Mechanism of Magma Ascent and Deep Feeding Channels of Island Arc Volcanoes

S. A. FEDOTOV
Institute of Volcanology, Petropavlovsk-Kamchatskiy, USSR

Abstract

The paper discusses the mechanism of deep magma activity beneath island arc volcanoes and similar structures on the basis of data from geophysical investigations in Kamchatka; the analyses of forces that cause the ascent of magma; and related phenomena that are due to hydrostatic forces.

It is shown that the ascent of magma through the asthenosphere occurs most likely in magma columns with a diameter of approximately 700-2,000 m and with a velocity of about 0.8-3 m/year. A regular line of such columns spaced in Kamchatka at a distance of about 30 km gives rise to a chain of separate large volcanoes or volcanic centers.

Ultrabasic magmas are most likely accumulated near the M discontinuity, whereas the apparent place of andesitic magma accumulation is in the earth's crust near the boundary between the basement and sediments.

Introduction

The ascent of magma and related phenomena constitute one of the major problems of volcanology, petrology of igneous rocks, and geodynamics. Most of the studies dealing with deep processes in island arcs consider a scheme of magma motion from the Benioff zone to the surface. It is supposed that the primary magma is formed in the Benioff zone and then rises to volcanoes. But the question of how magma moves from the Benioff zone to island arc volcanoes is still obscure.

The present paper considers this problem on the basis of numerous data from geophysical investigations in Kamchatka as well as on the basis of physical estimates of the phenomena related to magma ascent.

Geophysical Data on Deep Magma Activity beneath Kamchatka

Detailed seismic investigations, deep-seismic soundings (DSS), gravimetric studies, magnetotelluric soundings, etc. carried out in Kamchatka, enabled the author and his collaborators to obtain information concerning the deep structure and geodynamics of the areas of active volcanism; the relations between volcanoes and processes in tht Benioff zone; and on the deep-seated feeding channels and magma chambers of Kamchatka volcanoes (*Seismicity and Seismic Prediction* ..., 1974; FEDOTOV and TOKAREV, 1974; UTNASIN *et al.*, 1974; FARBEROV, 1974). In this paper the author uses also data on velocities of seismic waves and thickness of the lithosphere beneath the Kurile-Kamchatka arc.

Figure 1 shows the velocity variations of longitudinal waves with depth for various regions of the Kurile-Kamchatka arc. Such velocities at depths from 0 to 50-60 km beneath the volcanic regions of Kamchatka and South Kurile Islands are less than those beneath the wide area including not only the island arc, but also the adjacent areas of the Pacific Ocean and the Sea of Okhotsk. Most of the intrusions, magma chambers and feeding channels appear to be concentrated beneath the volcanic belt at a depth of less than 60 km, above the asthenosphere.

It is known tath the lithosphere may thin beneath tectonically active and volcanic regions. The boundary between the lithosphere and the asthenosphere can be determined from diagrams showing the variation of the number of earthquakes with depth. Diagrams of all the Kamchatka earthquakes (FEDOTOV, 1968; FEDOTOV *et al.*, 1974) and of earthquakes of the focal layer or Benioff zone (TOKAREV, 1974) indicate that the number of earthquakes in the focal layer is maximum at depths of 10-40 km, it decreases sharply at depths of 40-50 km, then it decreases slowly up to the depth of 250 km while it increases slightly at depths of 300 and more kilometers. Elastic deformations in the focal layer must diminish proportionally to the number of its earthquakes. Thus, the boundary between the lithosphere and the asthenosphere beneath the Kamchatka volcanoes seems to be at depths of 40-50 km and not deeper than 60 km.

A correlation between the lower and upper sections of feeding magma columns can be estimated by comparing the calderas area with that of the entire volcanic belt, and the convective heat flow in volcanoes and thermal springs with the mean value of the convective heat flow in the volcanic belt.

In South Kamchatka, the calderas occupy about 4-5 % of the area of the entire volcanic belt. The Kamchatka volcanoes convective heat flow versus the area of their calderas is $(100 \pm 50) \cdot 10^{-6}$ cal/cm^2 sec (KOVALEV and SLEZIN, 1974). The same convective heat

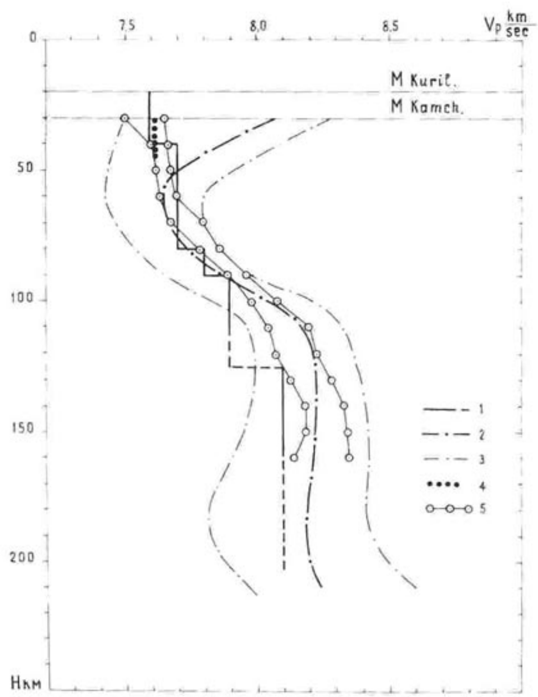

FIG. 1 - Velocity variations of longitudinal seismic waves in the mantle vs. depth: 1: beneath the South Kurile Islands (FEDOTOV and KUZIN, 1963); 2: beneath the South Kurile Islands, Hokkaido, South Sakhalin and adjacent areas of the Pacific Ocean and of the Sea of Okhotsk, a median line (ZHURAVLEVA and ROMANOV, 1972); 3: limits of possible velocity variations for the previous case; 4: mean velocity in the upper mantle beneath East Kamchatka (SLAVINA, FEDOTOV, 1974); 5: limits of possible velocity variations beneath East Kamchatka.

flow passes through the area of primary magma formation, through the calderas and volcanic craters. Convective heat flow, concentration of volatiles, and velocity of ascent of magma in deep feeding channels of volcanoes must increase by 20-25 times minimum from the focal layer to calderas. They can increase several thousand times in the craters of volcanoes.

The following conclusions may be drawn from tht overall geophysical data on deep magma activity beneath East Kamchatka.

The primary magma originates in the focal layer at depths of 100 to 220-250 km from where it rises vertically to volcanoes. About 70 % of magma rises within a band 45 km wide. Magma ascent occurs in the asthenosphere, in the lower layers of the lithosphere and in the earth's crust at depths of 50-250 km, 50-30 km, and 0-30 km, respectively. Some large volcanoes are connected with the focal layer through column-like feeding bodies with cross sections of severall kilometers. The concentration of convective heat flow and volatiles in the feeding magma channels of volcanoes increases 20-5,000 times from the focal layer to the surface. The great bulk of magma is accumulated above the asthenosphere at depths of 0-60 km.

Forces Causing the Magma Ascent

In order to find out how magma rises, it is necessary to consider the forces causing its ascent. This problem has been studied by volcanologists for more than 100 years, but till now which forces act it is still unclear.

The major possible reasons of magma ascent are: 1) the ascent of a layer of magma by a mechanism similar to zone melting; 2) boiling of magma; 3) tectonic pressure; 4) surplus pressure caused by the increase of the magma volume during melting; and 5) buoyancy of magma.

From the analysis of the role played by these forces in the Est Kamchatka volcanic belt (FEDOTOV, 1974) it may be inferred that the magma ascent in the asthenosphere takes place mainly due to buoyancy forces. The stresses arising in the asthenosphere because of the increase of the magma volume during melting, relax so quickly that they can be neglected. In the lower layers of the lithosphere buoyancy forces are aided by tectonic pressure, whereas in the earth's crust they are aided by forces relating to magma boiling. The activity of the latter increases with approaching the surface and becomes dominant during volcanic eruptions.

It is reasonable to consider the mechanism of activity of hydrostatic forces more in detail.

Phenomena Related to the Magma Ascent by Hydrostatic Forces

Using the simple equations of hydrostatic equilibrium, one may determine the correlation between the mean density of the mantle — ρ_2 — and that of the magma column — $\bar{\rho}$ — or between the increase in volume of the mantle material during melting — ΔV —

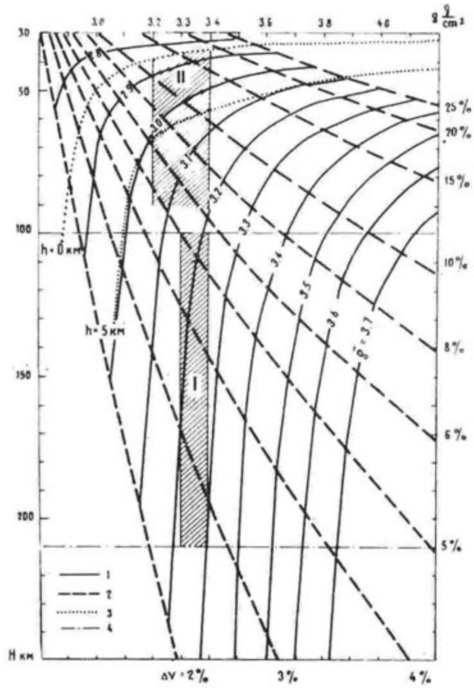

FIG. 2 - Depths from where magma columns can be raised to the surface by hydrostatic forces (the thickness of the crust is 30 km and its mean density is 2.75 g/cm³). H is the depth, ρ_2 is the mantle density, $\bar{\rho}$ is the mean density in the magma column, ΔV % is the volume increase during melting. 1: isolines $\bar{\rho} = f(\rho_2, H)$; 2: isolines $\Delta V \% = \varphi(\rho_2, H)$; 3: depths from where basalt magma with $\bar{\rho} = 2.85$ g/cm³ can be raised to the craters at heights of 0 and 5 km; 4: depth interval in the seismofocal layer where primary magma melting is supposed. I: probable area of primary magma formation, II: probable area of basalt magmas.

and the minimal depth H from where the magma column can be raised to the surface by hydrostatic forces. The diagrams showing these correlations for the East Kamchatka volcanic belt are given in Fig. 2.

These diagrams were constructed taking for the earth's crust the thickness and the mean density of 30 km and 2.75 g/cm^3, respectively.

A magma column of ultrabasic average composition can be raised to the surface only from depths of more than 100 km. If $\Delta V \simeq 3\,\%$, then the base of the magma column should be placed at depths of 170-200 km. At the same time, andesite and acid magmas can be raised to the surface practically from any depth in the mantle. Therefore, one may suppose that considerable amounts of andesite magma are formed in the mantle only beneath areas of andesitic volcanism.

Many geologists and petrologists consider that magma rises from the focal layer of island arcs as large drops of melt — asthenoliths (BELOUSSOV, 1966; RINGWOOD, 1974). Figure 2 shows that this assumption is rather wrong, at least for basic magmas which should rise to the surface in long magma channels or columns with the base at depths of several tens or hundreds of kilometers.

Large volcanoes and composite volcanic edifices (volcanic centers) of the island arc volcanic belts are spaced in chains approximately at the same distance from each other. For instance, a line of stratovolcanoes and caldera volcanoes of Upper Pleistocene-Holocene age extend along the Pacific coast of South Kamchatka with intervals of 25-38 km, averaging 29 km. The physical essence of this phenomenon can be explained by the inversion of densities or gravitational convection during magma flotation in the asthenosphere. This mechanism has been investigated in many geophysical papers, particularly in those dealing with diapiric tectonics (RAMBERG, 1967; USHAKOV and KRASS, 1972). These papers show that a straight chain of domes and vertical channels originates regularly above the layer of floating fluid if this latter is elongated in a certain direction. The distance λ between them depends on the ratio of densities (ρ), viscosities (η) and thicknesses h of two convection layers. From available data and diagrams of $\lambda = f\left(\frac{\rho_1}{\rho_2}, \frac{\eta_1}{\eta_2}, \frac{h_1}{h_2}\right)$ (KRASS and NAPADENSKY, 1972), the primary magma layer beneath the Kamchatka volcanic belt has a thickness from 80 to 240 km. On the basis of seismic data such thickness is 100-150 km. Both approximate estimates are in agreement.

These data evidence once again that the magma ascent in the asthenosphere occurs mainly from buoyancy forces.

The character of current in a viscous fluid can be determined by the Reynolds number $Re = V \cdot \ell \cdot \rho / \eta$, where V is the velocity, ℓ is the characteristic dimension (*e.g.*, the radius of the magma column), ρ is the density of the fluid and η is its viscosity. In our case $Re \ll 1$. Thus, the velocity of stable ascent of a spherical asthenolith V_a will be:

$$V_a \simeq \frac{2 \cdot R^2 \cdot \Delta\rho \cdot g}{\eta_2} \cdot \frac{\eta_2 + \eta}{\left(\frac{2}{3} \eta_2 + \eta\right)} \tag{1}$$

where R is the radius of the asthenolith, $\Delta\rho$ is the difference in density between the asthenosphere and the asthenolith, g is the acceleration of gravity, η_2 is the viscosity of the asthenosphere, and η is the viscosity of the asthenolith magma. When $\eta_2 \gg \eta$

$$V_a \simeq \frac{\Delta\rho \cdot g \cdot R^2}{3\,\eta} \tag{2}$$

Magma columns in the asthenosphere were taken to have a form of a round cylinder with radius R on top of which there is a hemisphere with the same radius R. The approximate velocity of intrusion of this magma column into the asthenosphere, V_k is:

$$V_k \simeq \frac{\Delta\rho \cdot g \cdot (\bar{h} + h) \cdot R}{6\eta_2 + 8 \cdot h \cdot \eta / R} \tag{3}$$

where $\bar{h}$ is the thickness of the feeding magma layer and h is the height of the magma column in the asthenosphere. From (2) and (3) the ratio of the ascent velocities of the magma column and of the asthenolith with the same radius is:

$$\frac{V_k}{V_a} \simeq \frac{h + \bar{h}}{2R} \tag{4}$$

This ratio shows that the velocity of ascent of the magma column is always greater than that of the asthenolith with the same diameter (as many times as the height of the magma column is greater than the diameter of the asthenolith). The velocity of ascent of the magma column can be several orders higher than the velocity of ascent of the asthenolith. This fact supports the conclusion that the magma of island arc volcanoes rises upward through magma channels or columns.

Figure 3 shows the diagrams of (1) and (3) for different values of $\bar{\eta}$ and $\Delta\rho$ when $\eta_2 = 10^{19}$.

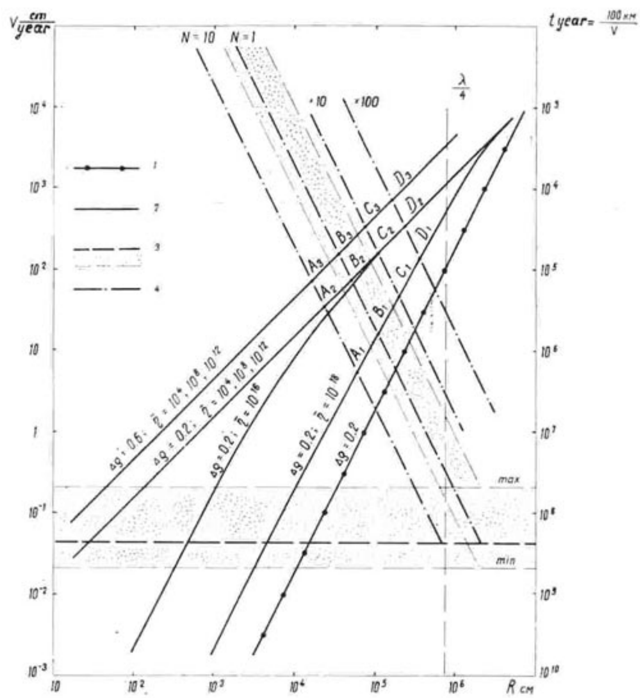

FIG. 3 - Probable radii and velocities of ascent of magma columns and asthenoliths in the asthenosphere. 1: velocity of flotation of a spherical asthenolith vs. its radius; 2: velocity of a magma columns moving through the asthenosphere vs. its radius for different $\Delta\rho$ and $\bar{\eta}$; 3: mean thickness of the layer of volcanic rocks erupted per year in the Eastern volcanic belt of Kamchatka during Quaternary times and limits of its variations; 4: velocity of magma ascent vs radius of the channel; the magma flow discharge changes 1,000 times.

The volume of the magma columns increases proportionally to R^2, and the velocity of their ascent, as shown by (3), is proportional to R. The volume of the magma flowing up a magma column is thus proportional to R^3. It indicates that the predominant part of magma will ascend through a few large channels.

The cross section of these channels is limited by the amount of the magma which is fused in the focal layer. The lower estimate of this amount can be obtained by using data on the volumes of erupted rocks (MELEKESTZEV *et al.*, 1974). Judging from these, beneath the Kamchatka volcanic belt, a layer of magma with average thickness of 0.4 mm was fused annually during Quaternary times. Seventy percent of the Kamchatka volcanoes are located in a band 45 km wide with a mean distance between large volcanoes and volcanic centers of about 30 km. The magma of each volcano can be considered as gathered from an area of about 1,300 km^2. Since the flow rate of the magma channels is proportional to R^3, this focal layer area must be connected with the volcanic center through one or, at the most, very few feeding channels. On the basis of the data listed, a mean dependence between the radius of the channel, R, and the velocity of magma flowing in it, $V(R)$, for the feeding channels of large volcanoes and volcanic centers in Kamchatka can be determined:

$$\lg V(R) \text{ cm/year} \simeq 11.3 - 2 \lg R \text{ cm} \qquad (5)$$

Graph (5) is marked by $N = 1$ in Fig. 3. The limits of $V(R)$ variations corresponding to the observed variations of the average annual volumes of Quaternary volcanites are also shown. If the volcanic center had had not one but ten feeding channels, then the velocities would have been ten times less (graph $N = 10$ in Fig. 3). The visible volume of erupted rocks are obviously smaller than the volume of fused magma. Unfotunately, their true correlation is unknown. Graphs $V(R)$ marked by $\times$ 10 and $\times$ 100 (Fig. 3) are plotted for the case when the volume of the magma flowing up a deep-seated feeding channel of a volcanic center is 10 to 100 times greater than the visible volume of eruptives from this center.

The intersections of graphs (3) and (5) in Fig. 3 give the most probable R values. The area of possible R values lies in the square $A_1 A_3 D_3 D_1$. R values in this area vary from 160 to 3,600 m, and V from 5 to 700 cm/year. The most probable R and V values are in the area $B_2 B_3 C_3 C_2$. There, R = 330-1,000 m and V = 80-300 cm/year respectively, if $\eta_l \leqslant 10^{12}$ poise.

Thicker magma columns can be formed by viscous magma with $\eta_l > 10^{16}$ poise. Columns with a diameter of 2-5 or at the most 7 km can be formed by very viscous magma with $\eta_l \simeq 10^{18}$ poise. The velocity of magma ascent along them is less than 1 m/year. They are probably more like protrusions of ultrabasic material, and not true magma columns.

Although calculations in Fig. 3 are rough, the order of magnitude, judging from additional estimates, seems to be correct.

It is possible that the thickness of the considered columns may increase in the area of transition from the asthenosphere to the lithosphere where the mantle viscosity increases considerably.

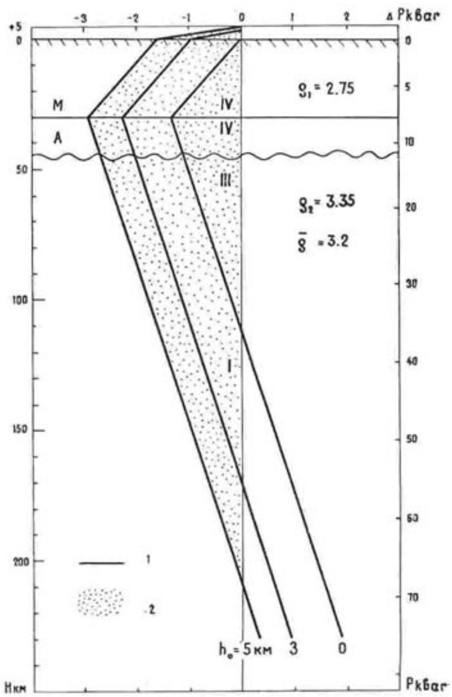

FIG. 4 - Difference between lithostatic and magmatic pressure for magma channels of ultrabasic composition with mean density of 3.2 g/cm^3. 1: isolines $\Delta P\,(H)$ for volcanoes 0, 3 and 5 km high; 2: area of surplus magmatic pressure. M: the bottom of the Earth's crust. A: possible position of the roof of the asthenosphere for East Kamchatka, P: lithostatic pressure.

The magmatic activity in the lithosphere depends, on a considerable extent, on the value of the surplus magmatic pressure, *i.e.* the ΔP difference between the hydrostatic pressure in the magma column and the lithostatic pressure of surrounding rocks. A certain difference of these pressures exists in the open magma column. Figures 4, 5 and 6 show the diagrams of ΔP versus depth for the Kamchatka volcanoes with heights of 0, 1, 2 and 5 km. In calculations, ultrabasic, basaltic and andesitic magmas were assumed to be non-compressible fluids with mean densities of 3.2, 2.85 and 2.65 g/cm^3 respectively. It is concluded that the surplus magmatic pressure in columns of ultrabasic magma can reach 1,000-3,000 bar near the bottom of the earth's crust (Fig. 4). In basalt columns it is maximum in the lower layers of the crust (ΔP = 300-2,000 bar; Fig. 5). In andesite columns ΔP is maximum in the upper layers of the crust, espe-

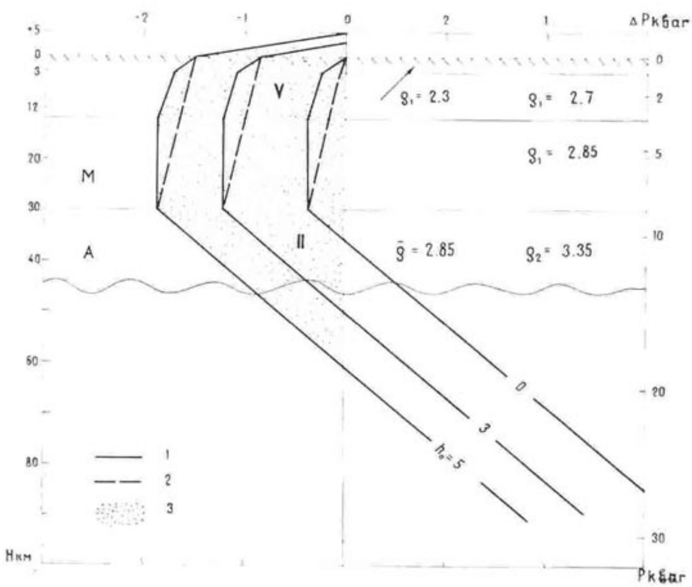

FIG. 5 - Difference between lithostatic and magmatic pressure (ΔP) for basalt magma columns with mean density of 2.85 g/cm^3. 1: isolines $\Delta P(H)$ for volcanoes 0, 3 and 5 km high for a one-layer crust with density of 2.75 g/cm^3; 2: $\Delta P(H)$ isolines for a three-layer crust with mean density as in 1; 3: area of surplus magmatic pressure. M, A and P are the same as in Fig. 4.

cially at the boundary between sediments and the crystalline crust (ΔP = 100-1,000 bar; Fig. 6). In the last case, the real ΔP may be considerably larger because in the fragile rocks the lateral pressure is only 0.3-0.7 of the vertical pressure at depths of several kilometers.

At the depths where ΔP is great, magma channels can widen and form large intrusions. Vast accumulations of ultrabasic magmas near the M discontinuity may create the transitional layer beneath the Kamchatka volcanoes (UTNASIN *et al.*, 1974; BALESTA *et al.*, 1974). Judging from the calculations, the most likely place for peripheral magma chambers of andesite composition is the boundary between the crystalline basement and sedimentary rocks.

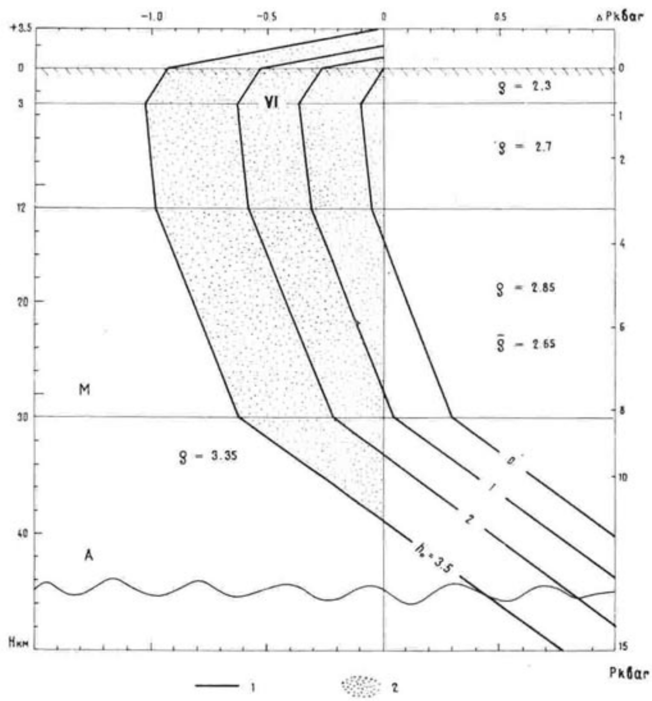

FIG. 6 - Difference between lithostatic and magmatic pressure (ΔP) for andesite columns with mean density of 2.65 g/cm³. 1: isolines $\Delta P(H)$ for volcanoes 0, 1, 2 and 3.5 km high; 2: area of surplus magmatic pressure. M, A and P are the same as in Fig. 4.

The above results do not depend upon petrological or geodynamic concepts. On the basis of the above data one may construct a probable geophysical model of deep magmatic activity beneath the active volcanic belt of East Kamchatka, all the propositions of which are supported by either geophysical observations or calculations.

Kamchatka is a typical link of the Pacific orogenic belt. Deep magmatic activity in other parts of this belt beneath island arcs and similar structures may be like that beneath East Kamchatka.

Probably the results obtained can be of help in studying the petrologic, thermodynamic, geodynamic and other aspects of the problem.

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