Bull Volcanol (1994) 56:459–465

ORIGINAL PAPER

M. A. Alidibirov

A model for viscous magma fragmentation during volcanic blasts

Received September 8, 1992 / Accepted May 18, 1994

Abstract Fragmentation of magma containing gas bubbles is of great interest in connection with developing models for the formation of pyroclastics and for volcanic blasts (explosions). This paper considers the problem of fragmentation of highly viscous ($\eta > 10^8$ Pa s) or solidified magma containing bubbles with excess gas pressure. It is suggested that the fragmentation of magma be considered on the basis of the fragmentation wave theory proposed by Nikolsky and Khristianovich, which is generally applicable to gas-dynamic phenomena occurring in mines. Then it becomes possible to derive the equations of conservation for the fragmentation wave front which moves into a body of magma from its free surface. As a result, the velocity, N, of magma fragmentation, and the velocity, u, of the movement of the gas-pyroclastic mixture behind the fragmentation wave front, are determined. Calculations show that N can reach ~ 5 m/s. Therefore the duration of the fragmentation of the magma body (blast duration) proves to be long. The suggested model explains the possibility of several explosions during the blast as a result of the fragmentation wave stopping, and accounts for the angular shape of pyroclasts by the brittle disruption of interbubble partitions during fragmentation wave propagation through the porous magma body. The initiation and cessation of fragmentation are defined by magma porosity, magma tensile strength, and the pressure differential between gas pressure in pores and the atmospheric pressure. The physical model of magma fragmentation developed explains the mechanism of energy release during volcanic blasts of the Vulcanian or Pelean types.

Mikhail A. Alidibirov
Institute of Volcanic Geology and Geochemistry, Russian Academy of Sciences, Petropavlovsk – Kamtchatsky, 683006, Russia

Present address:
Bayerisches Geoinstitut, Universität Bayreuth, D-95440 Bayreuth, Germany

Introduction

Explosive volcanic eruptions can be classified as magmatic or hydromagmatic. In the first case, a gas-pyroclastic mixture is ejected as a result of energy release from a compressed gas. In the second case this ejection is related to interaction of magma with ground or surface water (Wohletz 1983). The process of gas-pyroclastic mixture ejection from a vent may be steady or unsteady. Plinian eruptions are a close analog of a steady regime, and eruptions during which discrete explosions occur are related to an unsteady regime. While models for steady explosive eruptions are well developed (Slezin 1979; Wilson et al. 1980; Carey and Sigurdsson 1985), current models for volcanic blasts (explosions), such as Vulcanian events, have not explained to date all the observed phenomena. In most models, the blast is treated as a process of rapid expansion of some volume of originally compressed gas or gas-pyroclast mixture (Self et al. 1979; Wilson 1980; Kovalev et al. 1980). This volume is supposed to be originally separated from the atmosphere by a lava or pyroclastic plug, and the problem related to formation of the gas-pyroclastic mixture is not considered. Occasionally, the coalescence of bubbles is considered as the mechanism of magma fragmentation, but it was shown by McBirney and Murase (1970) and Sparks (1978) that coalescence is unlikely for highly viscous magmas. This study considers a model for highly viscous magma fragmentation which is essential for understanding the volcanic blast mechanism.

The approach to the problem and the main assumptions

We consider the fragmentation of magma during discrete blasts, which occur during eruptions of Vulcanian or Pelean types, when andesite, dacite or more acid low-temperature highly viscous magma is fragmented. The ejected fine-grained pyroclastic particles often have angular shapes. Many authors treat this as a result of fragmentation of magma being in an almost solid state (Gorshkov and Bogoyavlenskaya 1965). The volume of pyroclastics ejected can reach $\sim 10^6\,m^3$ during Vulcanian explosions and $\sim 10^8\,m^3$ during large "directed" blasts (Self et al. 1979; Bogoyavlenskaya et al. 1985). The blasts last from a few seconds to several tens of seconds, and the velocity of the ejecta can reach several hundred m/s. Some blasts consist of several separate outbursts of gas-pyroclastic mixture into the atmosphere. The time of blast preparation is usually longer than the duration of a blast. Therefore the stages of blast preparation and development may be considered separately. In the case of volcanic blasts having a magmatic nature, the compressed gas energy is accumulated in the magma in the course of blast preparation. Gas in the bubbles distributed in highly viscous magma has an excess pressure. The various mechanisms of excess pressure formation are considered in the literature (Sparks 1978; Wilson 1980). We simply consider the magma body containing bubbles with gas pressure reaching tens or a few hundred of bars. Before a blast this body may be located inside the conduit or edifice. During the blast, magma body fragmentation occurs and the gas-pyroclast mixture is ejected into the atmosphere.

We adopt the following assumptions about the characteristics of the magma body before its fragmentation: the behavior of highly viscous interbubble partitions is the same as that of solid ones; all bubbles (pores) have the same spherical shape and size and are uniformly distributed in the magma; the gas pressure in the bubbles exceeds the atmospheric pressure. Then we can estimate the magma body potential energy released during the volcanic blast, which is taken to consist of elastic energy stored in the material forming interbubble partitions and compressed gas energy.

Given the above assumptions, the volume fraction of gas in the magma is ε where:

$$\varepsilon = \frac{V_g}{V_g + V_l} \tag{1}$$

and where V_g is the volume of gas, and V_l is the volume of magma. ε is a constant if gas bubbles have equal sizes and are uniformly distributed in the magma. In this case we can choose a unit cell of the system to be a single gas bubble with radius a, surrounded by an envelope with radius b. The envelope thickness ($l = b - a$) equals a half-thickness of the interbubble partition d (so that $d = 2l$). Then a and l may be related to ε by:

$$\frac{a}{l} = \frac{\varepsilon^{1/3}}{1 - \varepsilon^{1/3}} \tag{2}$$

For $\varepsilon = 0.5$, the ratio $a/l = 3.85$, and for $\varepsilon = 0.7$ $a/l = 7.93$. Hence $a/l > 1$. We designate the gas pressure in the bubbles as P_g and the pressure on the outer surface of the envelope as P_a. In the case of a thin envelope ($a \gg l$) the stresses inside it are:

$$\sigma_{\theta\theta} = \sigma_{\varphi\varphi} = \frac{P_g}{2}\,\frac{a}{l} \tag{3}$$

$$\overline{\sigma_{rr}} = \frac{P_g}{2} \tag{4}$$

where $\overline{\sigma_{rr}}$ is the average radial stress throughout the envelope thickness; and $\sigma_{\theta\theta}$ and $\sigma_{\varphi\varphi}$ are tangential stresses (Landau and Lifshits 1965). Equations 3 and 4 show that tangential stresses are predominant. Hence, $\sigma = P_g(a/2l)$ may be used. The elastic energy stored in the unit volume of solid material, q_{el}, is

$$q_{el} = \frac{\sigma^2(1-\nu^2)}{2E} \tag{5}$$

Here ν is the elastic Poisson's ratio and E is Young's modulus. Then for $\sigma = P_g(a/2l)$, the elastic energy stored in the envelope is:

$$E^s_{el} = \frac{(1-\nu^2)\pi}{2E}\,\frac{P_g a^4}{l} \tag{6}$$

The buble radius a is connected with ε as:

$$a \simeq \left[\frac{3\varepsilon}{4\pi\rho_l n_m(1-\varepsilon)}\right]^{1/3} \tag{7}$$

where ρ_l is the magma density (without bubbles), and n_m is the number of bubbles in the unit mass of this magma. n_m is connected with the number of bubbles in the unit volume of magma n_v as:

$$n_v = n_m \rho_l (1-\varepsilon) \tag{8}$$

Equation 6 can be rewritten taking into account Eqs. 2 and 7 as:

$$E^s_{el} = \frac{3(1-\nu^2)P_g^2}{8En_m\rho_l}\,\frac{\varepsilon^{4/3}}{(1-\varepsilon)(1-\varepsilon^{1/3})} \tag{9}$$

Hence the elastic energy of the unit volume of a porous solid magma body is:

$$E_{el} = E^s_{el}\, n_v \tag{10}$$

or:

$$E_{el} = \frac{\pi(1-\nu^2)}{2E}\,\frac{P_g^2 a^4 n_v}{l} \tag{11}$$

Then, using Eqs. 2, 7, and 8 we obtain:

$$E_{el} = \frac{3(1-\nu^2)}{8E}\,\frac{P_g^2\varepsilon^{4/3}}{(1-\varepsilon^{1/3})} \tag{12}$$

Hence, for $\nu = 0.2$, $E = 0.5$ MPa, $P_g = 20$ MPa, $\varepsilon = 0.5$ we have $E_{el} = 5.5 \cdot 10^3$ J/M^3.

On the other hand, the energy of a compresed gas stored in the unit volume of the porous magma body, E_g, is:

$$E_g = \frac{P_g\varepsilon}{\gamma - 1}\left[1 - (P_a/P_g)^{\frac{\gamma-1}{\gamma}}\right] \tag{13}$$

where γ is the isentropic exponent, and P_a is the atmospheric pressure. For $P_g \gg P_a$, the last equation can be written as:

$$E_g \simeq \frac{P_g \varepsilon}{\gamma - 1} \tag{14}$$

Hence, $E_g = 3.3 \cdot 10^7$ J/m^3 for $P_g = 20$ MPa, $\varepsilon = 0.5$, $\gamma = 1.3$. Comparison of this value with $E_{el} = 5.5 \cdot 10^3$ J/m^3 shows that $E_g \gg E_{el}$. This ratio is also valid for other values of P_g. Thus the compressed gas energy is the major one.

The strength of interbubble partitions material is an important parameter. Unfortunately, experimental data on the strength of near-solidus or solidified magma samples of small sizes (≤ 10 μm) are absent. The influence of temperature and crystals on the strength is unknown.

We estimated the pressure differential, ΔP, required for the solid interbubble partitions disruption. Assuming for simplicity the plain shape of partitions we have:

$$\Delta P = \frac{\sigma}{0.25} (d/a)^2 \tag{15}$$

where a is the linear size, and $d = 2l$ is the thickness. For ε_0, $a/l = 3.85$, and $a/d = 1.92$. Thus for $\varepsilon_0 \sim 0.5$, $\Delta P \sim \sigma$.

It should be noted that we have considered the static strength and the ultimate static pressure. For dynamic loads, σ_t can differ from the static one. For instance, McBirney and Murase (1970) considered the "short-term stresses" of magma, sufficient for the interbubble partitions to be disrupted, to be some bars.

The model for magma fragmentation

We consider that the energy of compressed gas contained in the porous magma body is released during volcanic blasts. In the case of highly viscous or solid interbubble partitions, the release of this energy occurs as a result of the magma body fragmentation. Consequently, the volcanic blast development is directly determined by the process of the fragmentation of the porous magma body existing in the volcanic edifice or conduit before the blast. In this case, the products of magma body fragmentation (pyroclastics) are ejected into the atmosphere. Unfortunately, the problem of fragmentation of porous bodies having excess pressure gas in their pores is poorly developed in mechanics. Apparently, the studies of Nikolski and Khristianovich are rare exceptions, and our work is based on their approach (Nikolski 1953; Khristianovich 1953, 1979).

We consider the porous magma body with solid (or very highly viscous) interbubble partitions to be inside the volcanic conduit (Fig. 1). We then assume that a rapid decompression is created over this body at some point, so that the interbubble partitions can be disrupted by the pressure differential between gas and atmosphere. Alternatively, this pressure differential can be caused by gas pressure increase inside the pores (bubbles). The force condition for the breaking of interbubble partitions in the upper layer of pores, bounding with the atmosphere, can be expressed as:

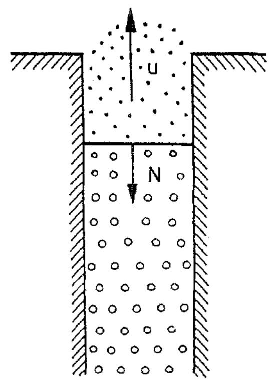

Fig. 1 A sketch of fragmentation wave propagation with velocity N inside a magma body and gas-pyroclast mixture ejection with velocity u into the atmosphere

$$\varepsilon(P_0 - P_a) \geq \sigma_t \tag{16}$$

where P_0 is the gas pressure in pores, P_0 is the atmospheric pressure, and σ_t is the tensile strength. A similar condition was suggested by McBirney and Murase (1970).

After the breaking of interbubble partitions in the upper layer of the porous magma body, gas contained in pores begins to expand to lower (atmospheric) pressure, accelerating the fragments of broken interbubble partitions. When the pressure drop over the newly formed free surface has taken place, the pressure differential is applied to interbubble partitions in the next layer of pores. The new layer of pores disrupts and the process is repeated. Thus, a layer-by-layer breaking away is realised in the course of disruption of a porous magma body with solid interbubble partitions. It should be noted that the term "solid interbubble partitions" implies here that the mechanical behavior of the partition material under a rapid loading is like that of a solid. However, the same material could also be characterised as highly viscous under a slow loading. Hence, as a result of the successive breaking of partitions, the disruption surface will move downward into the magma body. The magma body will not be fragmented in front of this moving surface, called the "fragmentation wave front". Behind the fragmentation wave front the mixture of expanding gas and fragmentation products will flow out into the atmosphere (see Fig. 1).

We can derive the main conservation equations for the fragmentation wave front according to Khristianovich (1979), and so determine the velocities of the fragmentation wave and of the gas-pyroclastic mixture ejection. All considerations will be made for the steady-state case. We shall adopt the following limitations to the model: one-dimensional geometry; equal velocities of expanding gas and particles behind the fragmentation wave front; and lack of heat transfer between gas and particles and of gas diffusion from particles. We introduce the following designations: N – velocity of fragmentation wave front, u – velocity of gas-pyroclastic mixture behind the fragmentation wave. ρ_l – density of magma (ρ_l=const). ρ_g – density of gas. The subscriptions "0" and "a" refer to parameters before and behind the fragmentation wave front respectively. In a coordinate system travelling with the moving front of the fragmentation wave, the magma body will run into an observer with the velocity N, and the products of the magma body fragmentation will move away with the velocity $N+u$. Then an equation of mass conservation for the solid phase, assuming its incompressibility (ρ_l=const), can be written as:

$$N(1-\varepsilon_0)=(N+u)(1-\varepsilon_a) \tag{17}$$

The similar equation for the gaseous phase has the appearance:

$$\rho_{g0} N \varepsilon_0=(N+u)\,\rho_{ga}\varepsilon_a \tag{18}$$

The energy balance of the process of magma body fragmentation can be written as:

$$e_0=e_a+e_k+e_l \tag{19}$$

This equation implies that the internal energy e_0 of a unit mass of gas in the magma body is expended for the gas–pyroclastic mixture acceleration leading to the kinetic energy e_k and for the irreversible losses (e_l), including the friction, crushing, and disturbance formation losses. In Eq. 19 e_a is the internal energy of the unit mass of gas after the expansion. Then one can write the equation of energy conservation for the magma body volume $SN\Delta t$, which is increased to the volume $Su\Delta t$ after fragmentation. Here S is the conduit area (see Fig. 1) and Δt is a short time interval. Hence:

$$N\rho_{g0}\varepsilon_0\left[e_0(1-\xi)-e_a-\frac{u^2}{2}\right]=P_a\varepsilon_a u+N(1-\varepsilon_0)\,\rho_l\frac{u^2}{2} \tag{20}$$

Here terms containing u^2 account for the kinetic energy of fragmentation products, and the term containing $P_a\varepsilon_a u$ defines the expansion work against the outer pressure P_a. The crush loss and other irreversible losses are allowed for by the ratio ξ:

$$e_l=\xi e_0 \tag{21}$$

which occurs in Eq. 20. From Eqs. 17 and 18 we obtain:

$$\frac{(1-\varepsilon_a)}{\varepsilon_a\rho_{ga}}=\frac{(1-\varepsilon_0)}{\varepsilon_0\rho_{g0}}=A \tag{22}$$

Then:

$$\frac{\varepsilon_a}{\varepsilon_0}=\frac{\rho_{g0}/\rho_{ga}}{1+[(\rho_{g0}/\rho_{ga})^{-1}]\,\varepsilon_0} \tag{23}$$

From Eqs. 17 and 18 we can obtain the relationship between N and u:

$$u=N\varepsilon_0[(\rho_{g0}/\rho_{ga})^{-1}] \tag{24}$$

Substitution of Eqs. 22 and 24 into Eq. 20 gives the equation of energy conservation in the form:

$$\rho_{g0}\varepsilon_0\left[e_0(1-\xi)-e_a-\frac{u^2}{2}(1+\rho_l A)\right]=\varepsilon_a\varepsilon_0 P_a[(\rho_{g0}/\rho_{ga})^{-1}] \tag{25}$$

Thus we have obtained equations of energy (25) and mass (17 and 18) conservation. Assuming a perfect gas, so that $e=P/[\rho_g(\gamma-1)]+a$ const, these expressions give equations for the calculations of u and N:

$$u=\frac{u_0}{\sqrt{1+A\rho_l}}\sqrt{\frac{2}{\gamma(\gamma-1)}}\sqrt{1-\xi-B} \tag{26}$$

$$N=\frac{u}{\varepsilon_0[(\rho_{g0}/\rho_{ga})^{-1}]} \tag{27}$$

where

$$B=\frac{P_a}{P_0}\frac{\rho_{g0}}{\rho_{ga}}\{1+\varepsilon_0(\gamma-1)[\varepsilon_0+(\rho_{g0}/\rho_{ga})]^{-1}\} \tag{28}$$

$$u_0^2=\gamma R T_0=\gamma\frac{P_0}{\rho_{g0}} \tag{29}$$

The momentum conservation equation is needed to complete Eqs. 26 and 27. However, a problem exists in deriving such an equation (Khristianovich 1979). The transfer of momentum in the cross-section S for the time interval Δt depends on the process of gas liberation from disrupted pores and is not known. Therefore, it has been suggested that Eqs. 26 and 27 should be completed by the relationship between P_a/P_0 and ρ_{ga}/ρ_{g0} for the fragmentation wave front in the form:

$$P_a/P_0=(\rho_{ga}/\rho_{g0})^n \tag{30}$$

Then Eqs. 26 and 27 can be rewritten to express the dependence of u and N on the ratio P_0/P_a at the fragmentation wave front:

$$u=\sqrt{\frac{2P_0\varepsilon_0}{\rho_m^0(\gamma-1)}}\cdot\sqrt{1-\xi-[P_0/P_a]^{\frac{1-n}{n}}\left\{\frac{1+\varepsilon_0\gamma[(P_0/P_a)^{1/n}-1]}{1+\varepsilon_0[(P_0/P_a)^{1/n}-1]}\right\}} \tag{31}$$

$$N=\frac{u}{\varepsilon_0[(P_0/P_a)^{1/n}-1]} \tag{32}$$

where:

$$\rho_m^0=\rho_{g0}\varepsilon_0+\rho_l(1-\varepsilon_0) \tag{33}$$

Here ρ_m^0 is the density of the porous magma body before fragmentation. If the weight portion of gas (water vapor) is small ($\rho_{g0}\varepsilon_0\ll\rho_l(1-\varepsilon_0)$), then the following equation may be used instead of Eq. 33:

$$\rho_m^0\simeq\rho_l(1-\varepsilon_0) \tag{34}$$

It should be noted that when $P_0/P_a\to\infty$, $n\to\gamma$. Hence, in this limiting case $n=\gamma$ can be used in calculations. The pressure P_a behind the fragmentation wave front may differ from the atmospheric pressure, and ξ is usually less than a few percent (Khristianovich 1979).

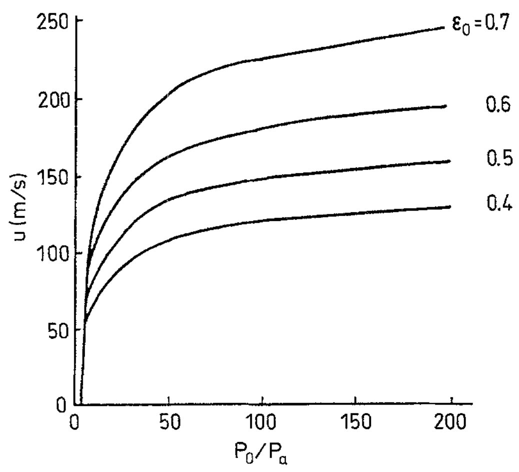

Fig. 2 Gas-pyroclast mixture ejection velocity u as a function of pressure ratio P_0/P_a in front of and behind the fragmentation wave front for initial magma porosity $\varepsilon_0=0.4$, 0.5, 0.6, 0.7; initial gas pressure $P_0=20$ MPa; magma density $\rho_l=2500$ kg/m^3; isentropic exponent $\gamma=1.333$, and ratio of crushing work and other losses to initial internal energy of gas $\xi=0.05$

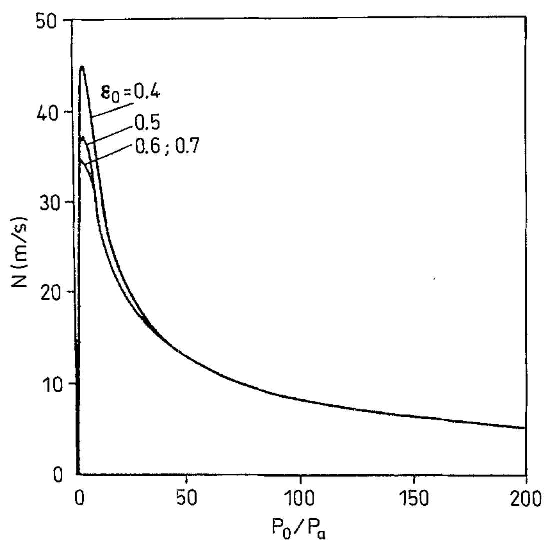

Fig. 3. The fragmentation wave front velocity N as a function of the pressure ratio P_0/P_a. Parameters used for calculations are the same as in Fig. 2

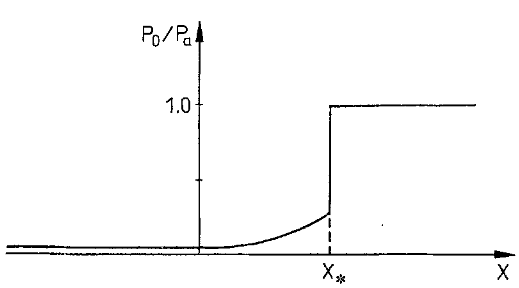

Fig. 4. The qualitative pattern of pressure ratio distribution in front of and behind the fragmentation wave front. x is the direction of fragmentation wave propagation; x_* is the location of the fragmentation wave front at some point

The results of calculations of the gas-pyroclastic mixture velocity as a function of the ratio P_0/P_a are presented in Fig. 2. One can see the continuous growth of u with the increase of P_0/P_a. The calculations of the fragmentation wave front velocity N as a function of P_0/P_a are shown in Fig. 3. There is a maximum of N in the region of low values P_0/P_a. For reasonable values of P_0/P_a, the value of N does not exceed 10 m/s. The pattern of pressure distribution in front of and behind the fragmentation wave front is shown qualitatively in Fig. 4.

If the solidified magma in the volcanic channel has a volume V, and the length $L=V/S$, where $S=\pi D^2/4$ is the channel area, then the duration of the porous magma body fragmentation t_f is:

$$t_f=L/N \tag{35}$$

This duration characterises the energy release rate during volcanic blasts. The compressed gas energy stored in the magma body with the volume V_m equals $E_0=V_m E_g$, where E_g is expressed by Eq. 13. Then the average power of this energy release may be expressed as:

$$Q=\frac{E_g V_m}{t_f} \tag{36}$$

The average mass flux during the blast is calculated as:

$$q_m=N\rho_l(1-\varepsilon_0) \tag{37}$$

Discussion

Some peculiarities of the approach used and the assumptions adopted will now be discussed.

One of the main problems is the creation of conditions sufficient for realisation of fragmentation. Equation 16 was suggested as the condition of fragmentation and volcanic blast onset. Two ways of accomplishing this condition are possible. First, continuous pressure increase inside pores (bubbles) in the upper layer of the magma body can lead to the disruption of interbubble partitions when ΔP is sufficient (Eq. 16). Second, a rapid external pressure decrease over the magma body can also increase the pressure differential applied to the interbubble partitions in the upper layer of the porous magma body. When Eq. 16 is satisfied, the fragmentation process begins. On the other hand, this condition (Eq. 16) defines the ceasing of the fragmentation process. Clearly, porosity decreases with increasing depth in the magma body and also the interbubble partitions may be "liquid" at depth so that pore expansion can precede the arrival of the fragmentation wave. As a result, the pressure differential decreases and the disruption process will not be brittle. During the fragmentation wave propagation in the porous magma body, excess pressure can arise behind the wave front. This may be caused by the sudden collapse of the walls of the conduit. The pressure disturbances produced will be propagated with a sound speed c in the medium behind the fragmentation wave front; they can reach this front if $c-u>N$. This situation is realised because of the low velocity of a fragmentation wave. In its turn, the pressure increase behind the fragmentation wave front can influence the wave motion through the change of the ratio P_0/P_a, which is a member of Eqs. 26 and 27. As a result of this influence, the velocities N and u can be changed. Also, the increase of P_a behind the fragmentation wave front leads to a decrease of the pressure differential $\Delta P = P_0 - P_a$. Therefore in accordance with Eq. 16 the fragmentation process can cease. Alternatively, when P_a decreases so that Eq. 16 is satisfied, the fragmentation process begins again.

Thus the process of magma body fragmentation may consists of several episodes, just as a volcanic blast may consist of several discrete explosions, which are observed during explosive volcanic eruptions. It should be noted that the velocities of the gas-pyroclast mixtures ejected into the atmosphere during subsequent explosions can be the same as during the first explosion. This fact is inadequately explained in models which consider the high-pressure gas volume do not allow for the process of fragmentation. In accordance with these models, the pressure inside the gas volume decreases as gas flows out, therefore the dusty gas velocity during subsequent ejections should be less. The multi-phase character of fragmentation during volcanic blasts explains the origin of secondary air pressure waves. The low velocity of fragmentation wave propagation explains the long duration of volcanic blasts. In our model this duration is $t_f \sim L/N$, but in the "gas pocket" model the characteristic duration of the blast should be $t \sim L/c$. It is clear that $t_f > t$, because $N < c$. t_f may increase due to the stopping and restarting of the fragmentation wave and to the additional time required for the gas-pyroclastic mixture to flow out after the ultimate fragmentation of the magma body.

This paper has considered the one-dimensional steady-state regime of fragmentation wave propagation. The results obtained show the dependence of u and N on P_0 and ε_0. Inasmuch as these parameters can change with depth, u and N also change. Data on the dependence of P_0, ε_0, and other parameters on depth, are needed for a complete description of the magma body fragmentation. An accurate value for ξ is also needed. It is known that in the case of coal and sandstone outbursts, ξ is not more than a few percent (Khristianovich 1979), but in the case of fragmentation of a solid or highly viscous magma body by a fragmentation wave this value is obscure. For instance, we could try to evaluate ξ, calculating the energy loss $\xi e_0 = S_s \lambda$, where λ is the specific energy of unit surface formation. However, values of λ are known only for the cases of room temperature and large size samples. λ must decrease with increasing T and decreasing l (Ivanov and Mineev 1979). A special investigation of near-solidus magma mechanical characteristics is needed.

The difference between gas and pyroclast velocities should be taken into account (Kurlaev 1980). We must also obtain the momentum conservation equation for the fragmentation wave front. Many problems arise in describing the unsteady outflow of gas-pyroclast mixture (Kieffer and Sturtevant 1984; Ivandaev 1984; Gelfand et al. 1985).

The assumption of solidified interbubble partitions narrow the range of phenomena which can be described. For instance, low-viscosity magma cannot be considered as brittle. Unfortunately, no theory of such pehnomena exists. However, the model suggested in this work seems useful for understanding volcanic blast mechanisms. The absence of a fragmentation theory for magma with gas bubbles seems to have led to the consideration of the stage of gas-pyroclast mixture outflow only (Eichelberger and Hayes 1982; Wohletz et al. 1984). It should be noted that McBirney and Murase (1970), and Bennett (1974) suggested that the process of the interbubble partition disruption could be of importance for the solution of the problem of pyroclast formation, but their ideas were not sufficiently developed.

Conclusions

A model for highly viscous magma fragmentation during volcanic blasts explains the observed values of ejecta velocities, the long blast duration, the possibility of several explosions during the blast, and the angular shape of pyroclasts. This model is based on the idea that a porous magma body can be fragmented by a fragmentation wave that arises when successive interbubble partitions break due to a pressure differential. The initiation and cessation of fragmentation wave propagation is determined by the relationship between porosity, the tensile strength of partitions, and the pressure differential between gas pressure in bubbles and the atmospheric pressure. Calculations show a low velocity (≤ 10 m/s) for fragmentation wave propagation. This essential fact determines the long blast duration, the low rate of energy release during volcanic blasts, and the possible influence of pressure change behind the fragmentation wave front on its motion and consequently on the fragmentation. The possibility that the fragmentation wave may temporarily stop explains the availability of several explosions during a blast. The model suggested shows the mechanism of angular pyroclast formation as a result of brittle disruption of highly viscous or nearly solidus interbubble partitions. For a given initial porosity and gas pressure of magma, ejecta velocities, duration, mass flux and some other parameters can be calculated. A model for magma fragmentation indicates that volcanic blasts have an "unloading" character, so that the compressed gas energy initially stored in gas bubbles inside highly viscous or near-solidus magma is only released during the blasts. This distinguishes volcanic blasts from chemical explosions. The model suggested may be applied to Vulcanian or Pelean blasts occurring on volcanoes having andesite or more acid low-temperature highly viscous magmas.

Acknowledgements. I thank V. N. Rodionov and S. A. Fedotov for encouragement and V. V. Adushkin, S. V. Kuznetzov, B. A. Ivanov and Yu. B. Slezin for their useful discussions, and the anonymous reviewer for his useful comments.

References

Bennett FD (1974) On volcanic ash formation. Am J Sci 274:648–661

Bogoyavlenskaya GE, Braitseva OA, Melekestsev IV, Kiriyanov VY, Miller CD (1985) Catastrophic eruptions of the directed-blast type at Mount St. Helens, Bezymianny and Shiveluch volcanoes. J Geodynam 3:189–218

Carey S, Sigurdsson H (1985) The May 18, 1980, eruption of Mount St. Helens. 2. Modelling of dynamics of the plinian phase. J Geophys Res 90:2948–2958

Eichelberger JC, Hayes DB (1982) Magmatic model for the Mount St. Helens blast of May 18, 1980. J Geophys Res 87:7727–7738

Gelfand BE, Gubanov AV, Medvedev SP, Timofeev EI, Tsyganov SA (1985) Shock waves during expansion of the compressed gas volume containing solid particles. Doklady AN SSSR 281:1113–1116 (in Russian)

Gorshkov GS, Bogoyavlenskaya GE (1965) The Bezymianny volcano and features of its last eruption. Nauka, Moscow, pp 1–172 (in Russian)

Ivandaev AI (1984) The flow with particles in the shock tube. Phys Goreniya Vzryva 20:105–111 (in Russian)

Ivanov AG, Mineev VN (1979) On the scale effects during disruptions. Phys Goreniya Vzryva 15:70–95 (in Russian)

Khristianovich SA (1953) On the outburst wave. On the crush wave. Izvest Acad Nauk SSSR, OTN 12:1679–1699 (in Russian)

Khristianovich SA (1979) Unrestricted flow of soil mass caused by the expansion of the high-pressure gas contained in pores. The crush wave. Preprint 128, Inst Probl Mechan AN SSSR, Moscow, pp 1–61 (in Russian)

Kieffer SW, Sturtevant B (1984) Laboratory studies of volcanic jets. J Geophys Res 89:8253–8268

Kovalev GN, Samoilenko BI, Storcheus AV (1980) On the nature of volcanic explosions of the Grate Tolbachik fissure eruption. Volcanol Seismol 3:62–71 (in Russian)

Kurlaev AR (1980) Estimation of the effect of degassing of the unrestricted flow of soil mass containing the compressed gas in pores. Preprint 163, Inst Probl Mechan AN SSSR, Moscow, pp 1–33 (in Russian)

Landau LD, Lifshits EM (1965) The theory of elasticity. Nauka, Moscow, pp 1–204 (in Russian)

McBirney AR, Murase T (1970) Factors governing the formation of pyroclastic rocks. Bull Volcanol 34:372–384

Nikolski AA (1953) On the disruption waves of aerated rocks. Doklady AN SSSR 91:1035–1038 (in Russian)

Self S, Wilson L, Nairn IA (1979) Vulcanian eruption mechanisms. Nature 277:440–443

Slezin YB (1979) The condition of the dispersive flow occurrence during an explosive volcanic activity. Volcanol Seismol 3:69–76 (in Russian)

Sparks RSJ (1978) The dynamics of bubble formation and growth in magmas: a review and analysis. J Volcanol Geotherm Res 3:1–37

Wilson L (1980) Relationships between pressure, volatile content and ejecta velocity in three types of eruptions. J Volcanol Geotherm Res 8:297–313

Wilson L, Sparks RSJ, Walker GPL (1980) Explosive volcanic eruptions. IV The control of magma properties and conduit geometry on eruption column behavior. Geophys J Roy Astron Soc 63:117–148

Wohletz KH (1983) Mechanics of hydrovolcanic pyroclast formation: grain size, scanning electron microscopy and experimental studies. J Volcanol Geotherm Res 17:31–63

Wohletz KH, McGetchin TR, Sandford II MT, Jones EM (1984) Hydrodynamic aspects of caldera-forming eruptions: numerical models. J Geophys Res 89:8269–8285

Editorial responsibility: C. Jeuport