flegrei system. In this view, Campi Flegrei might very well represent a modern analogue of a mineralized porphyry system, as has been demonstrated for White Island, New Zealand (Rapien et al., 2003). The authors used fluid and melt inclusion data from Campi Flegrei and other volcanoes of the Neapolitan area (Vesuvius, Ponza and Ventotene) to demonstrate the linkage with porphyry systems. Fluid inclusions in all the above volcanic systems show clear evidence of various stages of silicate melt/hydrosaline melt/aqueous fluid/CO_2 immiscibility during the magmatic evolution and its transition from magmatic to hydrothermal stage, comparable to the plastic, lithostatic domain in porphyry systems. In contrast, convectively driven fluids are found only in the volcaniclastic sediments of the Campi Flegrei caldera (in the geothermal wells of San Vito and Mofete fields), and are representative of the brittle, hydrostatic domain. The coexistence of liquid-dominated and vapor-dominated inclusions in the same fluid inclusion assemblage is strong evidence of boiling conditions during inclusion trapping, whereas fluid inclusions with daughter crystals trapped in samples from deeper, hotter levels indicate a high concentration of solute (brines), as confirmed by drilling.

The scenario suggested by fluid inclusion data indicates that the Campi Flegrei system receives an influx of saline water (magmatic + seawater), localized in aquifers at depths of ~2.5–3 km. The fluids are heated by the underlying crystallizing magma and remain under lithostatic pressure for long periods. The pressure in the upper, apical part of the magma chamber increases as water exsolves from the magma and causes uplift of the overlying rocks (positive bradyseism). When the system ruptures, due to the increasing pressure, the regime changes from lithostatic to hydrostatic, resulting in boiling, hydraulic fracturing, volcanic tremors and finally pressure release leading to deflation of the ground. Afterward, the system begins to seal again due to the precipitation of newly formed minerals and a new phase of positive bradyseism will occur only after several years when the system "reloads" under new lithostatic pressure conditions. In this scenario, a hydrothermal eruption can still occur, but only if the fluids pass from lithostatic to hydrostatic pressure when the overlying rocks have a thickness <500 m. If this happens, the hydrothermal eruption could trigger a magmatic eruption, as was probably the case of the Monte Nuovo eruption in 1538 AD.

*Corresponding author. *E-mail address*: bdevivo@unina.it (B. De Vivo).

1. Introduction

The phenomenon of slow, vertical, ground movements in the Campi Flegrei area, Italy, has been known since Roman times. The ground movement is named bradyseism from the Greek words meaning, literally, "slow movement".

The most recent intense deformation in Campi Flegrei occurred in 1538, in 1970–1972 and in 1982–1984. The ground deformation of 1538 AD culminated with the eruption of Monte Nuovo, whereas no eruptions occurred in 1970–1972 and 1982–1984. Probably many other bradyseismic events occurred in the Campi Flegrei before 1538 leaving no historical reference to their occurrence. In the last 2000 years at Campi Flegrei, there has been an eruption only in 1538 AD.

Secular deformations and intense unrests are typical manifestations of activity at calderas, and a distinctive feature of such deformation episodes is that they are generally not followed by eruptions (Dzurisin and Newhall, 1984).

Two different mechanisms have been proposed for Campi Flegrei bradyseism. The first explains the phenomenon with the classical model of intrusion of new magma at shallow depth (Corrado et al., 1976; Barberi et al., 1984). The second one ascribes (Oliveri del Castillo and Quagliariello, 1969) the ground movements to the effect of heating and expansion of ground water (Casertano et al., 1976). In recent years, different researchers have presented models, emphasizing both the magmatic and aquifer effects, to explain the observed features (Bianchi et al., 1987; Gaeta et al., 1998; Bonafede, 1991; De Natale et al., 1991; Dvorak and Berrino, 1991; De Natale and Pingue, 1993; De Natale et al., 1997; Troise et al., 1997; Scandone et al., this volume). In particular, De Natale et al. (2001) developed a mechanical fluid-dynamic model to explain the main features of Campi Flegrei deformation. The model involves two processes: the first one is related to the elastic response of the shallow crust to increasing pressure within a shallow magma chamber; the second involves the fluid dynamics of shallow aquifers in response to increasing pressure and/or temperature at depth.

We extend the contribution of Oliveri del Castillo and Quagliariello (1969) by describing an active role for the deformations at Campi Flegrei to only hydrothermal fluids, both of magmatic and/or meteoric/seawater origin, and not to the magma. In our model, the magma plays only the role of a "furnace" to heat the system. To support our view we use data obtained from fluid inclusions from geothermal boreholes at Campi Flegrei (De Vivo et al., 1989), as well as data from fluid and melt inclusions in the nearby sub-volcanic systems of the Pontine Islands (De Vivo et al., 1995; Belkin et al., 1996) and Vesuvius (Lima et al., 2003, this volume).

Our model is based on the model of the porphyry systems (Henley and McNabb, 1978; Burnham, 1979; Fournier, 1999, and references therein) as an analogue of the Campi Flegrei sub-volcanic system. In other words, the Campi Flegrei represents a modern analogue of mineralized systems associated with former magmatic systems (Beane and Titley, 1981; Beane, 1982; Beane and Bodnar, 1995; Roedder and Bodnar, 1997). White Island (New Zealand) is an example of an active magmatic system that represents an embryonic copper porphyry system that has not reached the productive stage of copper mineralization (Rapien et al., 2003). The Burnham model is used extensively and accepted by many ore deposits experts. In this model, the active role played by hydrothermal fluids in the transport and deposition of mineralization is amply demonstrated. The occurrence of boiling fluids and hydrothermal explosions is well documented as well, where the magmas located at depths between 5 and 7 km act as a sub-volcanic heating engine of the system.

2. History of Campi Flegrei bradyseism and models proposed to explain the deformation

Initial studies concerning the slow movements at Campi Flegrei come from the observations of sea level markers on the archaeological site previously called Serapis Temple (but really a marketplace, Macellum) in Pozzuoli. Boreholes left by a marine mollusk (*Lithodomus lithophagus*) have been found on the columns of this monument. These bores (and the shells left in them) record ancient relative sea level changes. This interesting movement attracted the attention of many other researchers (Breislak, 1792; Forbes, 1829; Niccolini, 1839, 1845; Babbage, 1847; Lyell, 1872; Gunther, 1903), and Parascandola (1947), who reconstructed the history of secular ground movements at Campi Flegrei using the boreholes. Dvorak and Mastrolorenzo (1991) updated the studies of Parascandola by reconstructing the history of vertical movements at Campi Flegrei (Fig. 1). It has to be stressed that the above studies document the ground movements since Roman times but no data are available on the phenomenon before Roman times. We have a record of ground deformation at Campi Flegrei only for the past 2000 years, which in terms of geological phenomena represents a very short time span.

Historically, the first documented episode of fast uplift at Campi Flegrei is the one associated with the 1538 Monte Nuovo eruption (the ground rose about 7 m before the eruption). Bradyseism was again active in 1970–1972, when uplift of 1 m was registered; this episode was followed by subsidence of 30 cm. It has to be stressed that in the period 1970–1972, the uplift began to be recorded well after it had begun. The total uplift was probably on the order of 1.7 m and began in 1969. The uncertainty in our knowledge of the

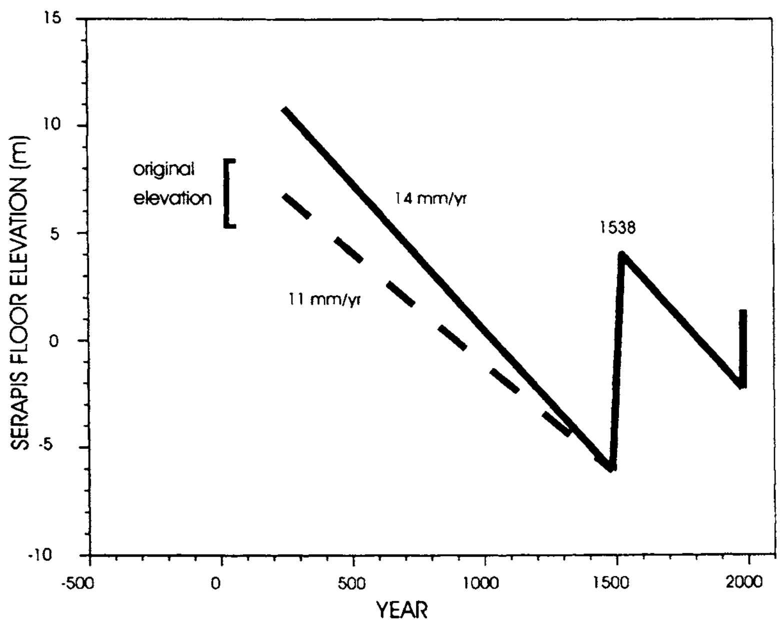

Figure 1. Vertical movements at Serapis temple, Pozzuoli (from Dvorak and Mastrolorenzo, 1991).

uplift that occurred only 33 years ago emphasizes that our knowledge about this phenomenon in pre-historic times is minimal.

A new episode of uplift of 1.8 m occurred from 1982 to 1984, then in 1985 was followed by a slow subsidence that continues today (Fig. 2). One of these episodes caused the evacuation of about 30,000 people from the city of Pozzuoli, and the building of a new town at Monte Ruscello, about 3 km away from the most active caldera center, but still inside the Campi Flegrei caldera! The areal distribution of ground movement at Campi Flegrei, with its circular symmetry, is shown in Figure 3.

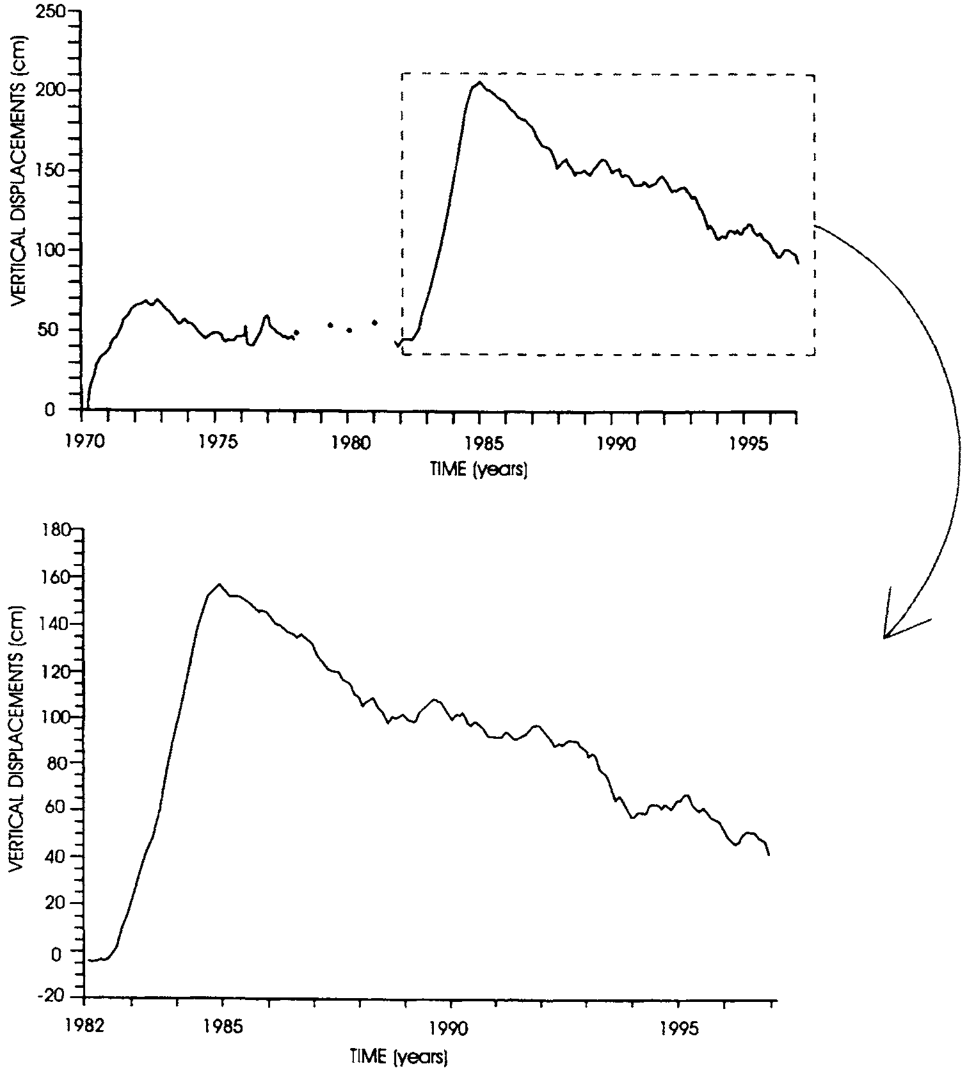

Figure 2. Vertical ground displacement at Pozzuoli harbour in the 1970–1996 period, as recorded by tide gauge (continuous line) and levelling data (closed circle) (from De Natale et al., 2001).

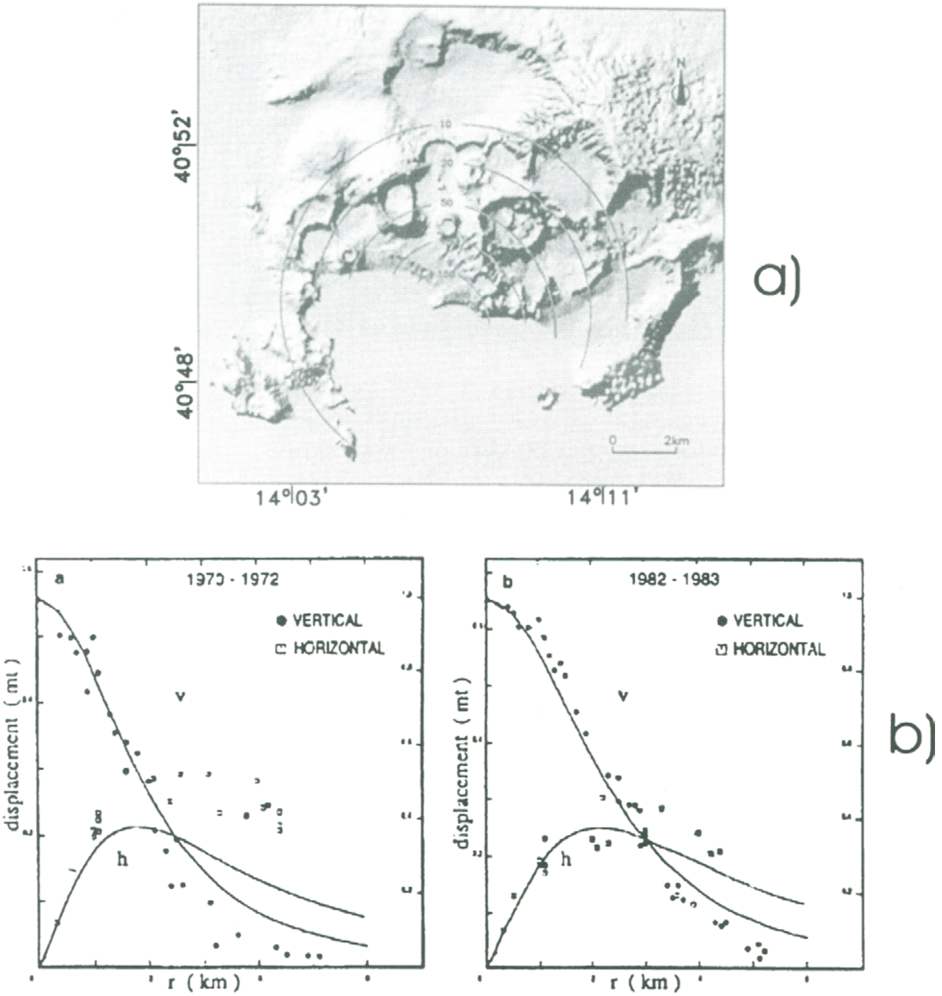

Figure 3. (a) Contours of vertical elevation at Campi Flegrei area in the period 1982–1985. (b) Ground deformation data as function of the distance from Pozzuoli, measured in the period 1970–1972 and 1980–1983; the fit with the best Mogi model is also shown (depth = 2.5 km for 1970–1972 and depth = 3 km for 1982–1983) (from De Natale et al., 2001).

Various models have been proposed to explain the deformation at Campi Flegrei. One purely mechanical model – mostly popular in the 1970s – attribute the unrest episodes to the intrusion of new magma at shallow depth (Corrado et al., 1976; Berrino et al., 1984; Bonafede et al., 1986; Bianchi et al., 1987). An alternative model, which has gained favor in recent years, explains the unrests mainly as a result of heating and expansion of aquifers (Oliveri del Castillo and Quagliariello, 1969; Casertano et al., 1976). Other researchers have published papers that explain the unrest mainly as a result of fluid-dynamic processes in the shallow geothermal system (Bonafede, 1990, 1991; De Natale, 1991; De Natale et al., 2001; Trasatti et al., 2005). Cortini et al. (1991) and Cortini and Barton (1993) suggested that the bradyseism is governed by the internal dynamics of the Campi Flegrei volcanic system whose physicochemical details are unknown. Analysis of the ground elevation at Pozzuoli, performed using nonlinear dynamics, showed that the Campi Flegrei system underwent a phase transition during the evolution from the sinking stage to that of uplift. They also suggested that the dynamics of Campi Flegrei, which appears fairly predictable on a timescale of a few days, could be driven by convection in the magma chamber. Cubellis et al. (2002) explain the different eruptive phases of Campi Flegrei in terms of chaotic convective cells operating inside the magma chamber, where the time evolution is hypothesized to be governed by three non-linear Lorenz equations (Lorenz, 1963). De Natale et al. (2001) make a comprehensive review of all pros and cons of the different models referenced above.

Seismicity at Campi Flegrei occurs only during unrest episodes (Corrado et al., 1976; De Natale et al., 1995). During 1982–1984 more than 15,000 earthquakes occurred, ranging from 0.4 to 4.2 in magnitude (De Natale and Zollo, 1986). De Natale et al. (1995) show that earthquake locations and mechanisms indicate the presence of faults that are associated with the inner caldera collapse structure and dip inward. Gravimetric and seismic methods (Berrino et al., 1984, 1992; AGIP, 1987; Berrino and Gasparini, 1995) show a marked Bouguer minimum centered at Campi Flegrei caldera (Fig. 4). Other seismic studies (Aster and Meyer, 1988; Ferrucci et al., 1992) define the 3D shallow velocity structure of the area and the location of the top part of the shallow magma chamber (Fig. 5). Ground deformations and seismicity are associated with the presence of intense fumarolic and hydrothermal activity, concentrated in the crater of Solfatara, where CO_2 and H_2O fluxes are particularly intense and probably represent magmatic degassing (Chiodini et al., 2001).

The above models for Campi Flegrei – and particularly the ones that propose a cause–effect relationship between ground uplift and the occurrence of an eruption – should be compatible with similar observations from other recently active calderas in the world (e.g., Rabaul, Mc Kee et al., 1989; Long Valley, Hill, 1984; Yellowstone, Dzurisin and Yamashita, 1987; see De Natale et al., 2001). However, studies of similar systems worldwide

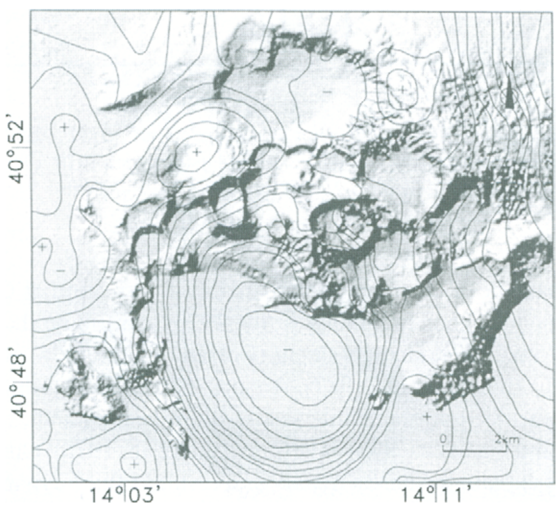

Figure 4. Bouguer anomaly contours at Campi Flegrei (from Scandone et al., 1991).

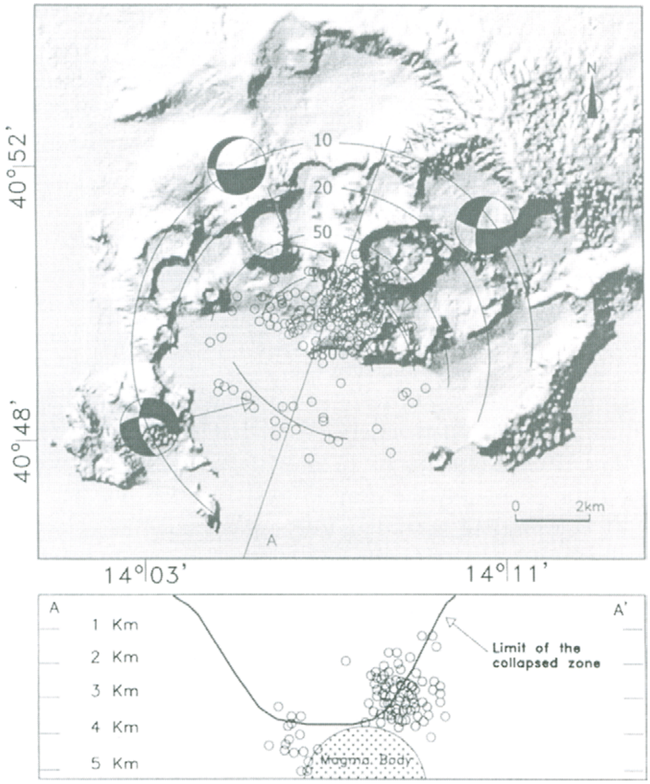

Figure 5. Map of various geophysical observations at Campi Flegrei. The contours of vertical elevation (in meters) are shown together with earthquake hypocenters; the projection of the collapsed zone as modelled from gravity anomalies is superimposed on the depth section of hypocenters. Composite focal mechanisms computed for three different seismic zones are also shown. Also shown is the location of magma chamber as inferred by Ferrucci et al. (1992) (from De Natale et al., 2001).

indicate that deformation episodes at calderas are generally not followed by eruptions (Dzurizin and Newhall, 1984). The eruption is an exception rather than the norm.

With these observations in mind, we describe the Campi Flegrei volcanic system and its vertical movements as the modern analogue of a mineralized porphyry system (Henley and McNabb, 1978; Burnham, 1979; Fournier, 1999; and references therein).

3. Magma and hydrothermal fluids

The role of hydrothermal fluids in the formation of Cu and Mo mineralization in magmatic systems (known as porphyry Cu and Mo) has been recognized since the 1960s (Burnham, 1967, 1979). In the following pages, we summarize briefly some relevant parts of this model from papers of Burnham (1979) and Fournier (1999).

According to the Burnham (1979) model, the first stage in the formation of a porphyry copper system involves the intrusion of water-undersaturated granodioritic magma.

Initially, the system is open and any volatiles generated during solidification of the melt escape from the system. Eventually, an impermeable rock rind develops at the top of the magma chamber, isolating the underlying magma from the overlying rocks. The magma becomes a closed system and only conductive loss of heat to the wall rocks is permitted. As the crystallization front migrates downward, the melt becomes saturated in water and an H_2O-saturated carapace composed of crystals+melt+fluid develops at the top of the magma chamber and below the crystalline rind (Fig. 6a).

The H_2O-saturated carapace migrates downward into the magma chamber as crystallization proceeds and plays a critical role in the development of porphyry Cu–Mo systems. The overlying crystalline rind serves as a barrier to the migration of volatiles, both outward to the wall rocks and inward from the wall rock. The H_2O-saturated carapace, especially in its upper parts, is the site of accumulation of an aqueous fluid phase as the system cools and evolves (Fig. 6b). The process of resurgent boiling (second boiling) is a natural consequence of cooling a melt saturated with respect to H_2O or one or more crystalline phases. The overall reaction, H_2O-saturated melt $\Rightarrow$ crystals + "vapor", takes place with the evolution of heat which is essentially the heat of crystallization (the heat of vaporization of H_2O from the melt is negligible, according to Burnham and Davis, 1974).

The exsolution of water from crystallizing hydrous melts produce high internal overpressures in the magma chamber. In isolated small pockets in crystallizing H_2O-saturated magma, values of ΔP_{in} (internal overpresure) $\geq$ 5 Kb are possible. Overpressures on the order of 400 bars are expected in igneous rocks deforming viscoelastically. An overpressure of this magnitude is enough to cause brittle failure of the overlying rocks (Burnham, 1972; Koide and Bhattacharji, 1975). The fractures are concentrated in and above the apical parts

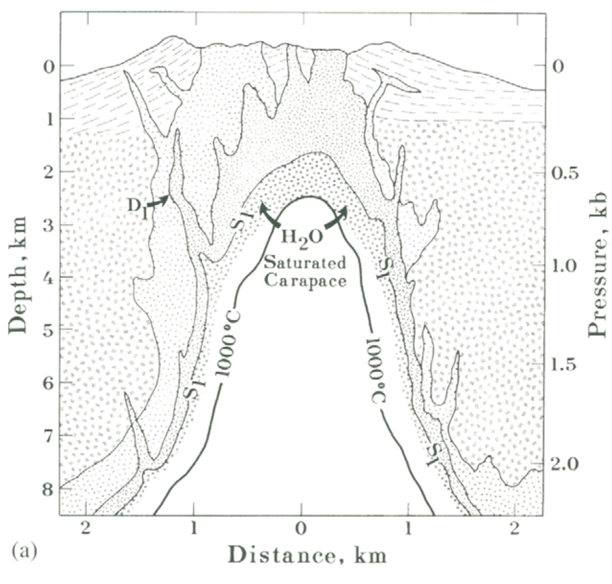

Figure 6a. Schematic cross section through a hypothetical granodiorite porphyry stock and associated dike (D_1). S_1 represents the H_2O-saturated solidus at this arbitrarily chosen initial stage in the development of a porphyry copper system and the circle pattern represents the zone of H_2O-saturated magma (H_2O-saturated carapace) (from Burnham, 1979).

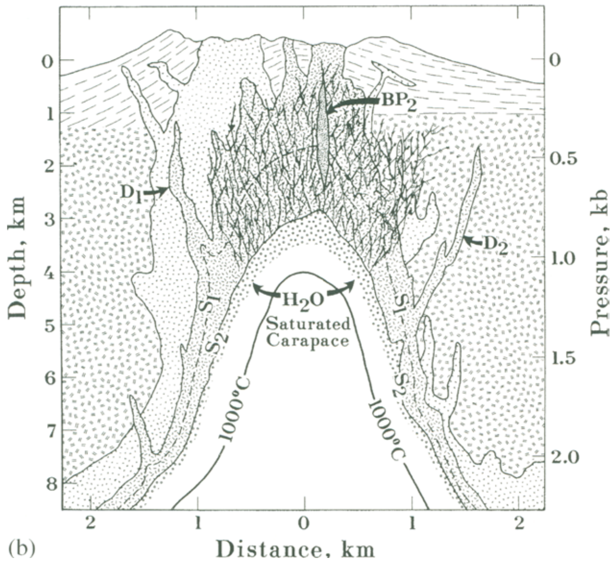

Figure 6b. Schematic cross section as in Figure 6a, except at a later (second) stage of solidification. BP_2 and D_2 schematically represent a breccia pipe and dike that formed as a result of wall rock failure between stages 1 and 2. Chaotic line pattern represents extensive fracture system that also developed during this period of activity and retreat of the H_2O-saturated carapace (from Burnham, 1979).

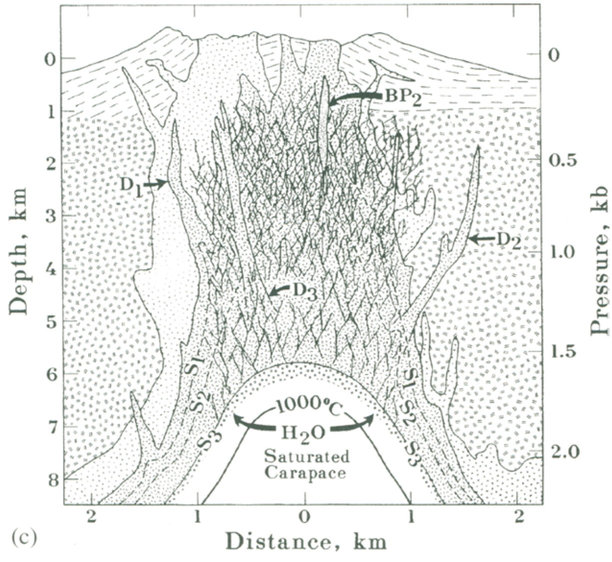

Figure 6c. Schematic cross section as in Figure 6a,b, except at a stage of waning magmatic activity in the development of a porphyry copper–molybdenum system (from Burnham, 1979).

of the stock (Fig. 6b), and tend to be steep, but their orientation depends upon the regional stress field.

An additional overpressure, referred to as "telluric pressure" that arises from differences in density between magma and wall rocks is superimposed on ΔP_{in} under quiescent conditions (Burnham, 1979). The "telluric pressure" is dependent on the height of the magma column above the depth of isostatic compensation, that is, the depth at which the wallrocks yield by plastic deformation under lithostatic pressure. It is this "telluric overpressure" that represents the driving force for intrusion as indicated in Figure 6a and which prevents a reduction in magma pressure that would have resulted from reduction in volume by crystallization.

The maximum mechanical energy ($P_t \Delta V_r$) released in the reaction H_2O-saturated melt $\Rightarrow$ crystals + "vapor" is enormous (Burnham, 1979), although it is only 1% of the total thermal energy content of the magma. At the depths indicated in Figures 6a–c, the mechanical energy released from the H_2O-saturated carapace presumably is expended mainly in fracturing a much larger volume of rocks. In the earlier stage of fracturing, the enclosing impermeable rocks are stretched laterally and may not be breached completely. Fluids penetrate this myriad of fractures and extend them outward and upward by hydraulic fracturing. This action results in lowering the fluid pressure in each fracture to below lithostatic pressure, except near the top of the fracture. With time, the H_2O-saturated carapace retreats to progressively deeper levels in the stock.

If major fractures breach the overlying rocks, breccia dikes and pipes are very likely to form (see BP_2 in Fig. 6b). If the breach occurs in the thinner, lateral flanks of the carapace, more normal dike intrusions result (D_2 in Fig. 6a,b). The mechanics of breccia pipe formation are hence visualized as primarily due to internal overpressure in the carapace (and not to contraction on cooling). As pressure decreases, the heat is lost to the wall rocks, and the magma dike devolatilizes and is quenched. In response to this process, more magma rises into the system until internal pressures are restored to near previous values.

At this stage, the magma system is restored to the same state as it was prior to the fracturing, except for the fact that the H_2O-saturated carapace is shifted downward in the magma chamber. In addition, the myriad of narrow fractures outside the solidus boundary promotes loss of H_2O and heat to the fracture system until the fractures become healed by precipitation of new minerals (mainly quartz). Further cooling of the magma leads to reactivation of the same process that operated before. The end result is a chimney-like fracture system (Fig. 6c) that serves to channel ore-bearing fluids and heat from the underlying magma system to higher levels in the stock. The above model of Burnham (1967, 1979) describes the mechanism that controls the deposition of ore (mostly Cu and Mo) in the classical "porphyry systems" (Lowell and Guilbert, 1970; Roedder, 1971; Wallace, 1974; Gustafson and Hunt, 1975; Henley and McNabb, 1978).

The transition from magmatic to epithermal conditions in a shallow sub-volcanic environment, such as in the case of Campi Flegrei, is shown in Figure 7. Dilute waters (dominantly meteoric) circulate through brittle rocks (Fig. 7a) at hydrostatic pressure at temperatures <370°C. The transition from brittle to plastic behavior occurs across the impermeable rocks between the underlying H_2O-saturated magma and the overlying highly fractured rocks at temperatures of 400–800°C. Within the lithostatic regime, hypersaline, magmatic brine and "steam" accumulate in relatively thin, horizontal, lenses or network of fractures that have limited vertical interconnectivity (when the least principal stress is the lithostatic load). When

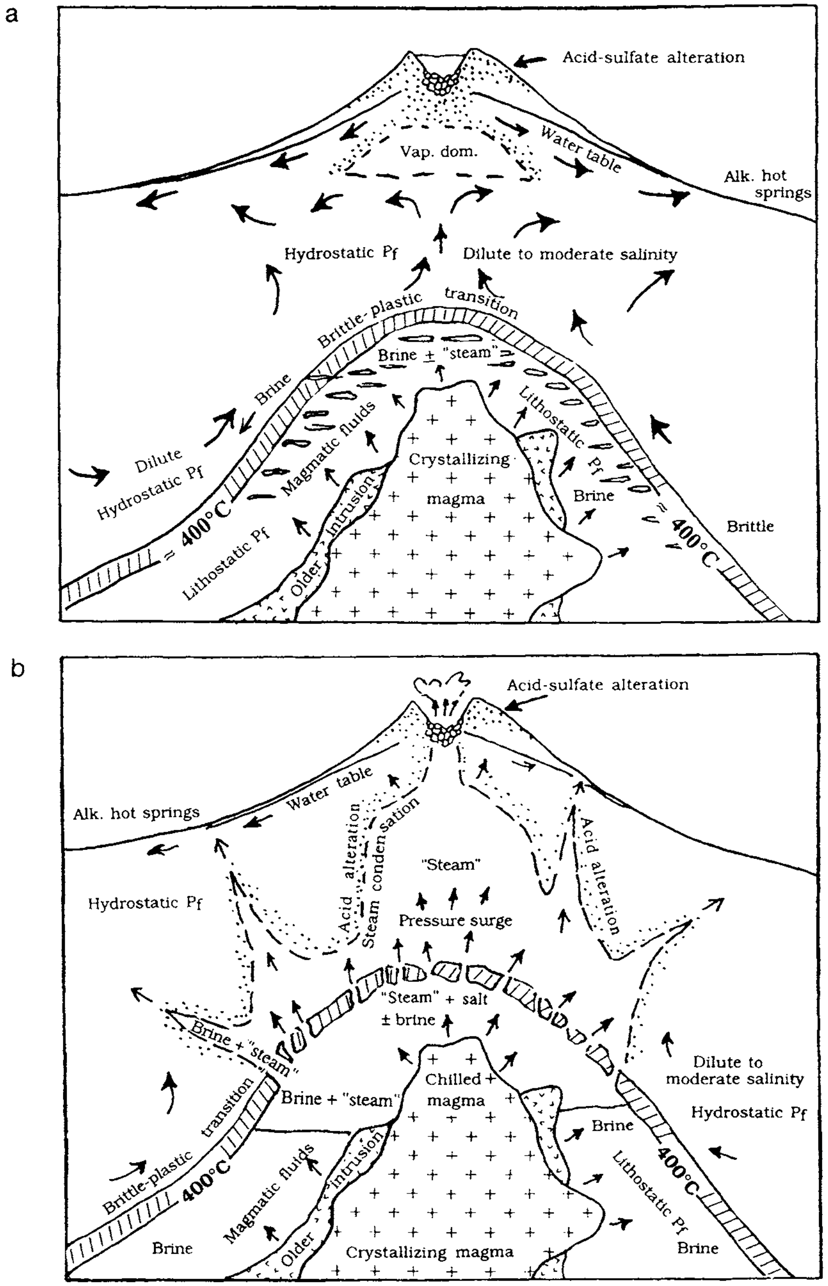

Figure 7. Schematic model of the transition from magmatic to epithermal conditions in a subvolcanic environment where the tops of intruded plutons are at depths in the range 1–3 km. (a) The brittle to plastic transition occurs at about 370–400°C and dilute, dominantly meteoric waters circulate at hydrostatic pressure in brittle rock, while highly saline, dominantly magmatic fluid at lithostatic pressure accumulates in plastic rocks. (b) Episodic and temporary breaching of a normally self-sealed zone allows magmatic fluid to escape into the overlying hydrothermal system (from Fournier, 1999).

magmatic bodies have been repeatedly intruded to relatively shallow depths, a large volume of 400–800°C plastic rocks develops, and the brittle–plastic interface can be as shallow as 2 km. In the SV1 well at Campi Flegrei, temperatures of 400°C have been found at ≈3000 m (De Vivo et al., 1989). Thus, sub-volcanic magmatic-hydrothermal systems are subdivided into a plastic, lithostatic domain, and an overlying hydrostatically pressured hydrothermal domain in which rocks deform brittlely. Several processes may play a role in triggering major breaches in the self-sealed, impermeable region that separates the plastic (magmatic) and brittle (hydrothermal) domains. One process is continued degassing of crystallizing magma and accumulation of the evolved fluid in the H_2O-saturated carapace beneath the impermeable barrier until the latter becomes stretched sufficiently to rupture by tensile failure (Northon and Cathles, 1973; Philipps, 1973; Burnham, 1979, 1985). A variation on this process is exsolution of volatiles at depth and transport to the top of the chamber by convection of the magma (Shinoara et al., 1995).

Another mechanism for breaching the self-sealed zone is upward injection of a new pulse of magma from depth. This would increase the strain rate within the overlying rocks to such a degree that the brittle to plastic transition temporarily migrates to a deeper and hotter enviroment. Fluids in the initially plastic rock would be at lithostatic pressure, and the change from plastic to brittle behavior with increasing strain rate would result in breaching of the self-sealed zone by shear failure in response to a small stress difference. A new pulse of magma would add volatiles upon crystallization and would heat previously existing brines trapped in horizontal lenses, inducing boiling. The resulting rapid expansion of the fluids would cause an additional increase in the strain rate, possibly leading to failure of the overlying rocks.

Large seismic events that affect permeability and rates of flow within an overlying hydrostatically pressured hydrothermal system may also lead to breaching of the self-sealed zone. In fact, the increase in flow rates may result in extraction of heat at the base of the circulating system more rapidly than heat can be supplied by conduction through plastic rocks from below. This leads to a progressive downward decrease in temperature and simultaneous expansion of the permeable region as a result of volumetric contraction and tensile cracking (Lister, 1974; Northon and Knapp, 1977; Northon and Knight, 1977; Carrigan, 1986).

Whichever mechanism produces the breaches in the self-sealed system, faulting, brecciation, hydrothermal alteration and ore deposition occur during this active process. This domain is characterized by hypersaline brines coexisting with "steam". For example, fluids present during the formation of "porphyry systems" are very hot (>400°C, and as high as 725°C), very saline (>30% to 60% salts), and consist of two phases, liquid and vapor (Roedder, 1984; Heinrich et al., 1992; Cline and Bodnar, 1994; Bodnar, 1995).

4. Fluid and melt inclusion data from Campi Flegrei and other Neapolitan sub-volcanic systems

The composition of residual liquids evolves during magmatic differentiation and these changes are recorded by fluid inclusions trapped in minerals (Roedder, 1984). Both magmatic and hydrothermal phenomena can thus be studied from observations of melt and fluid inclusions in igneous and hydrothermal minerals that crystallized at different depths and at various stages in the evolution of a magmatic system (De Vivo et al., 2005).

The observed composition of melt inclusions is determined by the composition of the original trapped melt and by the action of any later alteration processes, especially those due to hydrothermal activity. Hydrothermal changes frequently mask the original melt composition, particularly when the host igneous rock is introduced into the water table environment (Touret and Frezzotti, 1993; Student and Bodnar, 2004). Melt inclusion studies have thus concentrated mostly on the evolution of aqueous fluids in sub-volcanic systems in both ancient (porphyry ore deposits) and modern geothermal systems. Indeed, it was recognized from such studies that the majority of hydrothermal ore deposits in volcanic rocks or their subjacent plutonic suites were formed within geothermal systems similar to those active today (Henley and Ellis, 1983). At the same time, alteration processes have restricted the potential of fluid inclusion studies to follow the evolution of a melt during magma solidification from supraliquid temperatures (Roedder and Coombs, 1967; Frezzotti, 1992). Fortunately, however, melt evolution can be investigated unambiguously when magma immiscibility occurs in fluid inclusions (Roedder, 1992).

"Granitoid" xenoliths entrained in volcanic units from Campi Flegrei (and from sub-volcanic systems of other volcanoes in the Neapolitan area, including Vesuvius and the Pontine Islands (Ponza and Ventotene) provide samples of the deep environment. The source of such "granitoids" is the plastic, lithostatic, domain as discussed in the models of Burnham (1979) and Fournier (1999), where the fluids (melts + hydrosaline melts) trapped as inclusions represent magmatic fluids evolved from a subjacent crystallizing magma. In contrast, inclusions that trap vapor plus brines represent the plastic-brittle transition zone. In the Campi Flegrei system, information obtained from studies of fluid inclusions from geothermal boreholes (De Vivo et al., 1989) furnish data on the hydrostatic domain, which is characterized by dilute to moderate salinity fluids.

Geothermal exploration in the Campi Flegrei started in the years 1939 to 1954; this program was resumed in 1978 by a joint venture between the national utilities AGIP and ENEL and the Italian Geodynamic Project of CNR (Rosi and Sbrana, 1987) (Fig. 8). Several wells were drilled to depths up to 3 km (Carella and Guglielminetti, 1983). At shallow depths partially hydrothermally altered volcanic, volcaniclastic and sedimentary rocks are found. At greater depth, their thermometamorphic equivalents are encountered. The deep wells indicate the presence of a saline water-dominated geothermal field with multiple reservoirs. A fluid inclusion study (De Vivo et al., 1989) of cores from Mofete (MF1, MF2, MF5) and San Vito (SV1, SV3) geothermal wells indicates that the hydrothermal minerals were precipitated from aqueous fluids (containing $\pm$ CO_2) that were moderately saline (3–4 wt% NaCl equiv.) to hypersaline (up to 49 wt% NaCl equiv.), and were boiling. Three types of primary fluid inclusions were found in authigenic K-feldspar, quartz, calcite and epidote: (A) two-phase (liquid + vapor), liquid-rich inclusions with a range of salinity; (B) two-phase (liquid + vapor), vapor-rich inclusions with low salinity; and (C) three-phase (liquid + vapor + crystals – NaCl), liquid-rich inclusions with hypersaline fluids (Fig. 9).

Data from selected core samples reveal a general decrease in porosity and increase in bulk density with increasing depth and temperature. Hydrothermal minerals commonly fill fractures and pore spaces and define a zonation pattern, similar in all five wells studied, in response to increasing depth (pressure) and temperature. A greenschist facies assemblage, defined by albite + actinolite, gives way to amphibolite facies, defined by plagioclase (andesine) + horneblende, in the San Vito well at about 380°C. The fluid inclusion salinity values

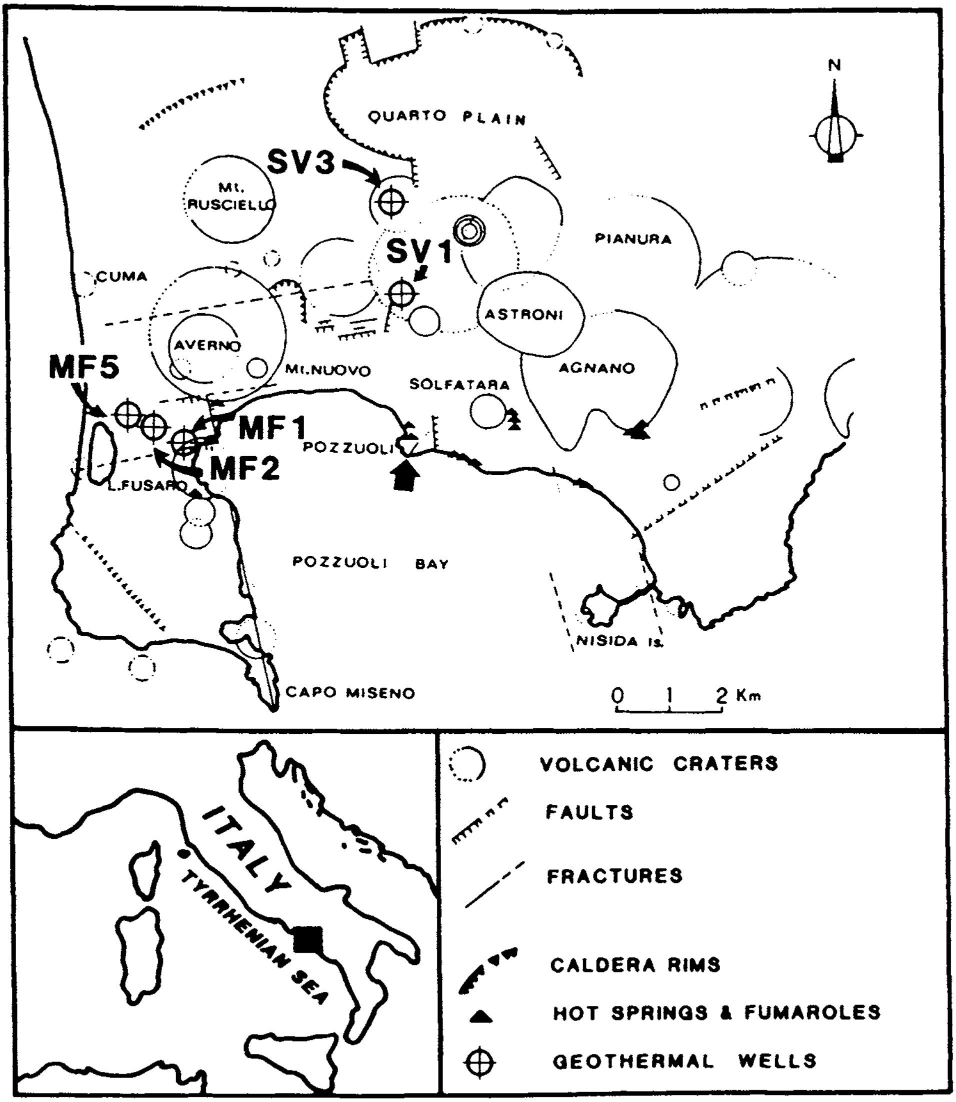

Figure 8. Volcano-tectonic schematic map of the Campi Flegrei volcanic complex showing the location of the studied geothermal wells. MF = Mofete, SV = San Vito. The arrow points to Pozzuoli, the center of uplift during the 1982–1984 bradyseismic crisis (from Barberi et al., 1984).

mimic the saline and hypersaline values found by drilling. Fluid inclusion (V/L) homogenization temperatures increase with depth and generally correspond to the extrapolated down-hole temperature. However, fluid inclusion data for MF5 and mineral assemblage for SV3, indicate a fossil, higher-temperature regime compared with the present one. The $^{87}Sr/^{86}Sr$ ratio in SV3 cores shows an approach to equilibrium with a fluid similar to modern seawater (De Vivo et al., 1989).

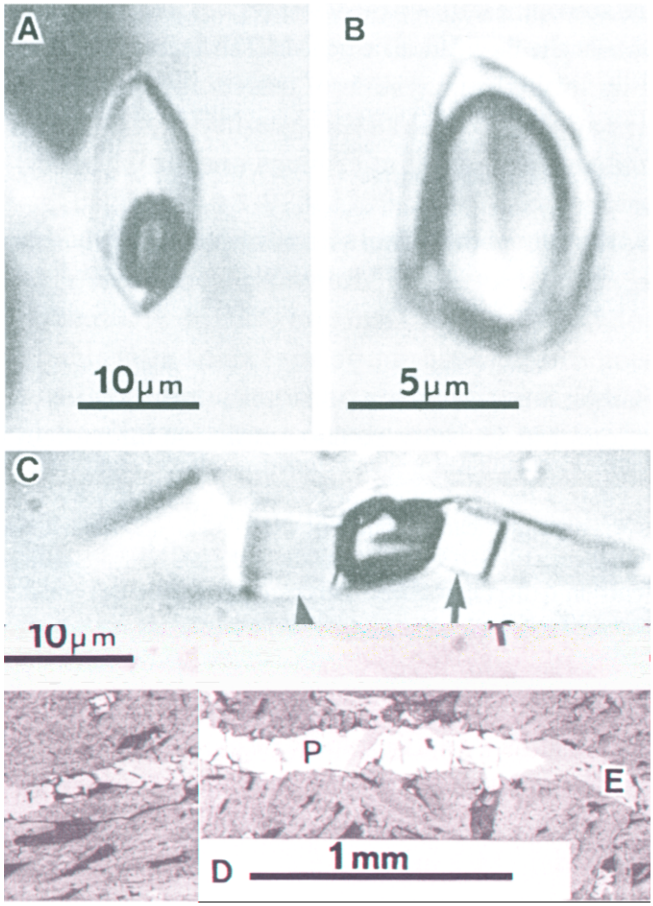

Figure 9. (A) Type A primary inclusion in calcite from San Vito 3-1948. Th = 290°C and 4 wt% NaCl equivalent. (B) Type B primary inclusion in quartz from San Vito 1-2676, Th = 321°C. (C) Type C inclusion from Mofete 5-2610 in quartz. Arrows point to halite cubes (daughter crystals). TmNaCl = 380°C, Th (V+L) = 390°C, salinity = ~44 wt% NaCl equivalent (assumes vapor present $NaCl+H_2O$ solution). (D) Typical filled fracture in Mofete 2-1824. The fracture oriented parallel to the scale bar is filled with pyrite (P) and epidote (E) (from De Vivo et al., 1989).

It has to be stressed that the common coexistence of type (A) and type (B) inclusions (with similar homogenization temperature) in samples from some of the studied cores suggests that the fluids were trapped while they were in a boiling condition and that a major component of the trapped fluids is CO_2. On the other hand, boiling conditions are also suggested by the association of calcite and adularia (Browne and Ellis, 1970; Simmons and Christensen, 1994) in the hydrothermal alteration mineralogy of different core samples.

Boiling, as explained by Henley and Mc Nabb, 1978, Burnham (1979) and Fournier (1999), is a common mechanism associated with ore deposition in the porphyry systems. A fluid in the lithostatic pressure domain (such as fluids in the plastic domain surrounding a crystallizing magma) may undergo boiling and effervescence during a pressure release caused by hydraulic/seismic fracturing. This process may permit the passage of fluids from the lithostatic to the hydrostatic domain. This process also results in the deposition of minerals in microcracks, fractures and breccias (Fig. 9) to re-seal the system and return the fluids to the lithostatic domain. The high salinity in the fluids trapped in newly formed minerals in the hydrostatic domain has been attributed to concentration processes affecting seawater-derived fluids. Fournier (1999) indicates that a brine is unlikely to form from a moderately saline solution (with seawater composition) by purely upward adiabatic decompression, especially when the heat source is relatively "shallow" and the fluids are heated to above 400–450°C, which are characteristics of the Campi Flegrei system.

The Campi Flegrei hydrothermal system represents a modern analogue of an ancient mineralized system and may have the potential for significant ore formation at depth, as has been suggested at White Island, New Zealand (Rapien et al., 2003). Although no major mineralization has yet been found at Campi Flegrei, galena, sphalerite, pyrrhotite, pyrite, arsenopyrite and hematite are found in minor amounts in fractures. In fact, the Campi Flegrei hydrothermal system has two characteristics, which would favor the formation of ore deposition. First, is the recognition of boiling. Boiling of a hydrothermal solution, especially one carrying significant amounts of CO_2, can be an effective way to deposit ore minerals (Cunningham, 1978, 1985). When the solution boils, the dissolved gases are strongly partitioned into the vapor phase. This can increase the pH and destabilize various metal complexes (sulfur species), and cause their precipitation. Second, brine stratification is present in the form of reservoirs with different salinities (Guglielminetti, 1986) (Table 1), and fluid mixing across the brine interface is a significant mechanism for ore mineral precipitation (Williams and McKibben, 1987; McKibben et al., 1988). Evidence that fluid inclusions in the Campi Flegrei record a transition from magmatic to hydrothermal conditions is found by Fedele et al. (this volume), who conducted fluid inclusion studies and SEM-EDS and electron microprobe analysis on late-stage minerals such as apatite, Zr-bearing minerals (zircon and baddaleyite), pyroclore group minerals, thorite and phosphate. Complex daughter mineral assemblages found in multiphase fluid inclusion of the xenoliths are evidence of high solute entrapment. Abundances of chlorides, sulfides and, to a lesser extent, sulfates and carbonates, suggest that the fluid inclusions trapped a hypersaline/sulfur-rich fluid (possibly with minor CO_2) which likely exsolved from a crystallizing magma. Microthermometry on secondary hypersaline fluid inclusions suggests two possible scenarios for fluid trapping: (1) circulation of non-boiling, high-temperature (up to 525°C), high salinity fluids which were trapped under decreasing P–T conditions; (2) circulation of boiling, hypersaline fluids, trapped at low pressure and temperatures up to 300°C. All data suggest that the xenoliths record the effects of a hydrothermal phase, possibly associated with a transition from a magma-dominated to a fluid-dominated stage at the margins of a magma chamber. The presence of base metals and tungsten minerals in the xenoliths suggests a potential for mineralization which is similar to that observed in the alkaline volcanic systems of Pontine Islands and Mt. Somma-Vesuvius, where fluids escaping from the upper part of a sub-volcanic magma chamber and trapped in the plastic, lithostatic domain have also been found in feldspathoid-bearing syenite xenoliths.

At Ventotene, De Vivo et al. (1995) found gabbroic cumulate and alkali syenite xenoliths. The gabbroic cumulates contain only silicate melt inclusions ± vapor bubble ± droplets of an opaque phase and rarely some CO_2. The alkali syenite xenoliths have three types of fluid inclusions: (1) single-phase vapor and silicate melt inclusions;

Table 1. Well chemistry data (in ppm) from Campi Flegrei (from De Vivo et al., 1989).

Mofete field, chemistry of water separated at atmospheric pressure (Guglielminetti, 1996)

	Shallow reservoir		Intermediate reservoir	Deep reservoir
	Mofete 1 550–896 m	Mofete 1 1273–1606 m	Mofete 2 1275–1989 m	Mofete 5 2310–2699 m
Na	14320	20860	10600	85160
K	1760	1880	2467	43380
Ca	792	2124	1005	53950
B	178	183	295	231
Sr	49	58	30	1310
As	13	17	22	nd
Li	36	46	28	480
Mn	10	28	52	5510
Fe	1	3	1	9450
SiO_2	568	690	938	210
Cl	25304	37800	21169	313850
HCO_3	116	77	85	TR
SO_4	72	7	12	TR
TDS	42860	65509	37880	515902
pH	7.5	6.5	6.0	4.5

Mofete field, water chemistry calculated at reservoir conditions (Guglielminetti and Tore, 1985; Guglielminetti, 1986)				San Vito 1, composition of fluid sampled during purge tests (Bruni et al., 1985)		
	Shallow reservoir		Intermediate reservoir	Sample	A	B
	Mofete 1 550–896 m	Mofete 1 1273–1606 m	Mofete 2 1275–1989 m			
Na	10025	12589	5090	Na	11750	6280
K	1230	2342	1180	K	8000	4025
Ca	555	1281	480	Ca	3290	1980
B	125	110	140	Mg	1120	540
Sr	34	41	14	Li	47	26
As	9	11	11	F	5	5
Li	25	28	13	SiO2	369	246
Mn	7	17	25	Cl	37755	20024
Fe	1	2	1	HCO3	nd	24
Mg	nd	5	0.61	TDS	62336	33150
Ba	nd	2.8	0.49	pH	3.2	4.4

(Continued)

Table 1. (Continued)

	Mofete field, water chemistry calculated at reservoir conditions (Guglielminetti and Tore, 1985; Guglielminetti, 1986)			San Vito 1, composition of fluid sampled during purge tests (Bruni et al., 1985)		
	Shallow reservoir		Intermediate reservoir	Sample	A	B
	Mofete 1 550–896 m	Mofete 1 1273–1606 m	Mofete 2 1275–1989 m			
Cr	nd	0.07	0.14			
Cu	nd	0.01	<0.03			
Pb	nd	0.36	<0.29			
Zn	nd	0.12	<0.02			
SiO_2	398	417	450			
Cl	17710	22810	10200			
HCO_3	81	46	41			
SO4	50	4	6			
TDS	30000	39500	18200			
CO2*	nd	17642	31980			
H_2S*	nd	236	1117			

*flashable gas, maximum content.
nd, not determined; TDS, total dissolved solids; TR, trace.

(2) two-phase silicate melt + salt, silicate melt + CO_2 (V), aqueous (L + V), and silicate melt + vapor inclusions; (3) three-phase and multiphase inclusions: CO_2 (L) + CO_2 (V) + H_2O; silicate melt + saline melt + H_2O ± birefringent or opaque-trapped minerals; H_2O + salt + silicate glass ± birefringent trapped minerals. The characteristics of the fluid phases trapped in xenoliths indicate that they evolved in a pressure regime varying from lithostatic to hydrostatic during the crystallization and cooling history of the host intrusive rocks. The gabbro crystallized at $T \sim$ 900–1000°C, around 1 to 1.4 kb, whereas the alkali syenite crystallized at $T \sim$ 600–700°C, between ~200 and 400 bars; the transition rock crystallized between 800°C and 1000°C (Fig. 10). The Ventotene xenoliths show clear evidence of various stages of silicate melt – hydrous saline melt – aqueous fluid – CO_2 fluid immiscibility (Roedder, 1992, and references therein) during the magmatic evolution and its transition from magmatic to hydrothermal stage. Fluids from the hydrostatic, brittle domain are recorded in the alkali syenite only by very subordinate secondary H_2O inclusions, which were trapped in the later stage of the hydrothermal process at much lower temperature (130–290°C), but at pressures relatively close to those of alkali syenite crystallization. The trapping of dilute to moderate salinity fluids in the hydrostatic domain occurred after breaching of the impermeable rocks separating the plastic from the brittle zones (see Fig. 7). At Ponza Island, Belkin et al. (1996) found evidence similar to that found in Ventotene in feldspathoid-bearing syenite xenoliths entrained in trachyte. Fluid inclusions show evidence of heterogeneous immiscible phases during crystallization, represented by silicate melt + CO_2 + H_2O and

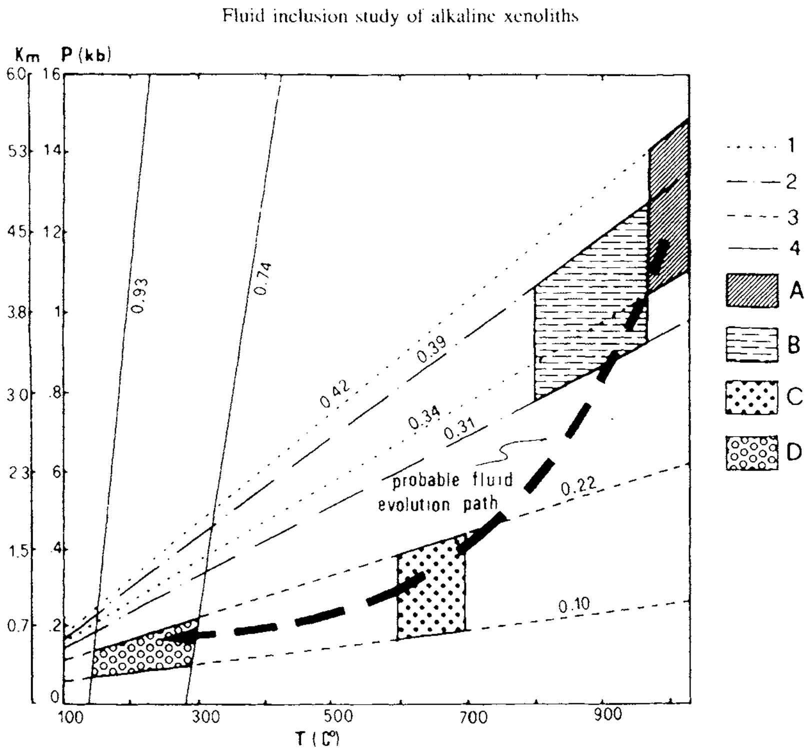

Figure 10. P–T plot for fluid inclusions in Ventotene xenoliths. Isochores for CO_2 and H_2O–NaCl are from Brown and Lamb (1989). (1) CO_2 isochores from VT-2A (gabbro); (2) CO_2 isochores from VT-4 (transition rock); (3) CO_2 isochores from VT-5B (alkali syenite); (4) H_2O–NaCl isochores from secondary inclusions (VT-4; VT-5). Field A, B and C represent the P–T conditions of crystallization of gabbro (A), transition rocks (B) and alkali syenite (C); field D represents the P–T conditions of late-stage aqueous secondary inclusion trapping (from De Vivo et al., 1995).

silicate melt/hypersaline/sulfur-rich aqueous inclusions. In contrast to Ventotene, however, no evidence for immiscibility between silicate melt and hydrosaline melt was found in Ponza. The potassium feldspar contains primary H_2O + CO_2 inclusions with small amounts of silicate melt and hypersaline/sulfur-rich aqueous inclusions. In the latter, small amounts of silicate melt are sometimes present. Silicate melt inclusions are more commonly observed coexisting with hypersaline/sulfur-rich aqueous inclusions along healed fractures. The evolution of the fluid phases under sub-solidus conditions is represented by secondary aqueous + CO_2 inclusions. These inclusions indicate moderate- to high-temperature hydrothermal fluids (between 359°C and 424°C) with low-to-moderate salinity (between 2.9 and 8.5 wt% NaCl equiv.). As for Ventotene, the xenoliths record the magmatic/hydrothermal transition, most probably in the upper part of the magma chamber. The peripheral zone, also in Ponza, has become enriched in various incompatible elements, including Th, U, REE, Zr, Y and others, as reported for other potassium-enriched magmatic systems (Tait, 1988; Turbeville, 1992; Belkin et al., 1994; Federico et al., 1994; Renzulli et al., 1995).

At Mt. Somma-Vesuvius, a mineralogical, fluid inclusion and isotope study (Fulignati et al., 2001; Gilg et al., 2001) carried out on skarn-bearing xenoliths found evidence of immiscibility in good agreement with the findings from the Pontine Islands (De Vivo et al., 1995; Belkin et al., 1996; Kamenetsky et al., 2003). In particular four major types of fluid inclusions were trapped in various minerals (wollastonite, vesuvianite, gehlenite, clinopyroxene and calcite: (a) primary silicate melt inclusions; (b) $CO_2 \pm H_2S$-rich vapor inclusions; (c) multiphase aqueous brine inclusions, with sylvite and halite daughter minerals; (d) complex chloride–carbonate–sulfate–fluoride–silicate-bearing saline melt inclusions, indicating immiscibility between silicate melt–aqueous chloride-rich liquid-carbonate/sulfate melt. The study also indicates that there is no evidence – as for Ponza and Ventotene – for a convectively cooling hydrothermal system at the magma-carbonate wall rock interface, indicating a lack of participation of externally derived fluids, such as meteoric waters or formational fluids. Salt-rich aqueous fluids must have separated from a silicate melt (magma) during cooling, infiltrated the carbonate rocks, and reacted with them to form skarns.

The saline melt inclusions contain high concentrations of Cl, carbonate, sulfate and sometimes F. These saline fluids might be related to the extremely Cl- and SO_2-rich melts of the Plinian to sub-Plinian eruptions (Webster et al., 2001; Lima et al., this volume). As found also in Ponza and Campi Flegrei, the fluids played an important role in the transport of Ti, Zr, Th, U and REE from the magma to the wall rocks.

The Pb, Nd, C and O isotope compositions of skarn and the presence of silicate melt inclusion-bearing wollastonite nodules indicate assimilation of carbonate wall rock by the alkaline magma at moderate depth (<5 km) which promoted exsolution of CO_2-rich vapor and complex saline melts from a locally contaminated magma.

5. A model of the bradyseism at Campi Flegrei

Fluid and melt inclusion studies indicate that the Campi Flegrei system is subdivided into a shallower permeable zone, where convectively driven fluids (encountered in the San Vito and Mofete wells) circulate, and a deep plastic, lithostatic domain dominated by magmatic fluids. The occurrence of two distinct fluid regimes is thought to be related to the presence of a large caldera filled by volcaniclastic sediments. The fluids have mostly either meteoric or seawater origin, with a magmatic component. This difference, i.e., the lack of a caldera filled with sediments above the volcanic plumbing systems of the other Neapolitan volcanic centers, might explain the occurrence of the bradyseism in the Campi Flegrei, as opposed to the other nearby volcanic areas where such ground movement has never been observed.

The drilling of MF1, MF2 and MF5 penetrated three reservoirs (Carella and Guglielminetti, 1983). A shallow reservoir was found in MF1 with a temperature of 250°C and a calculated reservoir fluid composition between 30,000 and 40,000 ppm TDS (total dissolved solids). A more dilute, intermediate reservoir was observed in MF2 with a temperature of 337°C and calculated fluid composition of ~18,000 ppm TDS. Well MF5 briefly produced hypersaline fluids from 2700 m (>500,000 ppm TDS, corresponding to 150,000 ppm in the reservoir) with a bottom hole temperature of 347°C (Table 1). The measured pCO_2 in MF1 at 1606 m (295°C) was 26 bars and in MF2 at 1989 m (345°C) was 51 bars (AGIP internal report), and is in agreement with the Carella and Guglielminetti (1983) estimate of the CO_2 content of the Mofete field of between 1 and 5 mol%. The data from San Vito field indicate fluid compositions that resemble compositions reported from the shallow and intermediate reservoirs of the Mofete field.

The coexistence of liquid-dominated and vapor-dominated inclusions indicates that the system, at some stage, contained a two-phase fluid. This coexistence would reflect a causal relationship between boiling and mineral deposition with the resulting inclusion trapping. The inclusion salinities (Figs. 11 and 12) reflect the chemical characteristics of the fluids found by drilling (Table 1). The variation of salinity with depth and location of MF1, MF2 and MF5, agrees with the fluid systematics found by Carella and Guglielminetti (1983), who stress the existence of multiple reservoirs, separated by impermeable layers. The shallow reservoir found at MF2 produced moderately saline fluids whereas the deep reservoir of MF5 contains a hypersaline fluid. Zones of low transmissivity prevent a major inflow of groundwater from penetrating and diluting the deep reservoirs. Hydrothermal activity in the Campi Flegrei is related to the onset of volcanic activity in the area (~50,000 ybp).

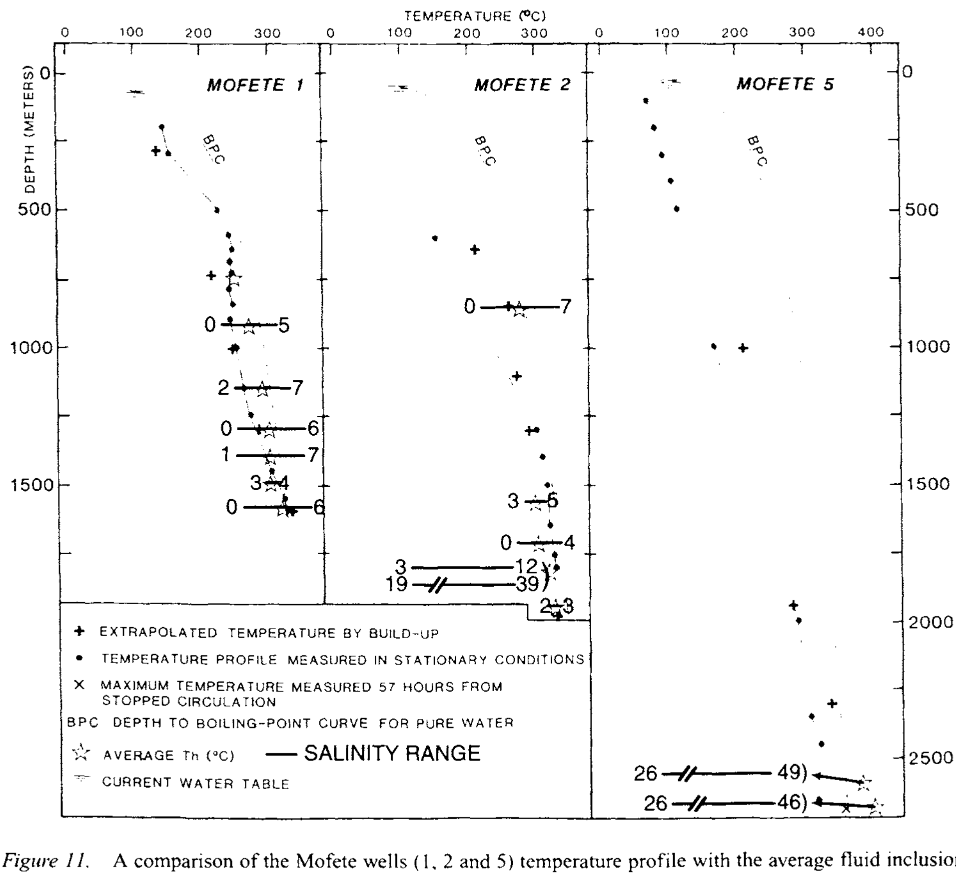

Figure 11. A comparison of the Mofete wells (1, 2 and 5) temperature profile with the average fluid inclusion homogenization temperature and range of salinity (wt% NaCl equivalent). The depth to the boiling-point curve for pure water is shown calculated to the current water table (from De Vivo et al., 1989).

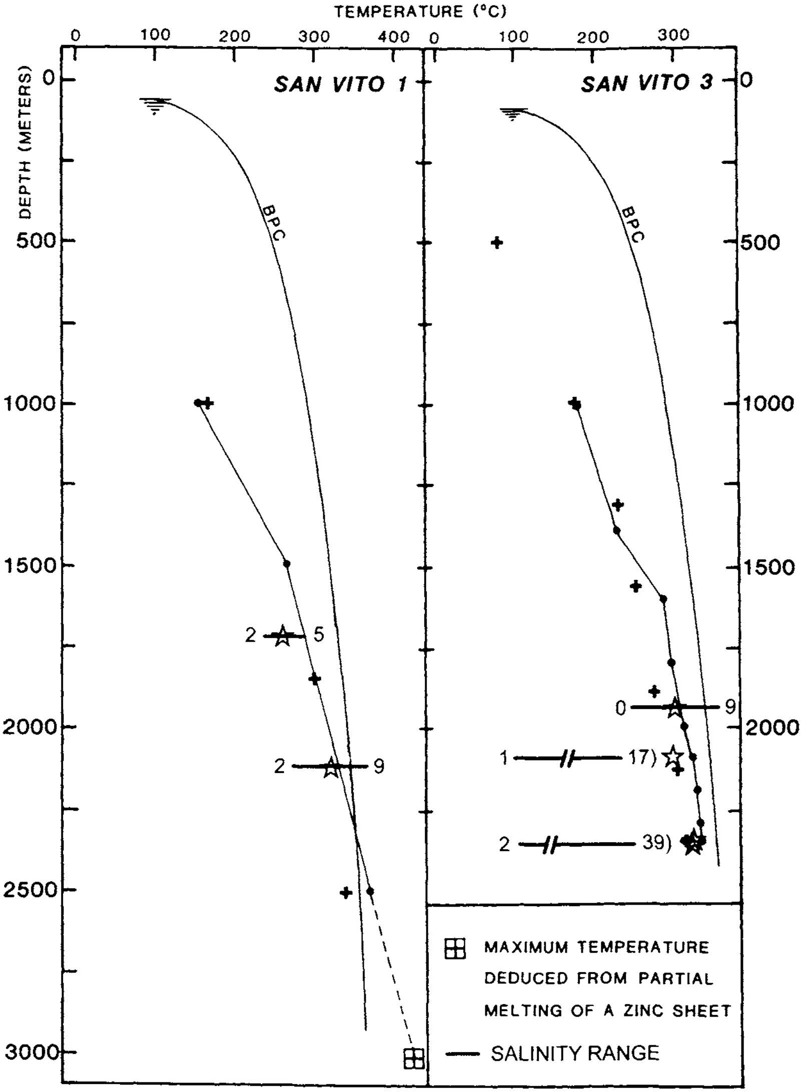

Figure 12. A comparison of the San Vito wells (1, 2 and 5) temperature profile with the average fluid inclusion homogenization temperature and range of salinity (wt% NaCl equivalent). The depth to the boiling-point curve for pure water is shown calculated to the current water table (from De Vivo et al., 1989).

This is confirmed by geochronology of zircons from alkali syenite nodules, which gives an age of ~50,000 ybp (Fedele et al., this volume). On the other hand, the homogenization temperatures of fluid inclusions found in hydrothermal minerals from the wells (except for MF5) correlate with the current thermal regime and the general characteristics of the fields (Figs. 11 and 12). This suggests that the host minerals are relatively recent and/or that the fields have remained stable during the later stages of Campi Flegrei evolution. Drilling data suggests that deep MF5 fluids have cooled subsequent to inclusions trapping.

In summary, primary fluid inclusions trapped in various authigenic minerals yield homogenization temperatures and salinities that mimic the temperature increase with depth and the salinities of the sampled reservoirs. The coeval nature of liquid-rich and vapor-rich inclusions is strong evidence of boiling conditions during inclusion trapping. Daughter crystals (halite) in core samples from deeper, hotter, levels indicate a high concentration of solute, as confirmed by drilling. This means that the above fluids are trapped in the brittle zone (see Fig. 7), in a hydrostatic pressure domain, immediately above the brittle-plastic transition zone. Conversely, the fluid inclusions that record immiscibility (in alkali syenite xenoliths) are trapped in the lithostatic domain and contain mostly fluids of magmatic origin.

The overall scenario of fluid inclusions and isotope characteristics is compatible with a Campi Flegrei geothermal field with a deep influx, from magmatic + seawater sources, of saline water which remains localized in aquifers at a depth of ~2.5–3 km. These aquifers are mostly under lithostatic pressure, being heated by an underlying, crystallizing magma body that is not moving upwards. The magma, in other words, does not increase the pressure on the overlying load of volcanoclastic sediments, but it furnishes to the overlying system heat and fluids, which escape from the apical portion of the magma chamber as shown in the Burnham model (1979).

The brittle–ductile transition zone is not fixed in space, but migrates up as the system is recharged with magma and migrates downward as the system cools. This is predicted in the Burnham model (Burnham et al., 1979) and documented in porphyry-mineralized systems (Beane and Titley, 1981; Beane, 1982; Beane and Bodnar, 1995; Roedder and Bodnar, 1997). The vertical and lateral extension of the hydrothermal system is relatively limited because of the presence of the overlying impermeable rocks.

In our model of Campi Flegrei, in addition to the inner, ~5–7 km deep, impermeable layer separating the plastic, lithostatic region from the brittle (hydrostatic) domain, there is a second, outer, shallower, impermeable layer. This is at a depth of ~2–2.5 km in the permeable volcaniclastic rocks, and its impermeability is due to the precipitation of newly formed minerals (as a consequence of the boiling process), which heal all fractures and faults that extend toward the surface. The continuous supply of heat and inflow of magmatic fluids creates an overpressure primarily against the inner, deeper, impermeable layer (transition from plastic to brittle domain), and subsequently also against the outer impermeable layer separating the more dilute surficial from the deep (at 2700 m) hypersaline geothermal aquifers (Carella and Guglielminetti, 1983). These internal overpressures, concentrated in the apical part of a crystallizing magma, can reach extremely elevated values ($\Delta P_{in} \geq 5$ kb). The amount of this overpressure, caused only by H_2O/CO_2-saturated fluids, would be enough to explain the positive bradyseism, which occurs episodically at Campi Flegrei.

When ΔP_{in} is high enough to breach the deeper impermeable layer and also the more surficial impermeable layer within the brittle domain, then hydrofracturing occurs, followed by intense boiling and consequent hydrothermal circulation. This is the moment when volcanic tremors occur and when the positive bradyseism is at its peak, as shown by geophysical experiments (Leet and Malone, 1990; Leet, 1991), which demonstrate that intense boiling in sub-volcanic environments can be responsible for volcanic tremor. In other words, in order to explain the seismic tremors, it is not necessary to invoke the presence of a magma, which rises toward the surface. At this point, the fluids that were in a lithostatic pressure regime pass into the hydrostatic domain, and the intense boiling causes the deposition of newly formed hydrothermal minerals; the mineral deposition, in turn, self-seals the system and leads to the restoration of the lithostatic domain.

Immediately after breaching of the impermeable layer, which separates the plastic from the brittle zone, deflation begins and lasts until the lithostatic domain is restored by the deposition of newly formed minerals (self-sealing process). In order to start a new episode of positive bradyseism it is necessary that the fluids escaping from the underlying crystallizing magma and/or fluids of external source (meteoric influx) "reload" the system, generating an overpressure in the apical portion of the magma body/chamber.

Considering that the hypocenters of the earthquakes registered during the last bradyseismic event of 1982–1984 were located at ~4 km depth (which should correspond to the location of the zone above the impermeable layer separating the plastic from the hydrostatic domains), we assume that a magma chamber, from which the magmatic fluids and the heat derive, should be located at a depth of ~5–7 km.

Our model, which is supported by the ideas of Cortini et al. (1991) and Cortini and Barton (1993), explains both the inflation and the deflation of the ground surface at Campi Flegrei, whereas a model which suggests that positive bradyseism is caused by the intrusion and rise of new magma toward the surface would explain only the positive event (inflation), but not the negative event (deflation). The crystallizing magma acts only as the "furnace" of the system, heating up and, at the same time, supplying fluids to underground, confined hydrothermal systems. Because of the depths at which these processes occur, we consider an eruption a very improbable event, because the ΔP_{in} is not enough to breach completely 5–7 km of overlying load. The case would be very different if a new batch of magma were intruded and rise toward the surface in Campi Flegrei. But, so far, no clear geophysical evidence of such occurrence has been reported. In our model an eruption could occur, as was the case of Monte Nuovo in 1538 AD. This would happen only if the process described above occurred with a load of volcanic rock/sediment <500 m. In this case, the high-pressured fluids passing from lithostatic to hydrostatic pressure domain could give rise to a hydrothermal eruption which in turn could trigger a magmatic eruption.

6. Conclusions

To explain the bradyseism at Campi Flegrei, we propose a model that emphasizes the role of hydrothermal fluids as opposed to the classical model of intrusion of new magma at shallow depths (Barberi et al., 1984; Corrado et al., 1976), suggesting that ground deformations could be generated by conductive heating of the hydrothermal fluids overlying the magmatic chamber. The authors describe a hydrothermal model based on the model of porphyry systems (Henley and McNabb, 1978; Burnham, 1979; Fournier, 1999). Campi Flegrei might represent a modern analogue of an ancient mineralized porphyry system (Beane and Titley, 1981; Beane and Bodnar, 1995; Roedder and Bodnar, 1997; Sasada, 2000), as has been suggested for White Island, New Zealand (Rapien et al., 2003). Deep, aqueous fluids, heated by an underlying crystallizing magma, would remain under lithostatic pressure for a long period. The heat and fluids supplied by the magma would determine the overpressure in the upper, apical, part of the magma chamber through accumulation of volatile phases confined by impermeable rocks, which cause uplift of the overlying rocks (positive bradyseism). A crisis occurs when the conditions change from lithostatic to hydrostatic pressure, with consequent boiling (De Vivo et al., 1989), hydraulic fracturing, volcanic tremor and then pressure release. At this point, the area experiences a sudden change from maximum inflation, followed by pressure release to beginning of subsidence (deflation of the ground). Afterward, the system, saturated with boiling fluids, begins to seal again due to the precipitation of newly deposited minerals. The beginning of a new positive bradyseism phase will occur only after several years when the system "reloads" under new lithostatic pressure conditions. This model is capable of explaining both inflation and deflation of ground, whereas a model which is restricted to the magma rising process as responsible for the bradyseism, would explain only the inflation. In the above scenario, a hydrothermal eruption can still occur, but only if the fluids pass from lithostatic to hydrostatic pressure when the overlying rocks have a thickness <500 m. If this happens, a hydrothermal eruption could trigger a magmatic eruption, which was probably the case of the Monte Nuovo eruption in 1538 AD.

Acknowledgments

The authors are grateful to the thorough reviews of Massimo Cortini (University of Napoli Federico II, Italy), Charles G. Cunningham (U. S. Geological Survey, Reston, Va, USA), Giuseppe De Natale (INGV, Napoli, Italy) and Robert J. Bodnar (Virginia Tech, Blacksburg, VA, USA), which have helped to greatly improve the final manuscript.

References

AGIP, 1987. Geologia e geofisica del sistema geotermico dei Campi Flegrea. Int. Report, p. 17.

Aster, R.C., Meyer, R.P., 1988. Three-dimensional velocity structure and hypocenter distribution in the Campi Flegrei caldera, Italy. Tectonophysics 149, 195–218.

Babbage, C., 1847. Observation on the Temple of Serapis at Pozzuoli near Naples. R. and J.E. Taylor, London, p. 35.

Barberi, F., Corrado, G., Innocenti, F., Luongo, G., 1984. Phlegraean Fields 1982 – 1984: brief chronicle of a volcano emergency in a densely populated area. Bull. Volcanol. 47(2), 175–185.

Beane, R.E., 1982. Hydrothermal alteration in silicate rocks. In: Titley, S.R. (Ed.), Advances in geology of the porphyry copper deposits of southwestern North America. University of Arizona Press, Tucson, pp. 117–137.

Beane, R.E., Bodnar, R.J., 1995. Hydrothermal fluids and hydrothermal alteration in porphyry copper deposits. Arizona Geol. Soc. Digest 20, 83–93.

Beane, R.E., Titley, S.R., 1981. Porphyry copper deposits. Part II. Hydrothermal alteration and mineralization. Econ. Geol. 75th Anniversary Vol. pp. 235–263.

Belkin, H.E., Cavarretta, G., Tecce, F., 1994. Petrogenesis of leucite-bearing nodules from Colli Albani volcano ejecta, Latium, Italy. Abstracts Int. Mineral. Assoc., 16th Gener. Meeting, 4–9 September, Pisa, Italy, p. 35.

Belkin, H.E., De Vivo, B., Lima A., Torok, K., 1996. Magmatic (silicate/sulfur-rich/CO_2) immiscibility and zirconium and rare-earth element enrichment from alkaline magma chamber margins: evidence from Ponza island, Pontine Archipelago. Eur. J. Mineral. 8, 1401–1420.

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

Berrino, G., Gasparini P., 1995. Ground deformation and caldera unrest. Cahiers du Centre Europeen de Geodynamique et de Sismologie 8, 41–55.

Berrino, G., Rymer, H., Brown, G.C., Corrado, G., 1992. Gravity-height correlation for unrest at calderas. J. Volcanol. Geotherm. Res. 53, 11–26.

Bianchi, R., Cordini, A., Federico, C., Giberti, G., Lanciano, P., Pozzi, J.P., Sartoris, G., Scandone R., 1987. Modelling of surface ground deformation in volcanic areas: the 1970–1972 and 1982–1984 crises of Campi Flegrei, Italy. J. Geophys. Res. 92 (B13), 14139–14150.

Bodnar, R.J., 1995. Fluid inclusion evidence for a magmatic source for metals in porphyry copper deposits. Miner. Assoc. Canada Short Course 23, 139–152.

Bonafede, M., 1990. Axi-symmetric deformation of a thermo-poro-elastic half-space: inflation of a magma chamber. Geophys. Int. 103(2), 289–299.

Bonafede, M., 1991. Hot fluid migration: an efficient source of ground deformation. Application to the 1982–85 crisis at Campi Flegrei, Italy. In: Luongo, G., Scandone, R. (Eds), Campi Flegrei. J. Volcanol. Geotherm. Res. 48, 187–198.

Bonafede, M., Dragoni, M., Quareni, F., 1986. Displacement and stress fields produced by a centre of dilatation and by a pressure source in visco-elastic half-space: application to the study of ground deformation and seismic activity at Campi Flegrei, Italy. Geophys. J. R. Astr. Soc. 87, 455–485.

Breislak, S., 1792. Essai mineralogiques sur le Solfatare de Pouzzole. Part 3, Observations sur l'exterieur du crater de la Solfatare. Giaccio, Naples, pp. 170–177.

Brown, P.E., Lamb, W.M., 1989. P-V-T properties of fluids in the system $H_2O \pm CO_2 \pm NaCl$: new graphical presentations and implications for fluid inclusion studies. Geochim. Cosmochim. Acta 53, 1209–1221.

Browne, P.R.L., Ellis, A.J., 1970. The Oaki-Broadlands hydrothermal area, New Zealand: mineralogy and related geochemistry. Am. J. Sci. 269, 97–131.

Bruni, P., Chelini, W., Sbrana, A., Verdiani, G., 1985. Deep exploration of the S. Vito area (Pozzuoli-NA) – Well S. Vito 1. In: Strub, A.S., Ungemach, P. (Eds), European Geothermal Update. Proc. Third Int. Seminar, pp. 390–406.

Burnham, C.W., 1967. Hydrothermal fluids at the magmatic stage. In: Barnes, H.L. (Ed.), Geochemistry of hydrothermal ore deposits. Holt, Rinehart and Winston, New York, pp. 34–76.

Burnham, C.W., 1972. The energy of explosive volcanic eruptions. Earth Mineral. Sci., The Pennsylvania State University, 41, 69–70.

Burnham, C.W., 1979. Magmas and hydrothermal fluids. In: Barnes, H.L. (Ed.), Geochemistry of hydrothermal ore deposits. Wiley, New York, pp. 71–136.

Burnham, C.W., 1985. Energy release in subvolcanic environments: implications for breccia formation. Econ. Geol. 80, 1515–1522.

Burnham, C.W., Davis, N.F., 1974. The role of H_2O in silicate melts: II. Thermodynamics and phase relations in the system $NaAlSi_3O_8$-H_2O to 10 kilobars, 700 to 1000°C. Am. J. Sci. 274, 902–940.

Carella, R., Guglielminetti, M., 1983. Multiple reservoirs in the Mofete fields, Naples, Italy. Workshop on Geothermal Reservoir Engineering, Stanford, 12 pp.

Carrigan, C.R., 1986. A two-phase hydrothermal cooling model for shallow intrusions. J. Volcanol. Geotherm. Res. 28, 175–192.

Casertano, L., Oliveri del Castillo, A., Quagliariello, M.T., 1976. Hydrodynamics and geodynamics in the Phlegraean Fields area of Italy. Nature 264, 154–161.

Chiodini, G., Marini L., Russo, M., 2001. Geochemical evidence for the existence of high-temperature hydrothermal brines at Vesuvius volcano, Italy. Geochim. Cosmochim. Acta 65(13), 2129–2147.

Cline J.S., Bodnar, R.J., 1994. Direct evolution of brine from a crystallizing silicic melt at the Questa, New Mexico, molybdenum deposit. Econ. Geol. 89, 1780–1802.

Corrado, G., Guerra, I., Lo Bascio, A., Luongo, G., Rampoldi, F., 1976. Inflation and microearthquake activity of Phlegraean Fields, Italy. Bull. Volcanol. 40(3), 169–188.

Cortini, M., Barton, C.C., 1993. Non linear forecasting analysis of inflation-deflation patterns of an active caldera (Campi Flegrei, Italy). Geology 21, 239–242.

Cortini, M., Cilento, L., Rullo, A., 1991. Vertical ground movements in the Campi Flegrei caldera as a chaotic dynamic phenomenon. J. Volcanol. Geotherm. Res. 48, 103–114.

Cubellis, E., Di Donna, G., Luongo, G., Mazzarella, A., 2002. Simulatine the mechanism of magmatic processes in the Campi Flegrei area (southern Italy) by the Lorenz equations. J. Volcanol. Geotherm. Res. 115, 339–349.

Cunningham, C.G., 1978. Pressure gradients and boiling as mechanisms for localizing ore in porphyry systems. J. Res. Geol. Survey 6, 745–754.

Cunningham, C.G., 1985. Characteristics of boiling-water-table and carbon dioxide models for epithermal gold deposition. In: Tooker, E.W. (Ed.), Geologic characteristics of sediment- and volcanic-hosted disseminated gold deposits – Search for an occurrence model. U. S. Geol. Survey Bull. 1646, 43–46.

De Natale, G., Petrazzuoli, S.M., Pingue, F., 1997. The effect of collapse structures on ground deformations in calderas. Geophys. Res. Lett. 24(13), 1555–1558.

De Natale, G., Pingue, F., 1993. Ground deformations in collapsed caldera structures. J. Volcanol. Geotherm. Res. 57, 19–38.

De Natale, G., Pingue, F., Allard, P., Zollo, A., 1991. Geophysical and geochemical modeling of the Campi Flegrei caldera. In: Luongo, G., Scandone, R. (Eds), Campi Flegrei. J. Volcanol. Geotherm. Res. 48, 199–222.

De Natale, G., Troise, C., Pingue, F., 2001. A mechanical fluid-dynamical model for ground movements at Campi Flegrei caldera. J. Geodyn. 32, 487–517.

De Natale, G., Zollo, A., 1986. Statistical analysis and clustering features of the Phlegraean Fields earthquake sequence (May 1984). Bull. Seism. Soc. Amer. 76(3), 801–814.

De Natale, G., Zollo, A., Ferraro, A., Virieux, J., 1995. Accurate fault mechanism determinations for a 1984 earthquake swarm at Campi Flegrei caldera (Italy) during an unrest episode: implications for volcanological research. J. Geophys. Res. 100 (B12), 24167–24185.

De Vivo, B., Barbieri, M., Belkin, H.E., Chelini, W., Lattanzi, P., Lima, A., Tolomeo, L., 1989. The Campi Flegrei (Italy) geothermal system: a fluid inclusion study of the Mofete and San Vito fields. J. Volcanol. Geotherm. Res. 36, 303–326.

De Vivo, B., Lima, A., Webster, J.D., 2005. Volatiles in magmatic-volcanic systems. Elements 1, 19–24.

De Vivo, B., Torok, K., Ayuso, R.A., Lima, A., Lirer, L., 1995. Fluid inclusion evidence for magmatic/saline/CO_2 immiscibility and geochemistry of alkaline xenoliths from Ventotene island, Italy. Geochim. Cosmochim. Acta 59(14), 2941–2953.

Dvorak, J.J., Berrino, G., 1991. Recent ground movement and seismicity activity in Campi Flegrei, Southern Italy: episodic growth of a resurgent dome. J. Geophys. Res. 96(B2), 2309–2323.

Dvorak, J.J., Mastrolorenzo, G., 1991. The mechanism of recent vertical crustal movements in Campi Flegrei caldera, Southern Italy. Geol. Soc. Amer. Spec. Pap., 263.

Dzurisin, D., Newhall, C.G., 1984. Recent ground deformation and seismicity at Long Valley (California), Yellowstone (Wyoming), the Phlegrean Fields (Italy) and Rabaul (Papua, new Guinea). In: Hill, D.P., Bailey, R.A., Ryall, A.S. (Eds), Proceedings of Workshop XIX: Active tectonic and magmatic processes beneath Long Valley Caldera, eastern California. Open-File Report – U.S. Geological Survey, pp. 784–829.

Dzurisin, D., Yamashita, K.M., 1987. Vertical surface displacements at Yellowstone caldera, Wyoming, 1976–1986. J. Geophys. Res. 92(B13), 13753–13766.

Fedele, L., Tarzia, M., Belkin, H.E., De Vivo, B., Lima, A., Lowenstern, J.B. (this volume). Magmatic-hydrothermal fluid interaction and mineralization in alkali-syenite nodules from the Breccia Museo pyroclastics deposit, Naples, Italy.

Federico, M., Peccerillo, A., Barbieri, M., Wu, T.W., 1994. Mineralogical and geochemical study of granular xenoliths from the Alban Hills volcano, Central Italy: bearing on evolutionary processes in potassic magma chambers. Contrib. Mineral. Petrol. 115, 384–401.

Ferrucci, F., Hirn, A., Virieux, J., De Natale, G., Mirabile, L., 1992. P-SV conversions at a shallow boundary beneath Campi Flegrei caldera (Naples, Italy): evidence for the magma chamber. J. Geophys. Res. 97(B11), 15351–15359.

Forbes, J.D., 1829. Physical notice in the Bay of Naples. Number 5. On the temple of Juppiter Serapis at Pozzuoli and the phenomena which it exhibits. Edin. J. Sci., New Series 1, 260–286.

Fournier, R.O., 1999. Hydrothermal processes related to movement of fluid from plastic into brittle rock in the magmatic-epithermal environment. Econ. Geol. 94(8), 1193–1211.

Frezzotti, M.L., 1992. Magmatic immiscibility and fluid phase evolution in the Mount Genis granite (southeastern Sardinia, Italy). Geochim. Cosmochim. Acta 56, 21–33.

Fulignati, P., Kamenetsky, V.S., Marianelli, P., Sbrana, A., Mernagh, T.P., 2001. Melt inclusion record of immiscibility between silicate, hydrosaline, and carbonate melts: applications to skarn genesis at Mount Vesuvius. Geology 29, 1043–1046.

Gaeta, F.S., De Natale, G., Peluso, F., Mastrolorenzo, G., Castagnolo, D., Troise, C., Pingue, F., Mita, D.G., Rossano, S., 1998. Genesis and evolution of unrest episodes at Campi Flegrei caldera: The role of thermal fluid dynamical processes in the geothermal system. J. Geophys. Res. 103 (B9), 921–933.

Gilg, H.A., Lima, A., Somma, R., Belkin, H.E., De Vivo, B., Ayuso, R.A., 2001. Isotope geochemistry and fluid inclusion study of skarns from Vesuvius. Mineral. Petrol. 73, 145–176.

Guglielminetti, M., 1986. Mofete geothermal field. Geothermics 15, 781–790.

Guglielminetti, M., Tore, G., 1985. Pit tests in the wells Mofete 1 and 2. In: Strub, A.S., Ungemach, P. (Eds), European Geothermal Update. Proc. Third Int. Seminar, pp. 473–483.

Gunther, R.T., 1903. The submerged Greek and Roman foreshore near Naples. Archaeologia 58, 62.

Gustafson, L.B., Hunt, J.P., 1975. The porphyry copper deposit at El Salvador, Chile. Econ. Geol. 70, 857–912.

Heinrich, C.A., Ryan, C.G., Mernagh, T.P., Eadington, P.J., 1992. Segregation of ore metals between magmatic brine and vapor – a fluid inclusion study using PIXE microanalysis. Econ. Geol. 87, 1566–1583.
Henley, R.W., Ellis, A.J., 1983. Geothermal systems ancient and modern: a geothermal review. Earth Sci. Rev. 19, 1–50.
Henley, R.W., McNabb, A., 1978. Magmatic vapor plumes and ground water interaction in porphyry copper emplacement. Econ. Geol. 73(1), 1–20.
Hill, D.P., 1984. Monitoring unrest in a large silicic caldera. The Long Valley-Inyo craters volcanic complex in the East-central California. Bull. Volcanol. 47(2), 371–395.
Kamenetsky, V.S., De Vivo, B., Naumov, V.B., Kamenestky, M.B., Mernagh, T.P., Van Achterbergh, E., Ryan, C.G., Davidson, P., 2003. Magmatic inclusions in the search for natural silicate-melt immiscibility: methodology and examples. In: De Vivo, B., Bodnar, R.J. (Eds), Melt inclusions in volcanic systems. Methods, Applications and Problems. Series "Developments in Volcanology 5", Elsevier, Amsterdam, pp. 65–82.
Koide, H., Bhattacharji, S., 1975. Formation of fractures around magmatic intrusions and their role in ore localization. Econ. Geol. 70, 781–799.
Leet, R.C., 1991. Investigation of hydrothermal boiling and steam quenching as possible sources of volcanic tremor and geothermal ground noise. PhD thesis. University of Washington, Seattle, WA, p. 282.
Leet, R.C., Malone, S.D., 1990. Boiling in hydrothermal systems may cause volcanic tremor. AGU 1990 Fall Meeting. EOS, Transactions 71(43), 1675.
Lima, A., Danyushevsky, L.V., De Vivo, B., Fedele, L., 2003. A model for the evolution of the Mt. Somma-Vesuvius magmatic system based on fluid and melt inclusion investigations. In: De Vivo, B., Bodnar, R.J. (Eds), Melt inclusions in volcanic systems. Methods, Applications and Problems. Series "Developments in Volcanology 5", Elsevier, Amsterdam, pp. 227–249.
Lima, A., De Vivo, B., Sintoni, M.F., An hydrothermal model to explain geochemical variations between 79 A.D. and 1944 A.D. Vesuvius eruptions.
Lister, C.R.B., 1974. On the penetration of water into hot rock. Geophys. J. Astron. Soc. 39, 465–509.
Lyell, C., 1872. Principles of geology. London, pp. 164–179.
Lorenz, E.N., 1963. Deterministic non-periodic flow. J. Atmos. Sci. 20, 130–141.
Lowell, J.D., Guilbert, J.M., 1970. Lateral and vertical alteration-mineralization zoning in porphyry ore deposits. Econ. Geol. 65, 373–408.
McKee, C., Mori, J., Talai, B., 1989. Microgravity changes and ground deformation at Rabaul caldera, 1973–1985. In: Latter, J.H. (Ed.), IAVCEI Proc. in Volcanol. 1, Volcanic Hazards, Springer, Wien, pp. 389–428.
McKibben, M.A., Williams, A.E., Okubo, S., 1988. Metamorphosed Plio-Pleistocene evaporates and origins of hypersaline brines in the Salton Sea geothermal system, California: fluid inclusion evidence. Geochim. Cosmochim. Acta 52, 1047–1056.
Niccolini, A., 1839. Tavola cronologica-metrica delle varie altezze tracciate dal mare fra la costa di Amalfi ed il promontorio di Gaeta nel corso di diaciannove secoli. Napoli, Flautina, pp. 11–52.
Niccolini, A., 1845. Descrizione della gran terna puteolana volgarmente detta Tempio di Serapide. Napoli, Stamperia Reale.
Northon, D.L., Cathles, L.M., 1973. Breccia pipe products of exsolved vapour from magmas. Econ. Geol. 68, 540–546.
Northon, D.L., Knapp, R., 1977. Transport phenomena in hydrothermal systems: nature of porosity. Am. J. Sci. 277, 913–936.
Northon, D.L., Knight, J.E., 1977. Transport phenomena in hydrothermal systems: cooling plutons. Am. J. Sci. 277, 937–981.
Oliveri del Castillo, A., Quagliariello, M.T., 1969. Sulla genesi del bradisismo flegreo. Atti Associazione Geofisica Italian, 18th Congress, Napoli, pp. 557–594.
Parascandola, A., 1947. I fenomeni bradisismici del Serapeo di Pozzuoli, Napoli, Privately published, pp. 156.
Philipps, W.L., 1973. Mechanical effects of retrograde boiling and its probable importance in the formation of some prophyry ore deposits. Inst. of Mining and Metallurgy Trans., Sect. B 82, B90–B98.
Rapien, M.H., Bodnar, R.J., Simmons, S.F., Szabo, C.S., Wood, C.P., Sutton, S.R., 2003. Melt inclusion study of the embryonic porphyry copper system at White Island, New Zealand. Econ. Geol., Spec. Publ. 10, 41–59.
Renzulli, A., Upton, B.G.J., Nappi, G., 1995. Magma chamber processes preceding the Pitigliano Formation eruption (Latera volcanic complex, central Italy): evidence from cognate plutonic clasts. Acta Vulcanologica 7(1), 55–74.

Roedder, E., 1971. Fluid inclusion studies on the porphyry-type ore deposits at Bingham, Utah, Butte, Montana, and Climax, Colorado. Econ. Geol. 66, 98–120.
Roedder, E., 1984. Fluid inclusions. Reviews in Mineralogy, Mineral. Soc. Am., Vol. 12, pp. 644.
Roedder, E., 1992. Fluid inclusion evidence for immiscibility in magmatic differentiation. Geochim. Cosmochim. Acta 56, 5–20.
Roedder, E., Bodnar, R.J., 1997. Fluid inclusion studies of hydrothermal ore deposits. In: Barnes, H.L. (Ed.), Geochemistry of hydrothermal ore deposits, 3rd ed. Wiley, New York, pp. 657–698.
Roedder, E., Coombs, D.S., 1967. Immiscibility in granitic melts, indicated by fluid inclusions in ejected granitic blocks of Ascension island. J. Petrol. 8, 417–451.
Rosi, M., Sbrana, A. (Eds). 1987. Phlegraean Fields. CNR, Quaderni de "La Ricerca Scientifica", Monog. 114(9), 175.
Sasada, M., 2000. Igneous-related active geothermal system versus porphyry copper hydrothermal system. Proceeding World Geothermal Congress 2000, Kyushu – Tohoku, Japan, May 28 – June 10, pp. 1691–1693.
Scandone, R., Bellucci, F., Lirer, L., Rolandi, G., 1991. The structure of the Campanian Plain and the activity of the Neapolitan volcanoes (Italy). J. Volcanol. Geotherm. Res. 48, 1–31.
Scandone, R., Giacomelli, L., Speranza, F.F. The volcanological history of the volcanoes of Naples: a review (this volume).
Shinoara, H., Kazahaya, K., Lowenstern, J.B., 1995. Volatile transport in a convecting magma column: implications for porphyry Mo mineralization. Geology 23, 1091–1094.
Simmons, S.F., Christensen, B.W., 1994. Origins of calcite in a boiling geothermal system. Amer. J. Sci. 294(3), 361–400.
Student, J.J., Bodnar, R.J., 2004. Silicate melt inclusions in porphyry copper deposits: Identification and homogenization behavior. Can. Mineral. 42(5), 1583–1599.
Tait, S.R., 1988. Samples from the crystallizing boundary layer of a zoned magma chamber. Contrib. Mineral. Petrol. 100, 470–483.
Touret, J.L.R., Frezzotti, M.L., 1993. Magmatic remnants in granites. Bull. Soc. Geol. Franc., II 229–242.
Trasatti, E., Giunchi, G., Bonafede, M., 2005. Structural and rheological constraints on source depth and overpressure estimate at the Campi Flegrea caldera, Italy. J. Volcanol. Geotherm. Res. 144, 105–118.
Troise, C., De Natale, G., Pingue, F., Zollo, A., 1997. A model for earthquake generation during unrest crises at Campi Flegrei and Rabaul calderas. Geophys. Res. Lett. 24(13), 1575–1578.
Turbeville, B.N., 1992. Relationships between chamber margin accumulates and pore liquids: evidence from arrested in situ processes in ejecta, Latera caldera, Italy. Contrib. Mineral. Petrol. 110, 429–441.
Wallace, S.R., 1974. The Henderson ore body: elements of discovery, reflections. Am. Inst. Min. Engr. Trans. 256, 216–227.
Webster, J.D., Raia, F., De Vivo, B., Rolandi, G., 2001. The behaviour of chlorine and sulfur during differentiation of the Mt. Somma-Vesuvius magmatic system. Mineral. Petrol. 73, 177–200.
Williams, A.E., McKibben, M.A., 1987. A brine interface in the Salton Sea geothermal system, California: mechanism for active ore formation. Geol. Soc. Am., Abstr. prog. 19, 890.