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ABSTRACT

Campi Flegrei (Phlegrean Fields) is an active volcanic center near Naples, Italy. Numerous eruptions have occurred here during the Quaternary, and repeated episodes of slow vertical ground movement (bradyseism) have been documented since Roman times. Here, we present a quantitative model that relates deformation episodes to magma degassing and fracturing at the brittle-ductile transition in a magmatic-hydrothermal environment. The model is consistent with field and laboratory observations and predicts that uplift between 1982 and 1984 was associated with crystallization of ~0.83 km^3 of H_2O-saturated magma at 6 km depth. During crystallization, ~6.2 × 10^{10} kg of H_2O and 7.5 × 10^8 kg of CO_2 exsolved from the magma and generated ~7 × 10^{15} J of mechanical ($P\Delta V$) energy to drive the observed uplift. For comparison, ~10^{17} J of thermal energy was released during the 18 May 1980 lateral blast at Mount St. Helens.

Keywords: Campi Flegrei, bradyseism, magmatic fluids, volcanic hazards, magmatic-hydrothermal system, eruption dynamics.

INTRODUCTION

Slow, vertical ground movement centered on the town of Pozzuoli at Campi Flegrei, Italy (Fig. 1), is referred to as "bradyseism" from the Greek words meaning "slow movement." The first studies of ground movement at Campi Flegrei involved sea-level markers at the Serapis Temple archaeological site in Pozzuoli. Boreholes left by a marine mollusk (*Lithodomus lithophagus*) on columns of this monument (Fig. 2) attracted the attention of early researchers (Breislak, 1792; Lyell, 1872), and Parascandola (Parascandola, 1947) reconstructed the history of secular ground movements using the boreholes. This work was later updated using reconstructions of the history of vertical ground movement at Campi Flegrei in recent times (Dvorak and Mastrolorenzo, 1991).

Radiocarbon dating of biological indicators at Serapis (Morhange et al., 2006) documents three 7 m relative sea-level highstands, one during the fifth century A.D., another during the early Middle Ages, and a third in 1538, when the ground rose ~7 m before the Monte Nuovo eruption (Fig. 3). Bradyseism was again active in 1970–1972 (Fig. 3), when uplift of 1 m occurred; this was followed by subsidence of 30 cm, without an eruption. Another episode of uplift of 1.8 m occurred from 1982 to 1984, with minor uplifts in 1988–1989, 1993–1994, and 2000–2001 (Fig. 3). Between periods of uplift, the bases of the columns at Serapis are underwater (Fig. 2, right), whereas during uplift events, the water level is below the base of the columns (Fig. 2, left).

The cause of uplift at Campi Flegrei is controversial and has important implications concerning volcanic risk in this area, which is home to several hundred thousand people. As a result of uplift in the early 1980s, 30,000 people were evacuated from Pozzuoli and relocated to a new town at Monte Ruscello, ~3 km away (yet still within the Campi Flegrei caldera!). This period of uplift, and others, did not produce an eruption, leading to uncertainty concerning steps that should be taken to protect public safety during future uplift events. In 2007, Campi Flegrei is entering a renewed phase of bradyseism; thus, a scientific understanding of the causes and possible consequences of uplift has important and immediate societal ramifications.

Models proposed to explain ground deformation at Campi Flegrei may be divided into two end-member types—those that involve only the input of magma at depth to explain uplift (mechanical models) and those in which fluids (either indigenous or externally derived) play an important role (magmatic-hydrothermal models). A recent comprehensive review (De Natale et al., 2001) discussed these models and considered the pros and cons of each. We interpret ground movement at Campi Flegrei to be the result of magmatic-hydrothermal processes associated with crystallization of hydrous magma at depth, in a manner analogous to the formation of porphyry copper deposits. These deposits involve the emplacement of hydrous magmas that crystallize and introduce magmatic fluids into the shallow hydrothermal system (Burnham, 1979). As the system evolves, melt and fluid inclusions with distinct properties are trapped in phenocrysts and veins in these deposits (Bodnar, 1995; Bodnar and Student, 2006), and similar melt and fluid inclusions have been observed in samples from Campi Flegrei (De Vivo et al., 1989; De Vivo et al., 1995).

MAGMATIC-HYDROTHERMAL MODEL

The Campi Flegrei stratigraphic sequence is well constrained based on the 3000-m-deep San Vito and Mofete geothermal wells (Fig. 1)

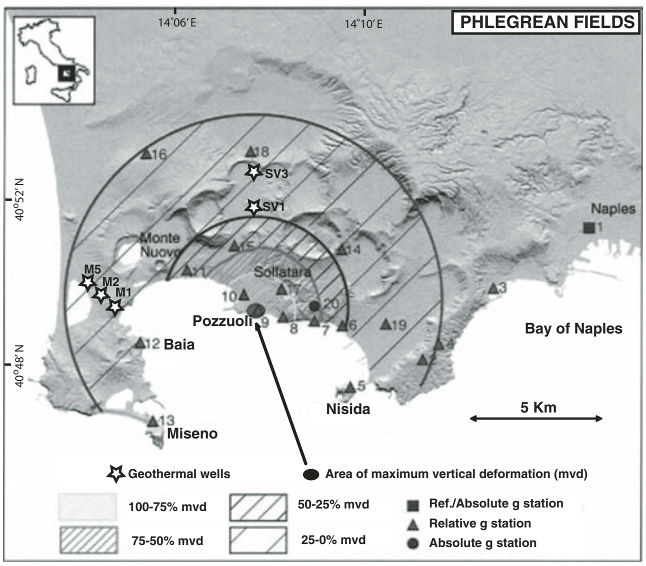

Figure 1. The Phlegrean Fields (Campi Flegrei) caldera, with the area of maximum vertical ground movement centered on Pozzuoli (arrow), site of the Serapis Temple. Also shown are locations of the 1538 Monte Nuovo eruption and Mofete (M1, M2, M5) and San Vito (SV1, SV3) geothermal wells. (Modified from Todesco and Berrino, 2005).

Figure 2. The ancient Roman marketplace of Serapis in Pozzuoli, Italy, shown during periods of uplift (left) and subsidence (right). The dark areas near the bottom of the three columns contain mollusk shells and borings, indicating that the columns were below sea level at some time in the past. The three large columns in the right foreground are each approximately 9 m tall.

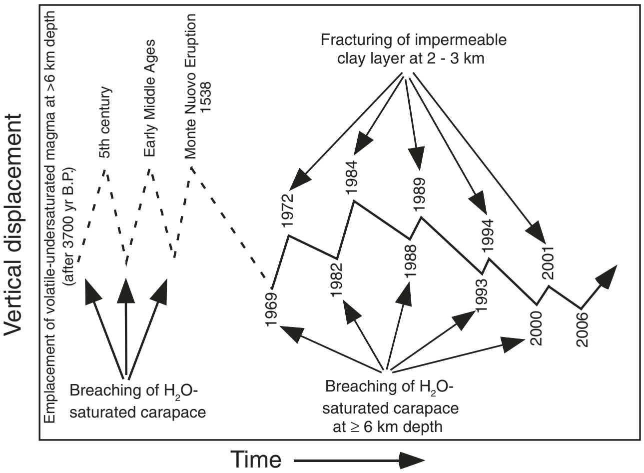

Figure 3. Vertical ground displacement versus time in the vicinity of Pozzuoli, Italy (modified from De Natale et al., 2001; Dvorak and Mastrolorenzo, 1991). Note that both vertical displacement and time are schematic and only indicate the direction of ground movement at different times.

(De Vivo et al., 1989), seismic studies (Judenherc and Zollo, 2004), and unpublished data. The uppermost 2000 m is composed of recent volcanoclastic rocks with minor trachytic lavas. Of specific relevance to the current study is an impermeable layer at a depth of 2–3 km. In the San Vito boreholes, located toward the center of the Campi Flegrei caldera (Fig. 1), a systematic increase in metamorphic grade with depth is observed below 2 km. The thermal profile associated with this metamorphic aureole suggests that at an earlier time, possibly during the active period from 4.5–3.7 k.y. B.P. (Rosi et al., 1983), magma occurred at a shallower depth (4–5 km) than today (≥6 km).

During the repose time following the period of peak volcanism, which ended ca. 3.7 k.y. B.P., we suggest that a new batch of volatile-undersaturated magma was emplaced beneath Campi Flegrei (Fig. 4A). The magma began to crystallize, and an impermeable rock rind developed at the top and sides of the magma body, isolating the underlying magma from the overlying rocks (Burnham, 1979). As anhydrous phases crystallized, the melt became saturated in volatiles, and a volatile-saturated carapace composed of crystals + melt + fluid developed at the top of the magma chamber and below the crystalline rind (Fig. 4B). Coexisting melt and fluid inclusions (Fig. 4H) characteristic of magmatic-hydrothermal systems are trapped in this environment (Bodnar and Student, 2006; Roedder and Bodnar, 1997).

The exsolution of volatiles (here expressed as H_2O) from crystallizing hydrous melts in a closed system may be represented by the reaction:

$$\text{Hydrous melt} \rightarrow \text{anhydrous crystals} + H_2O. \quad (1)$$

Reaction 1 involves a large, positive volume change (Fig. 5), owing to the difference in the partial molar volume of H_2O dissolved in the melt compared to the molar volume of H_2O at magmatic pressure-temperature (*P-T*) conditions (Student and Bodnar, 1996). The mechanical energy ($P\Delta V_r$, where ΔV_r is volume increase) released in the reaction hydrous melt → anhydrous crystals + H_2O can be large (Burnham and Ohmoto, 1980) (Fig. 5). In isolated small pockets in crystallizing H_2O-saturated magma, values of ΔP_{in} (internal overpressure) ≥ 500 MPa are possible and can cause brittle failure of the impermeable rind separating the volatile-saturated magma below from wall rocks above. Fluids and magma are then released into the overlying rocks (Fig. 4C).

The area that experienced uplift at Campi Flegrei during 1982–1984 may be approximated by an inverted cone with a diameter of 12 km and a height of 1.5 m at the center (De Natale et al., 2001), corresponding to an uplifted volume of ~0.05 km^3. Following Burnham (1985), crystallization of an H_2O-saturated magma at 6 km results in an 8% volume increase according to Equation 1 (Fig. 5). This volume increase assumes a pure H_2O volatile phase, but the magma at Campi Flegrei contains both H_2O and CO_2 (Chiodini et al., 2001). At 1100 °C and 170 MPa, ~4 wt% H_2O is soluble in a basaltic magma (Newman and Lowenstern, 2002). However, if the melt is saturated in CO_2 (= 358 ppm; Newman and Lowenstern, 2002), the solubility of H_2O is 3 wt%, or ~75% of the H_2O solubility in a CO_2-free melt. Since the volume change associated with crystallization of hydrous magma shown in Figure 5 is based on a CO_2-free melt, the volume change for a melt containing 358 ppm CO_2 in addition to H_2O would be only 75% of that for the CO_2-free system. Thus, we estimate that crystallization of volatile-saturated melt containing 358 ppm CO_2 and 3 wt% H_2O at 6 km produces a volume increase of ~6%.

If the volume associated with ground deformation at Campi Flegrei equals the volume increase (ΔV_r) of the magmatic system during crystallization shown on Figure 5, then the 0.05 km^3 uplift volume requires crystallization of ~0.83 km^3 of magma (i.e., 6% of 0.83 km^3 = 0.05 km^3). Approximating the shape of the magma body as a cylinder, 0.83 km^3 of magma corresponds to the top 265 m of a magma body with a radius of 1 km, or the top 66 m for a magma chamber with a radius of 2 km.

During crystallization of 0.83 km^3 of H_2O- and CO_2-saturated melt at 6 km, ~6.2 × 10^{10} kg of H_2O and 7.5 × 10^8 kg of CO_2 will be released into the magmatic-hydrothermal system. This amount of CO_2 is comparable to the amount (5000 t/d) associated with the "crisis" period between 1982 and 1984 (Todesco and Berrino, 2005).

According to Figure 5, crystallization and exsolution of the magmatic fluid generates ~7 × 10^{15} J of mechanical ($P\Delta V$) energy (Burnham, 1972), which is sufficient to accomplish the observed uplift. For comparison, the amount of energy associated with the lateral blast that destroyed Mount St. Helens in 1980 was estimated to be ~10^{17} J (Kieffer, 1981).

We emphasize that we are not proposing that the entire 0.83 km^3 of magma crystallized during the two-year period between 1982 and 1984. During some unknown number of years before 1982, magma was crystallizing and releasing fluid, but the fluid was trapped below the overlying impermeable, previously crystallized rocks (Fig. 4E). When this impermeable crystallized rind was breached in 1982, fluids escaped into the overlying rocks (beneath the claystone-siltstone impermeable layer) and began the uplift process (Fig. 4F). Catastrophic fracturing of the crystallized rind and the concomitant pressure drop resulted in a "pressure quench" of additional magma and migration of the volatile-saturated carapace to greater depth (compare Figs. 4E and 4F), which led to release of additional CO_2 (and H_2O) from the magma. Uplift resulting from this increased input of magmatic fluids contin-

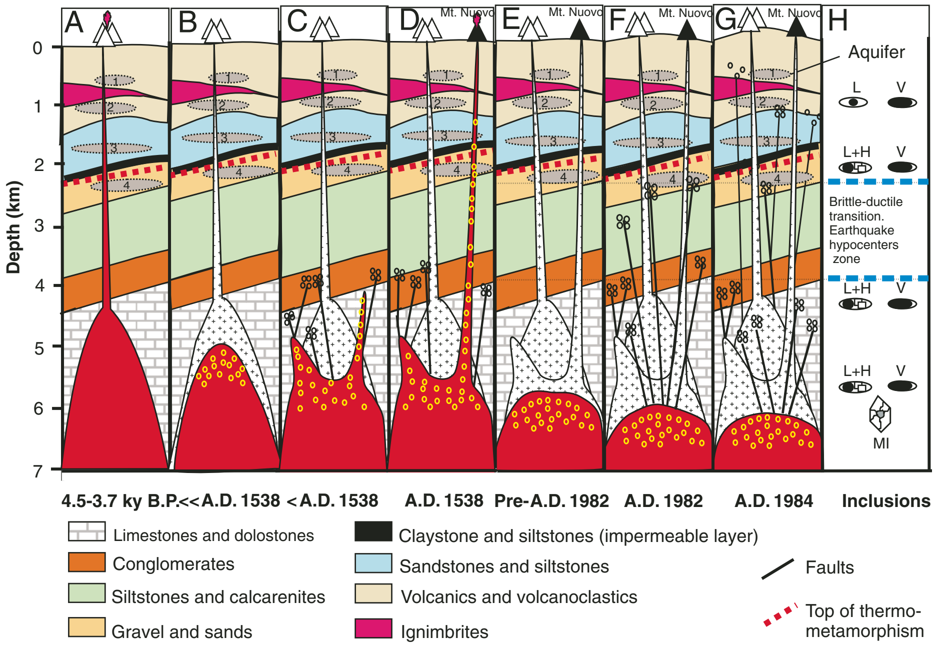

Figure 4. Schematic interpretation of subsurface magmatic-hydrothermal activity at Campi Flegrei since 4.5 k.y. B.P. (A) The last period of highly active volcanic activity was between 4.5 and 3.7 k.y. B.P., when numerous eruptions occurred. (B) During the period between 3.7 k.y. B.P. and the 1538 Monte Nuovo eruption, a volatile undersaturated magma was emplaced. An H_2O-saturated carapace developed under the impermeable rock rind surrounding the magma body as crystallization proceeded. (C) Shortly before the 1538 eruption, the rock rind surrounding the magma body fractured, allowing melt and magmatic fluid to escape. The high pressures associated with this event caused uplift of the ground surface immediately before the Monte Nuovo eruption. (D) In 1538, the melt-filled fracture breached the brittle-ductile boundary and intersected a shallow aquifer, leading to an initial hydrothermal eruption that evolved into a phreatomagmatic eruption. (E) Following the 1538 eruption, the magma body again became a closed system, and magmatic volatiles accumulated below the impermeable overlying rocks. (F) In 1982, the rock rind confining the magmatic system again fractured, allowing magmatic fluids to enter the overlying rocks beneath the brittle-ductile boundary, causing vertical ground deformation. (G) Ground deformation that began in 1982 ended and deflation began when fractures penetrated the brittle-ductile boundary, allowing the deep fluids to migrate into the shallow aquifers (gray oblate spheres labeled 1–4) and flow toward the surface. (H) In the deepest parts of the system, melt (MI) and coexisting liquid-rich (L) and vapor-rich (V) fluid inclusions are trapped. Following breaching of the impermeable rock rind, magmatic fluids escape into the overlying rocks, where phase separation occurs and coexisting high salinity (halite-bearing [H]) and vapor-rich inclusions are trapped. In the shallowest aquifers, high-salinity magmatic fluids mix with low-salinity meteoric water and seawater to produce boiling fluids that become less saline upward, as evidenced by halite-bearing inclusions at depth and lower-salinity liquid-rich inclusions at more shallow levels, both of which coexist with vapor-rich inclusions.

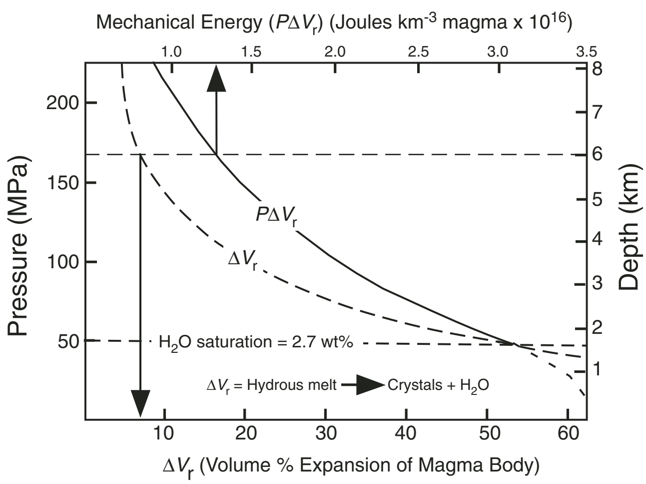

Figure 5. Volume change (ΔV_r) and mechanical energy ($P\Delta V_r$) associated with crystallization of an H_2O-saturated melt (modified from Burnham, 1972, 1985). The calculated values assume a closed system.

ued until the impermeable layer at 2–3 km was breached (Fig. 4G), which allowed complete connectivity between the deeper magmatic environment and the shallow aquifers. Evidence for connectivity between the deep and shallow systems is provided by CO_2/H_2O ratios of fumarolic fluids at Campi Flegrei, which increase during uplift and reach a maximum shortly after deflation begins (i.e., after the impermeable claystone and siltstone layer is breached).

As a result of fracturing of the overlying rocks, the magma body becomes an open system, pressure decreases, and magma and boiling fluids escape. Crystallization of magma and precipitation of minerals from fluids result in resealing of the magma body to again produce a closed system, as shown in Figure 4E. At this stage, the magma body is restored to its original state before fracturing, except that the H_2O-saturated carapace has migrated to greater depth, as illustrated schematically by comparing Figures 4D and 4E. Further crystallization produces magmatic fluid in the upper portion of the magma chamber, which increases the pressure and leads to reactivation of the same processes that operated before, as shown in Figure 4F.

TESTING THE VARIOUS MODELS

Various observations may be used to test models that have been proposed to explain bradyseism at Campi Flegrei (Table 1). While magma emplacement can result in surface uplift, deflation does not generally occur unless inflation leads to an eruption (Newhall and Dzurisin, 1988). Eruptions associated with uplift at Campi Flegrei are rare; the only documented occurrence in the past 2000 yr was the Monte Nuovo eruption in 1538 A.D. (Fig. 4D), and possibly a small phreatic event in 1198 A.D. (Rosi and Sbrana, 1987). Emplacement and crystallization of a magma to explain bradyseism at Campi Flegrei, without the production of magmatic fluids, is also inconsistent with observed increased fumarolic activity (Chiodini et al., 2003), CO_2/H_2O ratios of fumarolic fluids (Chiodini et al., 2003), H, C, O, He, Ne, and Ar isotopic evidence for magmatic fluids (Allard et al., 1991; Panichi and Volpi, 1999; Todesco and Scarsi, 1999), the heat budget at Campi Flegrei (Chiodini et al., 2003), gravity modeling (Gottsmann et al., 2006), and fluid inclusions (Bodnar, 1995; De Vivo et al., 1989, 1995). Moreover, the magmatic model requires a magma chamber at a depth of ≤2.5 km, and, while some studies suggest a magma body at 3–4 km (De Natale et al., 2001), more recent studies (Zollo et al., 2003) suggest that a magma body, if present, is at least 6 km deep. The increased seismicity at 2.5–4.0 km depth (Ferrucci et al., 1992) is now thought to represent the brittle-ductile transition (Fig. 4) rather than a magma body.

According to our model, as the magma cools and the H_2O-saturated carapace migrates to greater depth, the energy and volume change asso-

TABLE 1. CONSISTENCY OF OBSERVATIONS WITH THE MECHANICAL AND MAGMATIC-HYDROTHERMAL MODELS FOR BRADYSEISM AT CAMPI FLEGREI

Observation	Magma emplacement (mechanical) model	Magmatic-hydrothermal model
Uplift followed by deflation w/o an eruption	No[*]	Yes[†]
Increased CO_2-H_2O during uplift	No[§]	Yes[#]
Seismicity at 3–4 km depth	No[**]	Yes[**]
Seismicity at 3–4 km depth	Yes[††]	Yes[**]
Increased fumarolic activity during uplift	Yes[§§]	Yes[§§]
Thermal energy budget	No[§§]	Yes[§§]
Magmatic fluids in fumaroles	No[##]	Yes[##]
Gravity models	No[***]	Yes[***]
Fluid inclusions	No[†††]	Yes[†††]

[*]Newhall and Dzurisin (1988).
[†]Burnham (1972, 1985), Burnham and Ohmoto (1980).
[§]Assumes no input of fluid from magma.
[#]Chiodini et al. (2001).
[**]Ferrucci et al. (1992).
[††]De Natale et al. (2001).
[§§]Chiodini et al. (2003).
[##]Allard et al. (1991), Panichi and Volpi (1999), Todesco and Scarsi (1999).
[***]Gottsmann et al. (2006).
[†††]De Vivo et al. (1989, 1995), Bodnar (1995).

ciated with volatile exsolution decreases (Fig. 5). As such, the likelihood of an eruption at Campi Flegrei is lower today than at any time in the past 500 yr and is expected to continue to decrease. A future volcanic eruption is likely only if a new batch of magma is emplaced into the subsurface.

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