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THE SEARCH FOR MAGMATIC RESERVOIRS

FREDERICO MACHADO

Junta de Investigações do Ultramar, Lisbon (Portugal)

INTRODUCTION

The origin of magmas is still an open problem. Present trends of thought suggest that magma can be derived from the upper mantle, where it fills small pockets. When extruded from the mantle, this magma will rise to shallow depths, owing to the pressure of the overlying solid crust. An important question is whether volcanic activity is fed by vents discharging directly from the upper mantle, or from shallow reservoirs, where the magma stays for some time.

At present, it is assumed that most volcanoes have a relatively shallow magma chamber (cf. Macdonald, 1961). These chambers are, however, inaccessible to direct investigation. A convenient tool for research of this type is the usual seismic prospecting; unfortunately any magma layer will behave as a wave guide and, in this case, prospecting is unable to provide useful results. We must therefore rely upon other indirect methods which will be presently reviewed. On the whole, there is a considerable amount of evidence, but its reliability is frequently questionable.

PLUTONS, XENOLITHS AND DIFFERENTIATION

Intrusive bodies of plutonic rocks, embedded in the upper levels of the crust, are present in many areas. Some are definitely associated with volcanic rocks (Ustiyev, 1963) and may represent solidified magma chambers which fed some past volcanic activity. Very convincing evidence has been found in Scotland (see Richey, 1961), where the roots of the volcanoes were eroded by Quaternary glaciations. A detailed interpretation of the field surveys is rather difficult, but the data strongly suggest that magma chambers resulted from the sinking of conical blocks (subterranean cauldron subsidence), as shown in Fig.1. Magma seems to rise from the upper mantle through a ring dyke, and only after staying in the chamber it rises to the Earth's surface.

The form of the chamber varies widely, depending considerably on the structure of the surrounding rocks (Daly, 1933, Ch. 4); within sedimentary layers we can find laccoliths or phacoliths, whereas in an igneous

Fig 2

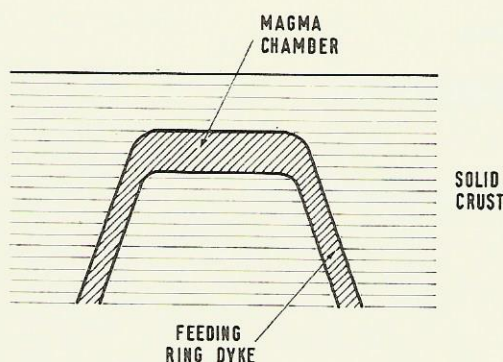


Fig.1. Scottish model of magma chamber (formed by subterranean cauldron subsidence).

environment more irregular bodies are expected. Ring structures are fairly common on continental areas; but they are also found on oceanic (or sub-oceanic) islands (Assunçao et al., 1968; Fúster et al., 1968, fig.48). Elongated structures are certainly frequent and many bodies described as sills (Daly, 1933, p.77) may probably also be considered as old magma chambers.

Gravitative differentiation is a usual feature of these intrusive bodies; they form layered intrusions, which at different levels can have compositions analogous to various volcanic lavas, and sometimes, in addition, exhibit at the bottom ultramafic cumulates formed mostly of olivine crystals (Jackson, 1967). This geological evidence suggests that actual active volcanoes can have shallow magma chambers which will eventually solidify into similar plutons. But the possibility of some volcanoes having no shallow magma chamber cannot be excluded.

The best-known case of xenoliths, giving information on a magma chamber, comes from the pyroclastic layers of Vesuvius (Rittmann, 1936, p.159). An explosive eruption, in the 12th century B.C., produced pumice layers with xenoliths of sediments down to the Triassic. The stratigraphy of the area is fairly well known and was lately confirmed by seismic studies (Imbò, 1950); the Triassic layers, which seem to form the roof of the magma chamber, lie at a depth of about 5 km. The Triassic xenoliths experienced considerable contact metamorphism, as would be expected.

Another source of indirect evidence comes from differentiated lavas. In fact, it appears that magma in the upper mantle is fairly uniform, as suggested by the similarity of the main (basic) volcanic rocks in widely distant areas of the Earth. Differentiation can, therefore, be regarded as a local phenomenon occurring in a separate chamber; the hypothesis of these shallow reservoirs has very often been invoked to account for differentiation (see, e.g., Maleev, 1964, p.218).

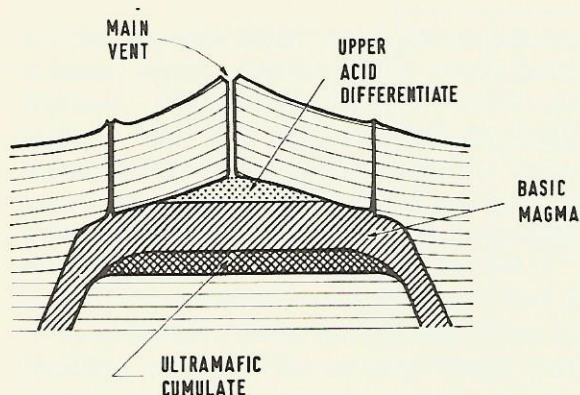


Fig.2. Idealized section of differentiated magma chamber.

Friedlaender (1929) drew attention to the fact that in many Azorean volcanoes the central vent produced acid lavas, whereas most peripheral cones are basaltic. This is what would be expected if the feeding chamber is vaulted and magma is layered by gravitative differentiation (Fig.2). During the 1563 eruption of the Agua-dePau Volcano (San Miguel, Azores), the activity started at the main vent with a Plinian explosion of trachytic pumice, but four days later basaltic lava flows came from a lower adventive vent (see Zbyszewski, 1963). Similar cases are known in a few other volcanoes.

SEISMOLOGICAL INVESTIGATIONS

As the usual seismic reflection or refraction methods are unsuitable for detecting magma chambers, other seismic methods have been tried for the purpose. Gorshkov (1958) found that seismic shear waves were not propagated at a depth of 50–70 km under some Kamchatka volcanoes. This was interpreted as being due to the presence of a magma layer at that depth (which corresponds to a somewhat high level of the upper mantle).

Fractional melting of a peridotitic upper mantle is indeed considered as the most probable source of volcanic magmas (Wager, 1958; Coats, 1962). Temperatures below a depth of about 50 km, derived from Gutenberg's (1959) seismic velocities, lie between the probable melting points of basalt and of forsteritic olivine (Fig.3).

In addition, the rigidity of the upper mantle (as derived from the velocity of shear waves) is lower than what would be expected in a completely solid peridotitic material (Fig.4); this can give a measure of the assumed molten (basaltic) fraction. In fact, Oldroyd (1956) has shown

$n = 5\%$. On the other hand, the abundance of radioactive matter in the oceanic upper mantle (heat production: $1.3 \cdot 10^{-13}$ cal/cm³ sec; Machado, 1968) compares well with the abundance in basaltic rocks. This suggests that the material could be mostly eclogite (or some kind of metamorphosed gabbroid or basaltic material) down to some 50 km; at lower levels the temperature is probably above the melting point of this rock which would therefore melt into a basaltic magma. Ringwood (1969) assumes for the oceanic upper mantle a "pyrolitic" composition, which would have a much lower radioactive heat production.

According to these hypotheses, either by the exudation of the molten fraction available (in oceanic areas) below about 50 km, or by the remelting of the assumed eclogite layer at that same depth, a basaltic magma can be produced and would accumulate in situ. In some cases, especially in rift eruptions, this magmatic layer can directly feed a volcanic eruption (see Gorshkov, 1967, p.271), as shown in Fig.5. In most cases, however, it appears that mantle magma will rise first to some shallow levels.

Shteynberg (1965), by studying volcanic tremor, obtained fair evidence of magmatic reservoirs at both levels. He thinks that the tremor is produced by a vertical vibration of the magmatic column, which fills the volcanic vent. The law of this vibration can be deduced very easily.

Let u be the displacement at time t of a magma particle in the vertical z direction. Neglecting body forces, the equation of motion (assumed as independent of the other space coordinates) is:

$$\frac{\partial^2 u}{\partial t^2} = -\frac{1}{\rho} \frac{\partial p}{\partial z} \quad (2)$$

p being the pressure and ρ the density (see Lamb, 1945, p.479). Noting that, for small displacements:

$$p - p_0 = -\chi \frac{\partial u}{\partial z} \quad (3)$$

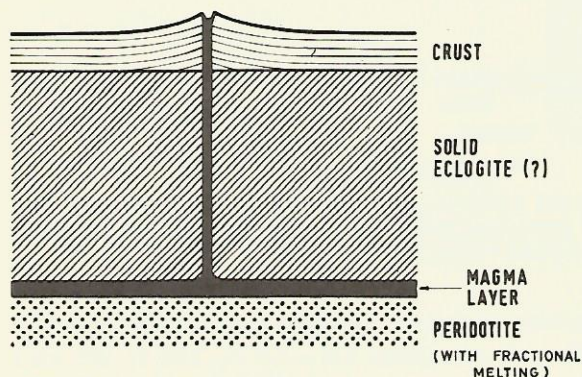


Fig.5. Idealized section of oceanic crust and upper mantle, with magma layer discharging directly to the surface.

where χ is the bulk modulus of the magma and p_0 the pressure in the undisturbed state, substituting in eq.2, we obtain:

$$\frac{\partial^2 u}{\partial t^2} = \frac{\chi}{\rho} \frac{\partial^2 u}{\partial z^2} \quad (4)$$

A solution, which satisfies the conditions $u = 0$ for $z = 0$ and for $z = h$ (h being the height of the magmatic column), is:

$$u = A \left[e^{i\omega(t+z/v)} - e^{i\omega(t-z/v)} \right] \quad (5)$$

where A is a constant, $v = (\chi/\rho)^{1/2}$ is the velocity of compressional waves in the magma, and ω is given by:

$$\omega = 2\pi/T = n\pi v/h \quad (6)$$

T being the period of the vibration and n a whole number.

The fundamental mode of vibration ($n = 1$) has therefore a period:

$$T = 2h/v \quad (7)$$

According to Shteynberg, in addition to the usual tremor with a period of 0.3–0.6 sec (cf. Minakami and Sakuma, 1953), there are components with periods of 2.5–3.5 sec and of 40–55 sec. With $v \cong 3$ km/sec, eq. 7 would give depths of 4 or 5 km and 60–90 km. These values are exactly those one would expect for the depths of the shallow chambers, and of the deep source of magma, respectively (see Fig.6).

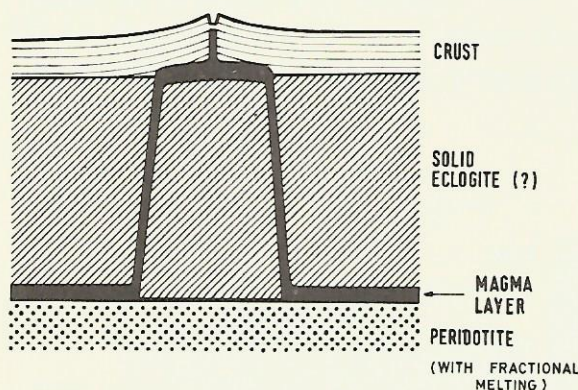


Fig.6. Idealized section of oceanic crust and upper mantle with deep magma layer and shallow magma chamber feeding a surface volcano.

Another type of investigation is based on the irregularities of the isoseismal lines in the vicinity of volcanoes (Machado, 1954; see also Howell, 1959, p. 94). In a homogeneous crust (and with a point focus) isoseismal lines would be concentric circles round the epicentre, each

radius Δ and the corresponding maximum ground acceleration a satisfying the approximate equation:

$$a(\Delta^2 + h^2) = \text{const.} \quad (8)$$

where h is the focal depth.

Maximum acceleration of the ground vibration is related to the intensity I (Modified Mercalli Scale of 1931) by the empirical formula:

$$3 \log a = I - 1.5 \quad (9)$$

The presence of a magma chamber will absorb seismic energy, producing a decrease of the intensities; this represents an anomaly which can be defined as:

$$\delta I = I - I_0 = 3 \log (a/a_0) = 1.3 \ln (a/a_0) \quad (10)$$

the subscript zero referring to the values in the absence of the magma chamber.

This anomaly can be related theoretically to the viscosity of the magma. Let the displacement of a given point be the real part of:

$$u = A e^{i\omega t} \quad (11)$$

where t is the time (with a convenient origin) and A and t , respectively, the amplitude and period of the vibration. The maximum acceleration is:

$$a = A \omega^2 = 4\pi^2 A/T^2 \quad (12)$$

In typical near-earthquakes the maximum acceleration is always associated with compressional waves. For shear waves both A and t are greater than for compressional waves, but as they are roughly proportional to each other (e.g., both 10 times as great) the maximum acceleration is found with the smaller periods. The propagation of a compressional wave is described by the equation:

$$u = A e^{i\omega (t - x/v)} \quad (13)$$

where x is the distance travelled along the wave path, and v the velocity of propagation given by:

$$v = [(\chi + 4\mu/3)/\rho]^{1/2} \quad (14)$$

χ being the bulk modulus, μ the rigidity modulus, and ρ the density.

For a Newtonian liquid of viscosity η , we substitute:

$$\mu = i\omega\eta \quad (15)$$

and obtain, if $\omega\eta/\chi$ is small:

$$v = v_0 \left(1 + \frac{2}{3} \frac{\omega\eta}{\rho v_0^2} i \right) \quad (16)$$

and:

$$u = A e^{-kx} e^{i\omega(t-x/v_0)} \quad (17)$$

with:

$$k = \frac{2}{3} \frac{\omega^2 \eta}{\rho v_0^3} \quad (18)$$

$v_0 = (\chi/\rho)^{1/2}$ being the velocity of compressional waves in the magma.

Maximum acceleration is now:

$$a = A\omega^2 e^{-kx} \quad (19)$$

and if damping in the solid crust is neglected:

$$a/a_0 = e^{-kL} \quad (20)$$

L being the length travelled by the wave inside the magma chamber. Substitution of eq.20 into eq.9 gives finally:

$$\delta I = -1.3 kL \quad (21)$$

For the use of this method, we need an earthquake with the focus suitably situated. We have also to decide what are the theoretical intensities in the absence of any magma chambers.

The theoretical distribution of intensities can usually be chosen in such a way that the anomalies are negative in the "shadow" zones and vanish elsewhere. As the assignment of field intensities is a subjective process, the use of a dense net of accelerometers would be an improvement.

The method, notwithstanding all its limitations, has been used with some success in the Azores and in Sicily (Machado, 1954, 1965). In Fig.7

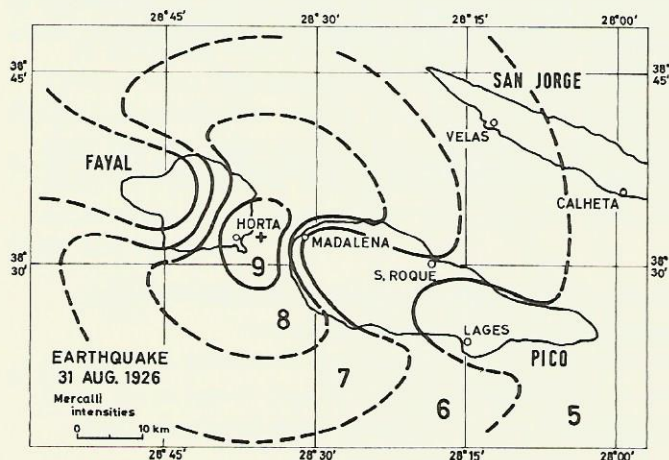


Fig.7. Isoseismal lines of the Azores earthquake of 31 August, 1926 (based on data from Agostinho, 1927).

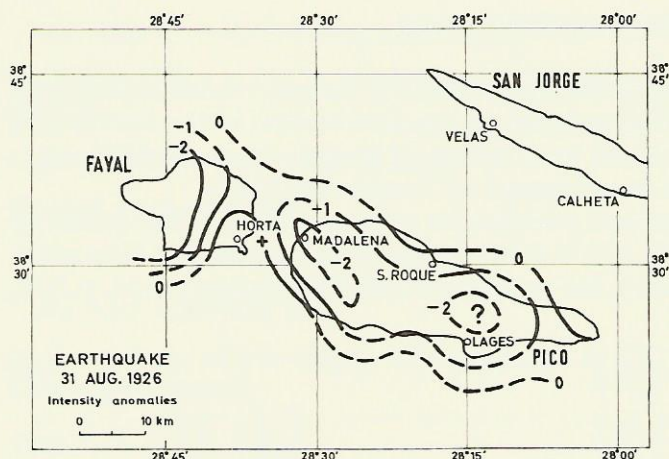


Fig.8. Intensity anomalies of the Azores earthquake of 31 August, 1926 (according to Machado, 1954).

the isoseismal lines of the Azores earthquake of 31 August, 1926 are presented (based on data from Agostinho, 1927). The assumed anomalies and a hypothetical section of the magma chambers of Fayal and Pico volcanoes are shown in Fig.8 and 9, respectively. The depth of 5 km was chosen to bring the centres of the chambers into the vertical of the main vents. The chambers of the individual volcanoes seem to be interconnected, forming a single elongated one.

A reasonable size for the magma chambers was obtained using $k = 0.5 \text{ km}^{-1}$. With this value and $T = 0.2 \text{ sec}$, $\rho = 3 \text{ g/cm}^3$ and $v_0 = 3 \text{ km/sec}$, we obtain, using eq. 18, $\eta = 0.6 \cdot 10^9$ poises. This viscosity is surprisingly high. The mechanism of damping can be slightly more complicated (especially by reflections at the boundaries) but no big change in order of magnitude is expected. At present, only a mush of crystals with an interstitial liquid is believed to have, perhaps, such a high viscosity.

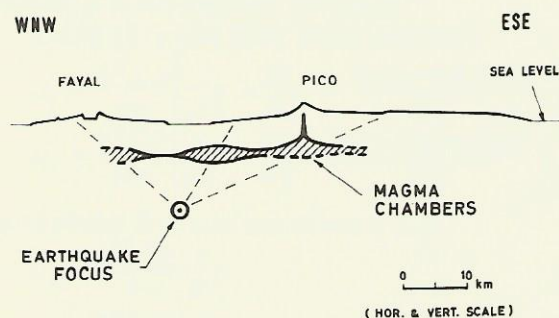


Fig.9. Longitudinal section of Fayal and Pico magma chambers, as deduced from seismic intensity anomalies (according to Machado, 1954).

CRUSTAL DEFORMATION NEAR VOLCANOES

An important approach to the problem of magma chambers is due to Mogi (1958). The method is based on the deformation of the surface of a semi-infinite solid, produced by a change of pressure in a spherical cavity; the theoretical solution was obtained by Yamakawa (1955).

Let r and z be cylindrical coordinates with the origin at the centre of the spherical cavity (Fig.10); b is the radius of the cavity and h its mean depth. With this symmetry, the stresses are (see Timoshenko, 1934, p.309; Love, 1952, p.276):

$$\sigma_r = \frac{\partial}{\partial z} \left(\nu \nabla^2 \phi - \frac{\partial^2 \phi}{\partial r^2} \right) \quad (22)$$

$$\sigma_\theta = \frac{\partial}{\partial z} \left(\nu \nabla^2 \phi - \frac{1}{r} \frac{\partial \phi}{\partial r} \right) \quad (23)$$

$$\sigma_z = \frac{\partial}{\partial z} \left[(2-\nu) \nabla^2 \phi - \frac{\partial^2 \phi}{\partial z^2} \right] \quad (24)$$

$$\tau_{rz} = \frac{\partial}{\partial r} \left[(1-\nu) \nabla^2 \phi - \frac{\partial^2 \phi}{\partial z^2} \right] \quad (25)$$

where ν is Poisson's ratio and ϕ is a function of r and z satisfying the differential equation:

$$\nabla^2 \nabla^2 \phi = 0 \quad (26)$$

Here $\nabla^2 = \partial^2 / \partial r^2 + (1/r) \partial / \partial r + \partial^2 / \partial z^2$.

The boundary conditions are:

$$\sigma_z = 0 \quad \text{for} \quad z = h \quad (27)$$

$$\tau_{rz} = 0 \quad \text{for} \quad z = h \quad (28)$$

$$\sigma_R = -\Delta p \quad \text{for} \quad (r^2 + z^2)^{1/2} = b \quad (29)$$

where Δp is the change of pressure and σ_R is the spherical radial stress given by:

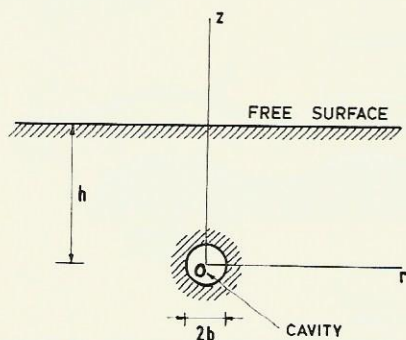


Fig.10. Theoretical model of crust with a spherical "pulsating" cavity.

$$\sigma_R = (\sigma_r r^2 + \sigma_z z^2 + 2\tau_{rz}rz)/(r^2 + z^2) \quad (30)$$

Assuming that $\nu = 1/4$, a solution of eq.26, satisfying conditions (27) and (28), is:

$$\phi = C [\ln (R_1 + z) + 2 (z-h)/R_2] \quad (31)$$

C being a constant and:

$$R_1 = (r^2 + z^2)^{1/2} \quad (32)$$

$$R_2 = [r^2 + (z-2h)^2]^{1/2} \quad (33)$$

In the vicinity of the origin, only the terms in R_1^{-1} are relevant; and we have approximately:

$$\sigma_r = -C \left(\frac{3r^2}{R_1^5} - \frac{1}{R_1^3} \right) \quad (34)$$

$$\sigma_z = -C \left(\frac{3z^2}{R_1^5} - \frac{1}{R_1^3} \right) \quad (35)$$

$$\tau_{rz} = -C \frac{3rz}{R_1^5} \quad (36)$$

and using eq.30:

$$\sigma_R = -2C/R_1^3 \quad (37)$$

Therefore, if b is much smaller than h , condition (29) is satisfied with fair approximation by making:

$$C = b^3 \Delta p/2 \quad (38)$$

The displacements are (see Love, 1952; p.276):

$$u_r = -\frac{1}{2\mu} \frac{\delta^2 \phi}{\delta r \delta z} \quad (39)$$

$$u_z = \frac{1}{2\mu} \left[2(1-\nu) \nabla^2 \phi - \frac{\partial^2 \phi}{\partial z^2} \right] \quad (40)$$

where μ is the rigidity modulus.

Using (31) and (38), these become:

$$u_r = \frac{b^3 \Delta p}{4\mu} \left[\frac{r}{R_1^3} + \frac{2r}{R_2^3} - \frac{6r(z-h)(z-2h)}{R_2^5} \right] \quad (41)$$

$$u_z = \frac{b^3 \Delta p}{4\mu} \left[\frac{z}{R_1^3} + \frac{2h}{R_2^3} - \frac{6(z-h)(z-2h)^2}{R_2^5} \right] \quad (42)$$

and at the free surface where $z = h$:

$$u_r = \frac{3b^3 \Delta p}{4\mu} \frac{r}{R^3} \quad (43)$$

$$u_z = \frac{3b^3 \Delta p}{4\mu} \frac{h}{R^3} \quad (44)$$

with $R = (r^2 + h^2)^{1/2}$. These are Yamakawa's (1955) results.

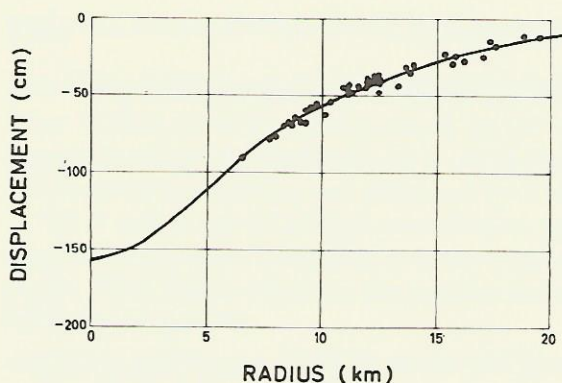


Fig.11. Vertical displacement near Sakurazima Volcano, in the years 1905–1914 (according to Mogi, 1958).

Eq.44 was used by Mogi (1958) for studying the deformation near Sakurazima Volcano (Fig.11). The fit is remarkably good if the mean depth of the magma reservoir is assumed to be about 10 km; usually the radius b cannot be obtained from eq.44 because Δp is unknown.

The same method was applied to Kilauea (Eaton, 1962; Decker et al., 1966; Fiske, 1968), and to Irazù (Murata et al., 1966). In both cases the depth of the “pulsating” chamber was estimated at 3 or 4 km.

A different type of deformation was observed during the eruption of Fayal (Azores) in 1958. After a violent seismic swarm (which preceded the 2nd phase of the eruption) the roof of the assumed elongated magma chamber seems to have buckled in three half waves, each about 5 km wide (Fig.12). The actual crustal strain ϵ and the approximate radius of curvature R could be measured by geodetic surveying (Machado et al., 1962; Machado and Nascimento, 1965). Using the elementary bending theory, the thickness h of the undulated roof can be computed by the equation:

$$h = 2 \epsilon R \quad (45)$$

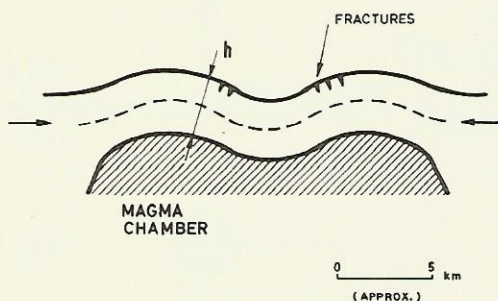


Fig.12. Buckling of the roof of Fayal magma chamber (according to Machado and Nascimento, 1965). The vertical displacement is much exaggerated.

With the measured values $\epsilon \cong 0.5 \cdot 10^{-3}$ and $R \cong 3000$ km (both for one of the convex belts; see Fig.12) a roof thickness $h \cong 3$ km could be estimated. This is compatible with a mean depth of the magma chamber of some 5 km, as obtained from the seismic anomalies.

RATE OF EXTRUSION AND EARTH TIDE CONTROL

Very interesting information about the mechanism of eruption can be obtained from the rate of lava extrusion (Machado, 1962). The flow of lava through a cylindrical vent probably obeys Poiseuille's law:

$$Q = \frac{\pi r^4 \Delta p}{8\eta h} \quad (46)$$

where Q is the rate of flow, r the radius of the vent, Δp the pressure increase at the base of the vent (above hydrostatic equilibrium), η the viscosity of the lava, and h the length of the vent (thickness of the chamber roof).

Let v be the volume of the magma chamber and χ the bulk modulus of the lava. A sudden decrease Δv_0 of this volume will increase the pressure by an amount:

$$\Delta p_0 = -\chi \Delta v_0 / v \quad (47)$$

and when flow starts Δp will change according to the equation:

$$\Delta p = -\frac{\chi}{v} (\Delta v_0 + \int_0^t Q dt) \quad (48)$$

Using eq. 46, the last equation can be written as:

$$Q = Q_0 - A \int_0^t Q dt \quad (49)$$

where:

$$A = \frac{\pi r^4 \chi}{8\eta h v} \quad (50)$$

$$Q_0 = -A \Delta v_0 \quad (51)$$

Differentiating eq. 49, we obtain:

$$\frac{dQ}{dt} + AQ = 0 \quad (52)$$

whose solution, satisfying eq. 49, is:

$$Q = Q_0 e^{-At} \quad (53)$$

As a rule, Δv_0 equals the total volume of extruded lava, so that by the end of the eruption hydrostatic equilibrium is established again.

Eq. 53 was verified approximately for several eruptions: Vesuvius in 1944, Kilauea in 1955, Fayal in 1957–1958 (see Fig.13). From the

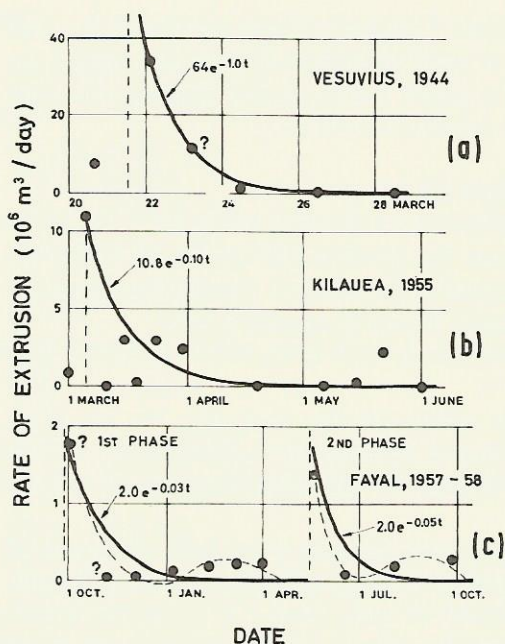


Fig.13. Rate of lava extrusion: (a) in Vesuvius, 1944 (according to Imbò and Bonasia, 1962); (b) in Kilauea, 1955 (according to Macdonald, 1959); (c) in Fayal, 1957–1958 (according to Machado, 1962).

diagrams we can obtain values for A and Q_0 (and Δv_0); we have also some idea of the order of magnitude of χ and η . We can therefore deduce $r^4/h\nu$, but the separate quantities cannot be obtained.

A problem close to the last-mentioned was investigated by Imbò (1954, 1955b) who discovered during the 1944 eruption of Vesuvius a probable vertical oscillation of the lava in the upper part of the vent. We can assume that gas (or vapour) fills the vent, which is only closed by a small upper plug of liquid lava. Let u be the vertical displacement of this plug, m its mass and s the section of the vent. The motion (if frictionless) is described by the equation:

$$m \frac{d^2 u}{dt^2} = s \Delta p \quad (54)$$

where t is time and Δp the excess of pressure at the bottom of the plug over the pressure (mg/s , g being gravity) due to its weight.

Assuming isothermal conditions, we have also:

$$s \Delta p / mg = -u/h \quad (55)$$

h being the height filled with gas.

Using eq. 55 and introducing a friction term, eq. 54 becomes:

$$\frac{d^2 u}{dt^2} + 2\kappa \frac{du}{dt} + \frac{g}{h} u = 0 \quad (56)$$

where k is a friction coefficient. This equation is satisfied by:

$$u = A e^{-k t} \cos (\omega t + \psi) \quad (57)$$

where A and ψ are constants and:

$$\omega^2 = \frac{g}{h} - \kappa^2 \quad (58)$$

From the record of air-borne pressure waves, during a given phase of the eruption, Imbò obtained $\omega = 0.0464 \text{ sec}^{-1}$ and $k = 0.012 \text{ sec}^{-1}$, and, using eq.58, $h = 4.3 \text{ km}$. Apparently the vent, except for a small lava plug, was filled with gas down to the magma chamber, an exceptional situation which certainly led to the following explosive activity.

The effect of semidiurnal Earth tides was recognized in several eruptions, namely in Vesuvius in 1944 (Imbò, 1955a, 1958), in San Jorge (Azores) in 1808 (Canto, 1884; Zbyszewski, 1963), and in the old Kilauea lava lake (Jaggard, 1938).

The effect of the semi-annual tide was also recognized in the Fayal eruption of 1957–1958 (Machado, 1962) and can be observed superimposed on the curve of Fig.13c. The problem is capable of theoretical treatment by including in eq.49 a term proportional to $\sin (\omega t + \alpha)$, $2\pi/\omega$ being the period of the tide and α an adequate constant. We have therefore:

$$Q = Q_0 + Q_m \sin (\omega t + \alpha) - A \int_0^t Q dt \quad (59)$$

where:

$$Q_m = -A \Delta v_m \quad (60)$$

$\Delta v_m/v$ being the cubical expansion corresponding to the maximum of the tide.

Now the solution is:

$$Q = [Q_0 - Q_m \sin \psi \cos (\alpha + \psi)] e^{-A t} + Q_m \cos \psi \sin (\omega t + \alpha + \psi) \quad (61)$$

where $\psi = \tan^{-1} (A/\omega)$.

For the 1957–1958 eruption (2nd phase, Fig.13c) we have approximately $Q_m = 0.5 \text{ million m}^3/\text{day}$, and $A = 0.05 \text{ day}^{-1}$. This gives, by eq. 60, $\Delta v_m = -10 \text{ million m}^3$.

On the other hand, the tidal cubical expansion is:

$$\Delta v_m/v = f z/R \quad (62)$$

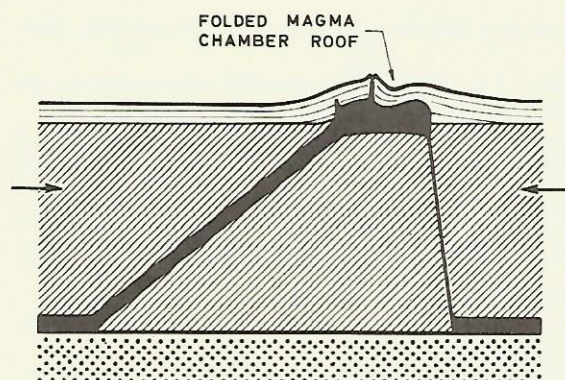


Fig.15. Idealized section of volcano on compressive belt of orogenic type.

Volcanoes are found not only along tensional belts (mid-ocean ridges), but also in the young mountain belts and island arcs, where compression seems to be prevalent. In either case, some difference is expected in both magma type (cf. Gorshkov, 1962) and form of the conduits through which magma rises to the upper levels. In fact, tensional fractures tend to be vertical (Fig.14), whereas in orogenic belts some of the feeding fractures probably correspond to reverse faults, dipping at angles of, say, $30-45^\circ$ (Fig.15). Although shallow chambers seem to be rather frequent in some cases (especially in tensional belts), magma is supposed to rise directly from the upper mantle to the surface (Gorshkov, 1967; Machado, 1969).

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