JOURNAL OF GEOPHYSICAL RESEARCH, VOL. 104, NO. B5, PAGES 10,639-10,654, MAY 10,1999

Gas bubble dynamics model for shallow volcanic tremor at Stromboli

Maurizio Ripepe

Dipartimento di Scienze della Terra, Universita* di Firenze, Florence

Evgenii Gordeev

Geophysical Service, Russian Accademy of Science, Petropavlovsk-Kamchatsky, Russia

Abstract. Volcanic tremor at Stromboli (Aeolian islands, Italy) is correlated to small infrasonic transients *[Ripepe et al,* 1996] which repeat almost rythmically in time in a range between 0.8 and 1.2 s. We demonstrate that infrasonic transients are associated to small gas bubble (~0.5 m) burstings which produces no transients in the seismic signal. Tremor ground displacement attenuates with the inverse of the distance from the craters indicating that the source is shallow. Short-term energy release shows that infrasonic and seismic signals aire linked to the same dynamical process, while at the long-term scale it is evident that the two signals are controlled by two distinctive mechanisms. We suggest that the possible physical model acts in two steps: first, gas coalescence and, then, gas bursting. In our model, the seismic signal is related to the coalescence of a gas bubble from a layer of small bubbles, while the infrasonic signal is linked to the bursting of the bubble when it reaches the magma surface. Gas bubbles could form by free coalescence in magma or could be forced to coalesce by a structural barrier. We calculate that forced coalescence induces in magma a pressure change ($\sim 10^4$ Pa) 2 orders of magnitude higher than free coalescence, and it explains best the tremor ground displacement (10^{-5} m) recorded at Stromboli. Moreover, forced coalescence evidences the role of a structural barrier, such as a dike, in volcanic tremor source dynamics. In this gas dynamic process, the delay time of 1-2 s between infrasonic pulses could reflect the gas nucleation interval of basaltic magma *[Thomas et al,* 1993; *Manga,* 1996]. We propose that the source function for the shallow volcanic tremor at Stromboli could be the viscoelastic reaction of the magma to the pressure decrease induced by gas bubble growth rate under constant depressurization. The spectrum of our source function is controlled by the time duration of the pressure pulse, which represents the viscoelastic relaxation time of the magma and gas bubble growth rate. The predicted asymptotic decay of the frequency contents fits the spectral behavior of the vocanic tremor ground displacement recorded at Stromboli. We show that the same spectral behavior can be found in ground displacement spectra of volcanic tremor recorded on different volcanoes.

1. Introduction

The dynamics of explosive volcanism is still far from being understood. More than one model has been proposed to explain the mechanism that could generate the different kind of seismic signals recorded on volcanoes. The numerous models of the volcanic tremor source, from simple resonance of a magma body *[Kubotera,* 1974] to complicated systems consisting of a number of vibrating conduits *[Seidl et al,* 1981], have merely highlighted the complex nature of the volcanic tremor source. Nevertheless, in spite of the diversity of the source models proposed all the theories agree on the role of gas as a possible energy source for generating volcanic tremor. Many investigators [e.g., *Chouet,* 1985; *Schick,* 1988; *Okada et al,* 1990; *Ferrazzini et al,* 1991; *Ripepe,* 1996] believe that the main source of seismic energy has to be related to the degassing process of magma. Great attention has been focused on the role of gas dynamics to explain the explosive mechanism *[Sparks and Brazier*, 1982]. Magma fragmentation produced by instantaneous decompression induces a rapid gas nucleation which progressively accelerates during the explosion *[Toramaru,* 1989; *Mader*

Paper number 98JB02734.
0148-0227/99/98JB-02734$09.00

et ai, 1994; *Sugioka and Bursik,* 1995]. The amount of gas produced by decompression probably generates large bubbles which explode at the free magma surface *[Manga,* 1996; *Ripepe,* 1996; *Vergniolle and Brandeis,* 1996]. The main evidence for such an explosive mechanism at Stromboli comes from infrasonic signals *[Braun and Ripepe,* 1993; *Vergniolle and Brandeis,* 1994; *Buckingham and Garces,,* 1996; *Vergniolle and Brandeis,* 1996; *Vergniolle et ai,* 1996] recorded close to the active craters as simple pressure pulses *[Gordeev,* 1993; *Okada et ai,* 1990; *Braun and Ripepe,* 1993; *Vergniolle and Brandeis,* 1994]. Seismic transients and infrasonic waves are related at Stromboli to explosive volcanic activity, indicating that strong energy partitioning is acting at the source *[Ripepe,* 1996]. However small infrasonic pulses (0.2-1.3 Pa at 150 m from the crater) have been recorded at Stromboli with no associated volcanic explosion *[Ripepe et ai,* 1996]. We will demonstrate that each infrasonic pulse represents a single bursting of a small gas bubble. Therefore, we assume that each infrasonic pulse is representative of the pressure change in the magma produced by the gas coalescence. On bubble dynamical bases, we will develop a source model for gas bubble coalescence in magma induced by a structural barrier, like a dike, and we will suggest that seismic spectral content reflects the viscoelastic reaction of the pressure drop in magma induced by bubble growth under accelerating decompression. The spectrum of our theoretical function shows the same asymptotic decay as the volcanic tremor, while the local frequency peaks are explained as being due to the variable delay time between two successive infrasonic pulses.

2. Experimental Constraints for Source Modeling

Since 1993, the Department of Earth Sciences of the University of Florence has monitored the explosive activity of Stromboli volcano by a network of five Mark Products vertical components and one Mark Products three-component L-4C seismometers with a natural period of 1 s and sensitivity of $1 \text{ V cm}^{-1} \text{ s}^{-1}$. The seismic station FOS, at 150 m from the active vents, (Figure 1) was provided with an acoustic pressure sensor realized by Istituto Nazionale di Ottica in Florence (Italy). The pressure sensor consists of a Monacor condensator microphone MC2005 with an attenuation of-3 dB shifted back to 2 Hz and a sensitivity of 0.46 V Pa^{-1}. Data were transmitted via cables to the recording site located at Semaforo San Vincenzo in the Geophysical Laboratory of the University of Florence at 1800 m from the vents. The acquisition system was PC based and provided with a 16-channel National Instruments NB-MIO-16H A/D converter at 12 bits; data were stored at a sampling rate of 100 samples s^{-1} channel^{-1} *[Napoleone et ai,* 1993]. During the experiments, the network was synchronized to a Hitachi VM-S7200E video camera with an objective of 52 mm, a sensitivity of 10, and spectral resolutions between 0.4 and 0.8μm. Seismic and acoustic signals were syncronized to the videocamera with a Deutschland C-band Frankfurt (DCF) radio code. In the last years, this combined data analysis has allowed for the definition of some of the explosive dynamical features at Stromboli *[Blackburn et ai,* 1974; *Chouet et ai,* 1974; *Vergniolle and Brandeis,* 1994; *Ripepe,* 1996], which could be summarized as follows: (1) volcanic explosions are produced by large gas volume; (2) gas volume fluctuates during explosions with a period of ~2 s; (3) gas fluctuation during the explosion controls the seismic energy fluctuations; (4) seismic signal onset always starts 1-6 s before the explosion; and (5) infrasonic waves are generated by the explosion of a large gas bubble. However, we have also observed that moderate degassing generates low-pressure (~1 Pa at 150 m distance) impulses but does not produce clear transients in the seismic signals (Figure 2). Delay time distribution between two successive infrasonic pulses shows that they repeat in time at a rate of almost 1 s (Figure 3) in a narrow range between 0.8 and 1.2 s. These infrasonic pulses have always been recorded during volcanic tremor. In order to verify if infrasonic pulses are linked to volcanic tremor, we calculated the mean square amplitudes (MSA) of the infrasonic and seismic signal in a 30-s time window every 10 min. Transients related to volcanic explosions have been removed from seismic and infrasonic records before MSA functions were calculated. MSA fluctuates in time according to how energy is released in the atmosphere, as acoustic pressure, and in the ground, as seismic waves (Figure 4a). MSA functions of the five seismic stations are strongly correlated (0.96) to each other, which demonstrates that we are dealing with a coherent seismic signal produced by the same source. If we detrend the long-term variations out of the infrasonic and seismic MSA functions (Figure 4b), the superimposed energy fluctuations are in phase with a correlation coefficent of 0.8. However, correlation between seismic and infrasonic MSA functions in the long term reduces to 0.43, which indicates that infrasonic and volcanic tremor are not always closely in phase. If we carfully analyze Figure 5, we observe (e.g., from 15 to hour 22 and from hour 32 to hour 38) that infrasonic and seismic MSA sometimes show an inverse trend. This is clear evidence that the source of energy for the two signals is sensitive in the long term to different physical conditions. Consequently, we infer that infrasonic and seismic signals are somehow linked to the same dymamies process but controlled by two different sources. Ground displacement amplitude, calculated as the square root mean value of each MSA seismic function, attenuates with the inverse of the distance from the craters (Figure 6). This indicates that seismic source is shallow, and it confirms the presence of body waves in the volcanic tremor wave field *[Chouet et ai,* 1997]. Analysis of video camera im-

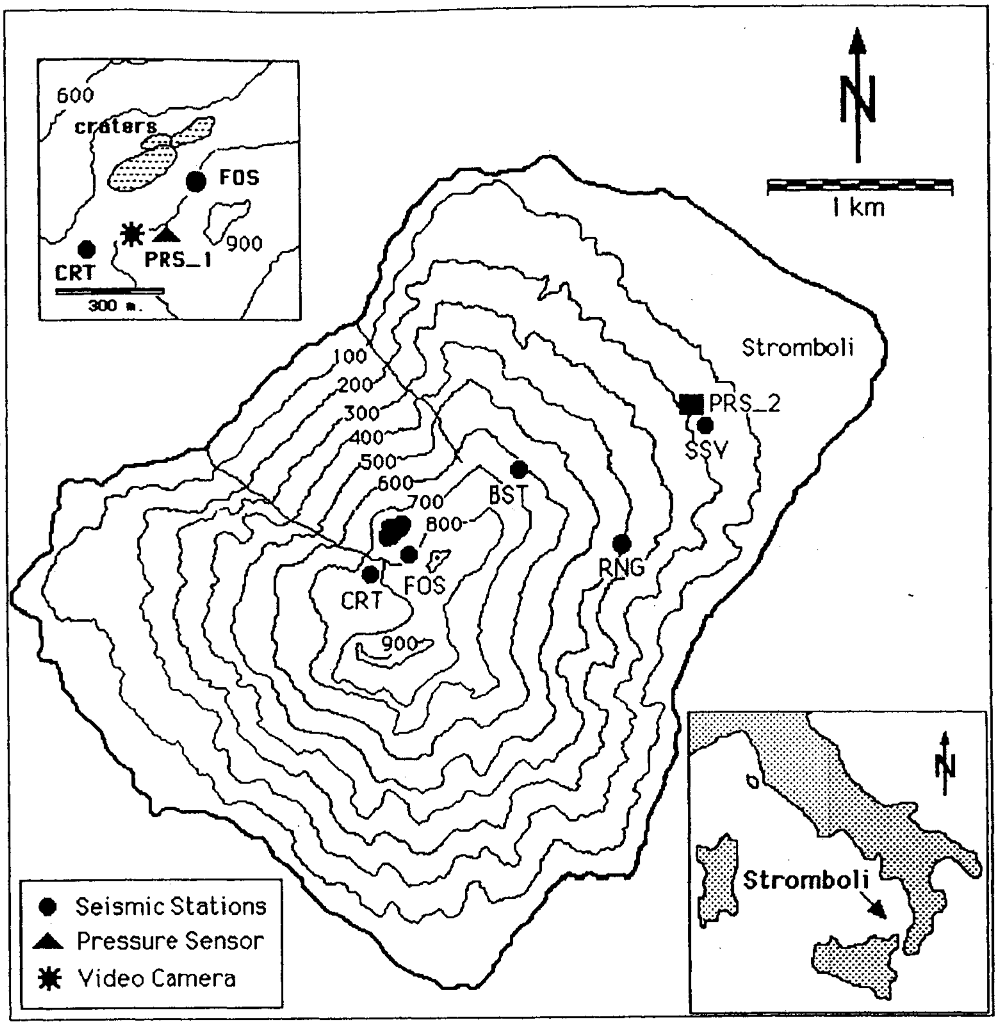

Figure 1. Map of Stromboli volcano with the location of the seismic stations, infrasonic sensors, and video camera.

ages reveals that infrasonic waves and volcanic tremor were both recorded during moderate degassing activity which we suggest as the possible dynamical process for both signals.

2.1. Gas Bubble Bursting as Source of Sound

We imagine the degassing process acting at Stromboli as the bursting of small gas bubbles at the top of the magmatic column. For large volcanic explosions, it has been suggested that infrasonic waves are generated by the vibration of a large gas bubble just before the explosion *[Vergniolle and Brandeis,* 1994; 1996]. If our small acoustic pulses are also generated by gas bubble burstings, acoustic pressure should be related to the frequency of the infrasonic signed according to *[Ltghthill,* 1978; *Lu et al.,* 1989; *Vergniolle and Brandeis*, 1994]

$$f = \frac{1}{2\pi}\sqrt{\frac{\gamma P_g}{\rho h R}} \qquad (1)$$

where P_g is the pressure inside the gas bubble, 7 is the ratio of the heat capacity (1.1 for hot gases), R is the radius of the gas bubble, p is the magma density and h is the thickness of the bubble film. Analysis of 400 infrasonic pulses at Stromboli indicates that, in logarithmic scale, amplitude and frequency of the pressure pulses are directly proportional by a factor of 2 (Figure 7) according to the linear relation

$$logf = logK + 0.5logP_g \qquad (2)$$

where K is equal to $1/2\pi\sqrt{\gamma/phR}$ and it is a constant term. This result points out that the radius of the gas bubble *(R)* should remain almost constant during degassing process and that frequency of the pressure pulses is mainly controlled by gas overpressure P_g. Bubble overpressure P_g can be directly estimated by the amplitude of infrasonic signals assuming that pressure decreases in the atmosphere by geometrical spreading ($\propto 1/r^2$). For example, 1 Pa recorded at a distance

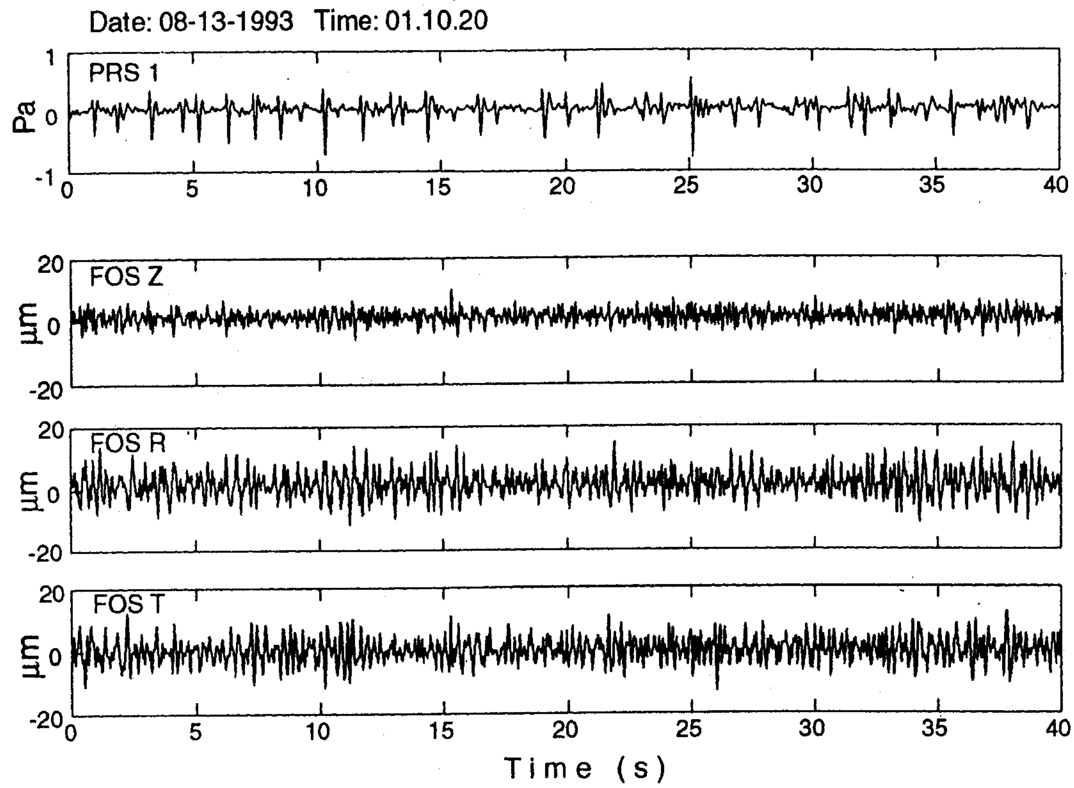

Figure 2. Infrasonic pulses and seismic signals recorded by the pressure sensor (PRS_1) and the three-component *(Z, R,* and *T)* station (FOS) located 150 m from the active vent. No evidence of transient signals is visible on the seismic signal.

r=150 m would correspond to a pressure at the source of 2.2x10^4 Pa. Considering a density of 2700 kg m^{-3} for the shoshonitic magma *[Francalanci et al,* 1989] of Stromboli and a thickness of the bubble film of 0.1 m *[Vergniolle et al,* 1996], we calculate from (2) that the mean radius of the bursting gas bubble is ~0.5 m. Therefore small infrasonic pulses (0.4-1.3 Pa) associated to volcanic tremor at Stromboli could be generated by small gas bubbles (~0.5 m) bursting at the surface of the magmatic column. We propose a model where infrasonic signal is related to the explosion of the gas bubble at the magma-air interface, whereas seismic waves are generated when the gas bubble forms in the magma. In section 3), we will consider the way gas bubbles of ~0.5-m radius could form, and we will try to quantify the pressure change in magma induced by this process.

3. Gas Coalescence Models

Gas dynamics is indicated as the main factor in explaining volcanic explosions. In the last years, many theoretical and experimental investigations have produced strong evidence for the role of gas bubble coalescence in volcanic eruption dynamics. Gas bubble nucleation and coalescence have been studied from a theoretical point of view *[Toramaru,* 1989; *Proussevitch et al,* 1993] and have been experimentally reproduced in laboratory simulations *[Mader et al.,* 1994; *Sugioka and Bursik,* 1995; *Zhang et ai,* 1997]. Experimental conditions require that gas starts to nucleate and to coalesce when the gas-saturated liquid almost instantaneously undergoes a decompression process. The sudden decompression creates the conditions which lead to

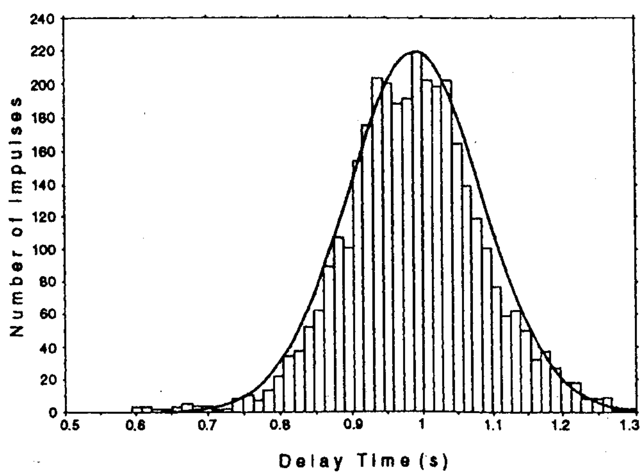

Figure 3. Frequency distribution of ~3500 delay times between infrasonic pulses. The data set has been smoothed with a five data point filter to reduce errors introduced by the threshold criteria used for identifying each single pulse. The delay times have a narrow distribution range of 0.8-1.2 s around a mean value of ~1s.

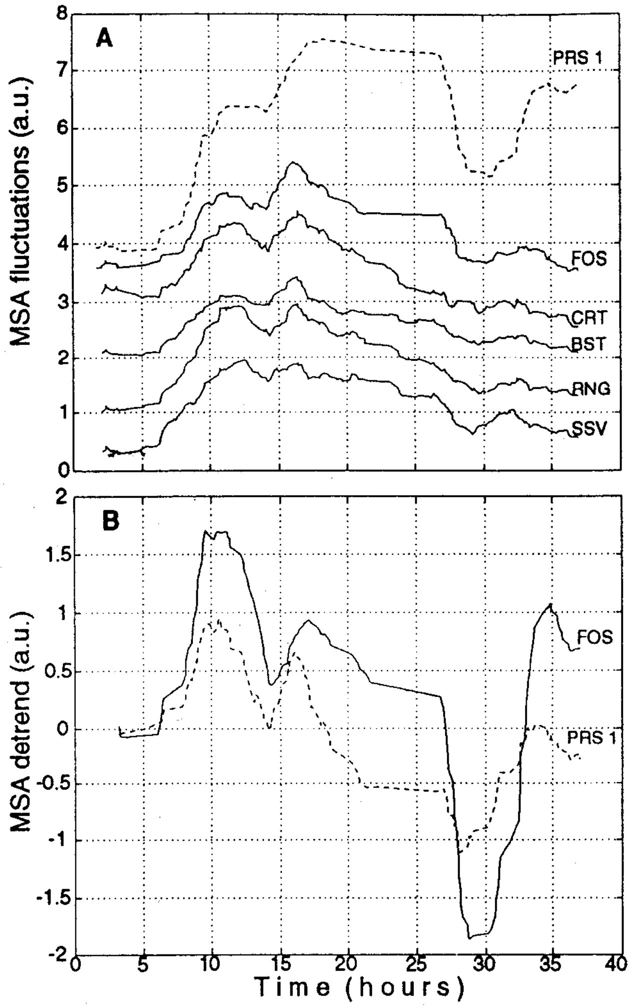

Figure 4. (a) Mean square amplitude (MSA) fluctuation of the infrasonic signals compared to the seismic MSA recorded at different distance from the craters (see map of Figure 1). MSA functions are proportional to energies and have been amplified by a factor of 2 (BST), 2.5 (RNG), and 2.7 (SSV) to improve data representation. (b) Infrasonic and seismic MSA functions have been detrended in order to filter out the long-term variations. Short-term fluctuations are closely in phase, indicating that the two signals are influenced by the same dynamical process.

the liquid being supersaturated and the gas to expand quickly. In our case, gats bubbles should form with a rate close to 1 s, which corresponds to the delay times of the infrasonic pulses. This time is far too short to allow the system to accumulate enough pressure to build up an overpressurized condition. If the volcanic tremor is generated by the formation of gas bubbles, the following mechanisms should be invoked. We explore here the possibility that volcanic tremor is produced by coalescence of a cloud of small bubbles, and we take into account two different models: free and forced coalescence. For both models we will try to estimate the amount of pressure involved in the two processes.

3.1. Free Coalescence Model

Gas bubbles start to nucleate and to grow freely in the magma because the hydrostatic pressure of magma progressively decreases toward the surface. Pressure inside the bubble is due to

$$P_g = P_{atm} + \rho g h + \frac{2\sigma}{R} \tag{3}$$

where a is the surface tension, R is the bubble radius, g is the gravity acceleration, p is the magma density, and h is the column of magma above the bubble. There is no gas coalescence until a close packing of bubbles is formed. It has been demonstrated [*Vergniolle and Jaupart*, 1986] that in magma the minimum gas volume fraction required to lead to coalescence is -70%. This percentage of gas volume fraction is reached at a depth of a few tens of meters in the magma column. Considering the density of the lava (p_l=2700 kg m^{-3}) during effusive activity at Stromboli and the density of the pyroclastic ejecta (p_e=1500 Kg m^{-3}) *[Chouet et al,* 1974], it can be easy calculated from

$$n = \frac{\rho_l - \rho_e}{\rho_l} \tag{4}$$

that gas represents (n) the 0.44 fraction of the magma volume during the explosions at Stromboli. In order to verify whether the free coalescence process could represent the source of the volcanic tremor, we tried to roughly estimate the pressure drop induced in the magma by a sudden formation of a 1-m-size gas bub-

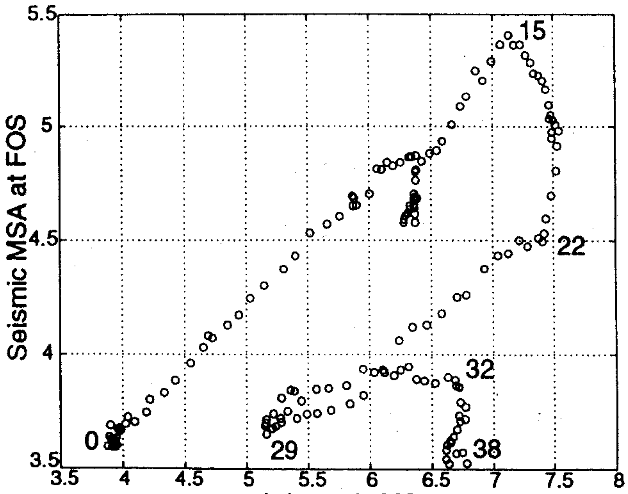

Figure 5. Infrasonic (PRS-1) MSA are represented as function of the seismic (FOS) MSA recorded near the craters. From 0 to 15 hours, infrasonic and seismic MSA are directly correlated, while the correlation shows an inverse trend from hour 15 to hour 22, and from hour 32 to hour 38. Instability of the long-term trend could be representative of a physical change (e.g., gas overpressure, magma column, or coalescence level in magma) in the dynamical process which has a direct influence only on one of the two parameters.

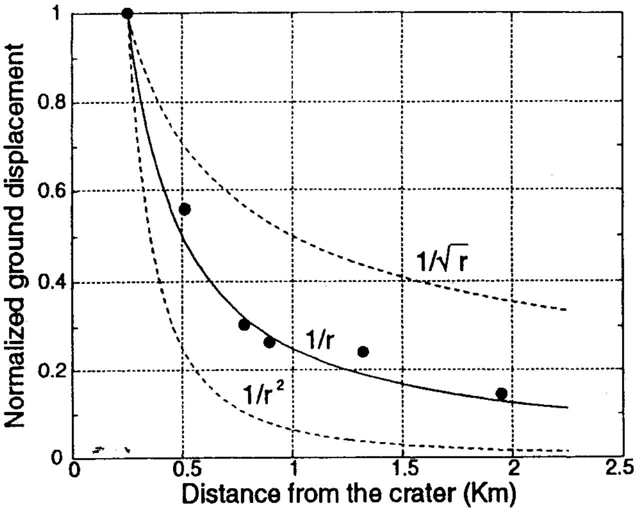

Figure 6. Attenuation law for the tremor ground displacement calculated as the square root of the MSA mean value of Figure 4a, The normalized ground diplacement attenuates following the function proportional to $1/r$. This indicates that the tremor source is shallow and that body waves are controlling the amplitude of the volcanic tremor wave field.

ble. When gas starts to nucleate in magma, bubbles with a 10^{-3}-m radius are formed [e.g., *Sparks,* 1978; *Toramaru,* 1989]. As soon as magma starts to rise inside the conduit, the pressure linearly decreases with depth; this process together with gas diffusion controls bubble growth in magma. If bubbles nucleate under a high-pressure regime, the growth rate will be slow, while if they nucleate at low pressure, the growth rate will be fast. In this way, the bubble size will tend to be uniform [*Scriven,* 1959; *Sparks,* 1978]. For basaltic magma, *Toramaru* [1989] calculated that the mean bubble radius is ~1 cm. Assuming that a 1-m-size gas bubble is instantaneously formed by the free coalescence process (Figure 8a), we can calculate the pressure drop P_w in the gas bubble as

$$P_w = \frac{2\sigma}{R_n} - \frac{2\sigma}{R} \tag{5}$$

where R is the initial radius of the bubble, while Rn is the radius of the final bubble, which can be considered as proportional to

$$R_n \approx \sqrt[3]{N}R \tag{6}$$

where N is the number of small bubbles with mean radius R that are coalescing in a larger bubble of radius R_n. Substituting (6) into (5), we obtain

$$P_w \approx \frac{2\sigma}{R}\left(\frac{1}{\sqrt[3]{N}} - 1\right) \tag{7}$$

which represents the pressure variation produced by free gas coalescence in a magma-gas medium. We estimate that if 10^6 bubbles of 1-cm size *(R)* will freely coalesce into a bubble with ~0.5 m radius, the pressure of the gas will drop to 80 Pa. The free coalescence can induce pressure changes of no more than $8x10^2$ Pa if we consider bubble size as small as 1 mm.

3.2. Forced Coalescence Model

Large gas bubbles are freely formed when the gas rising in the conduit approaches the surface. Bubble growth rapidly increases mainly under the decompression control *[Sparks,* 1978; *Barclay et al,* 1995]. It has been experimentally demonstrated *[Jaupart and Vergniolle,* 1988] that if the gas has the possibility of accumulating at a barrier, a large bubble can also form at depth. In their experiment, *Jaupart and Vergniolle* [1988] showed how different magma viscosity could drastically change gas flow regime. We considered that such mechanism could be active also at shallow depth, where we imagine the gas could accumulate below a structural barrier (e.g., a dike) along the conduit (Figure 8b). When gas reaches the critical volume of ~70%, a large bubble starts rising in the conduit forcing the magma to flow downward around the bubble to free space for the gas. The hydrostatic pressure will drop of a quantity equivalent to the bubble size

$$P_h \approx -2\rho g R \tag{8}$$

Assuming that the bubble has a spherical shape with radius R of 0.5 m, the drop of the hydrostatic pressure will be of $2.2x10^4$ Pa. The upward gas flow generates a force in the magma which is directed vertically downward *[Kanamori et al,* 1984]. Laboratory

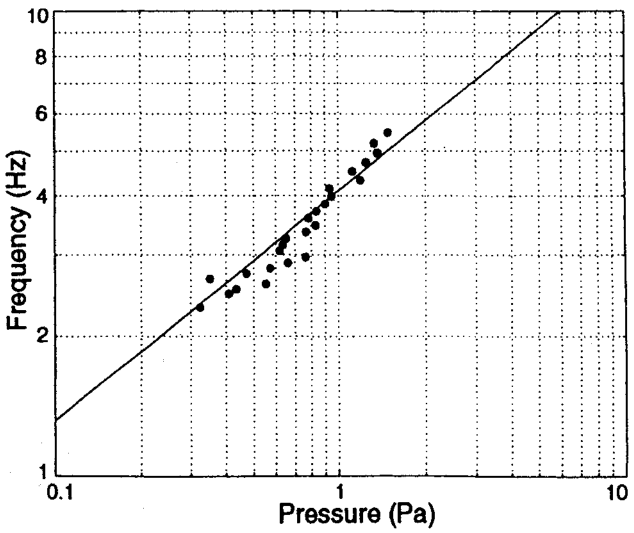

Figure 7. Logarithmic relationship between pressure and frequency content calculated for 400 infrasonic pulses recorded at 150 m from the vent. Pressure ranges between 0.3 and 1.2 Pa, while frequency ranges between 2.3 and 5.7 Hz. Frequency of the infrasonic pulse is linearly linked by a factor of 2 to pressure as predicted by the bubbling origin of the infrasonic waves (equation (2)).

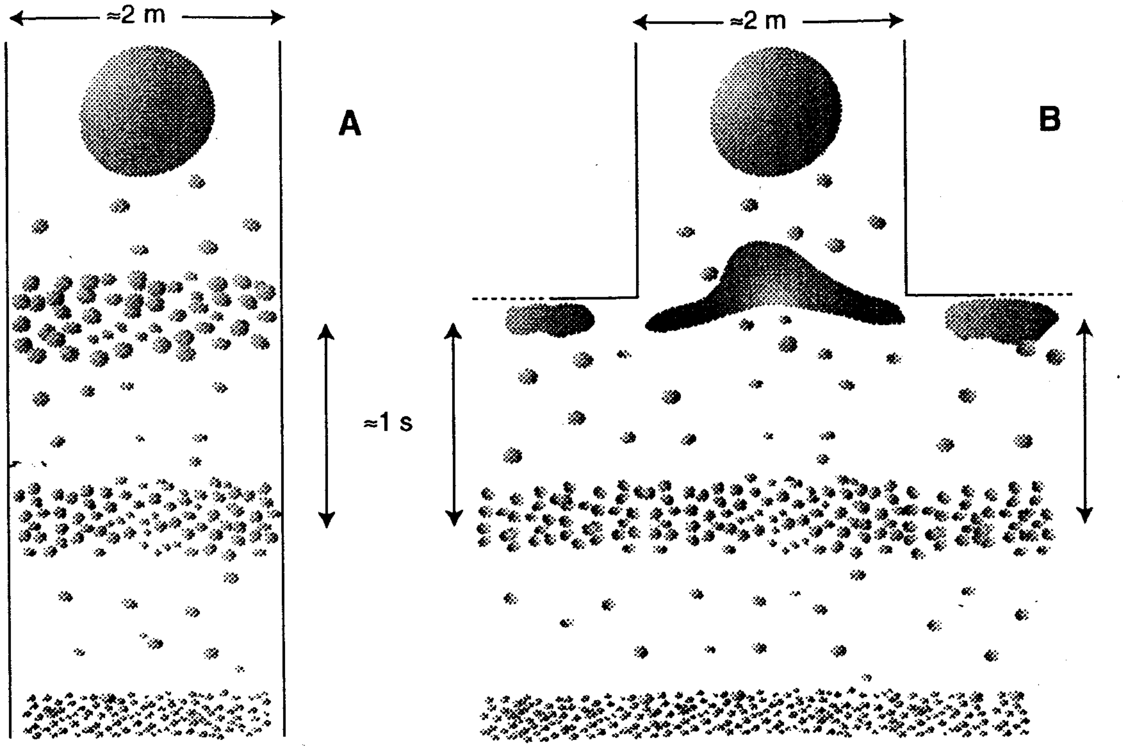

Figure 8. Schematic illustration of the two coalescence models considered, (a) Free coalescence model: bubble layers rising along the conduit freely coalesce in a larger bubble when hydrostatic pressure drops below a critical value, (b) Forced coalescence model: a layer of bubbles reaching a structural barrier is forced to coalesce and starts flowing in the above conduit inducing a hydrostatic pressure drop. We estimated that the free coalescence model releases pressure 2 orders of magnitude less than the forced one.

experiments demonstrated *[Delia Schiava et al.,* 1996] that this downward force generates oscillations in the magma, and it stops acting when the gas bubble stops flowing in the conduit. A single force, which represents a mass movement, and a seismic moment tensor, corresponding to a volume change, was proposed by *Uhira and Takeo* [1994] as a source time function of explosive eruption of Sakurajima. Ground displacement и induced by an isotropic source of radius R can be calculated in the far field at a distance $r >> R$ to the source as *[Kanamori et al,* 1984]

$$u = \frac{2\pi R^2 P_h}{3\rho r \alpha^3 T} \tag{9}$$

where p is the density of the medium, a is the P wave velocity and T is the wave period. Assuming density of 2300 kg m^{-3} for the ground and P wave velocity of 1600 m s^{-1} *[Braun and Ripepe,* 1993] at a period of 0.5 s, the pressure variation Ph of 2.2×10^4 Pa, induced in magma by forced coalescence, produces at a distance of 150 m a ground displacement of 1.4×10^{-5} m. This displacement is of the same order of magnitude as the recorded ground displacement at Stromboli (Figure 2). In terms of pressure release, a forced coalescence model best explains the tremor ground displacement recorded at Stromboli, and it evidences the role of a structural barrier in the volcanic tremor source dynamics.

4. Theoretical Time Source Function

Pressure time history under forced coalescence will be controlled by the bubble radius growth under continuous depressurization. The time of bubble growth under these conditions will be proportional to $t^{1/2}$ [*Toramaru,* 1989; *Mader et al,* 1994; *Zhang et al,* 1997] and larger than the value $t^{1/2}$ calculated for diffusional growth under constant pressure and temperature *[Toramaru,* 1989]. At these strain rates, magma behaves as a viscoelastic medium with a characteristic relaxation time which depends on viscosity and elastic modulus *[Mader et al,* 1994]. Following the theory of the linear viscoelasticity *[Bland,* 1960], we can understand the magma around the gas bubble as a viscoelastic media formed by the combination of N elements with the same viscous and elastic properties. During the forced coalescence process, the volume of the bubble increases and magma pressure decreases. This produces two different reactions in the medium in terms of magma deformation and decompression. If we consider an isotropic viscoelastic magma where the bubble grows almost instantaneously, a constant strain (Maxwell model) will generate a pressure which will decrease in time according to

$$P(t) = H(t) N P_h e^{-bt}, \qquad t > 0 \tag{10}$$

where P_h is the forced coalescence pressure and b represents the mean relaxation time for the N elements. The relaxation time b is defined by the ratio K/μ, between the bulk modulus K and the viscosity p. of the magma. However, under a constant stress (Kelvin-Voigt model), gas bubble volume grows as $t^{1/2}$ or $t^{2/3}$ depending on magma viscosity. This behavior could be explained by the combination of the Maxwell and Kelvin-Voigt models, where the stress and strain are changed simultaneously *[Bland,* 1960]. In this case, the pressure variation can be written as

$$P(t) = H(t) P_h t^m e^{-bt}, \qquad t > 0 \tag{11}$$

where the first part of this source time function

$$P(t) = H(t) P_h t^m, \qquad t > 0 \tag{12}$$

represents the rate m of pressure change in the magma. Laboratory experiments and theoretical calculations showed that during the degassing process under sudden decompression *[Zhang et al,* 1997] or at constant decompression rate *[Toramaru,* 1989] bubble radius grows as
In homogeneous viscoelastic media under fast process, the stress and the strain recover simultaneously, and the resulting relaxation time b is the characteristic viscoelastic parameter of the media defined as the ratio between bulk modulus K and viscosity μ of the media. In order to estimate a value for the relaxation time as close as possible to reality, we would consider the magma as a two-phase medium where the presence of gets drastically changes the physical parameters of the liquid.

4.1. Bulk Modulus of the Magma-Gas Medium

The elastic modulus of a liquid-gas mixture is the bulk modulus (K) or coefficient of compressibility a, where $(K=l/a)$. The bulk modulus in closed liquid-gas system depends on free gas content and decreases when gas content increases. For a gas volume fraction n, the bulk modulus of a two-phase gas-liquid system can be estimated as

$$K \approx \frac{K_g K_l}{n K_l + (1-n) K_g} \tag{13}$$

where K_l and K_g are the bulk modulus of the liquid and gas, respectively. The bulk modulus of a basaltic magma (K_g) is generally estimated as $\sim 10^{10}$ Pa *[Chouet,* 1985], while that of the gas can be calculated as

$$K_g = \gamma P \tag{14}$$

where y is the ratio of the heat capacity (1.1 for hot gases) and P is the gas pressure in magma which is mainly hydrosatic pressure. Assuming that the bubble forms at a few tens of meters, the bulk modulus (K_g) of the gas will be $\sim 10^6$ Pa of the same order as the hydrostatic pressure in magma. Considering that we calculated (equation (4)) a minimum gas volume content of n=0.44, the bulk modulus of the magma-gas medium at Stromboli can not exceed 2.3 K_g and it will be of the same order as that of free gas (K_{m+g} is $\sim 10^6$ Pa).

4.2. Viscosity of the Magma-Gas Medium

The viscosity (μ) of the basaltic melting at Stromboli varies from 10^2 Pa s to 10^3 Pa s *[Vergniolle and Jaupart,* 1986], but the effective viscosity (μ_{ef})for the fluid containing suspended small gas.bubbles is greater than the viscosity of the ambient fluid and can be expressed as

$$\mu_{ef} = c\mu \tag{15}$$

where c is the coefficient which depends on gas bubbles concentration. For a liquid containing a large number of suspended small bubbles with equal size, the coefficient c is equal to 4/3a, where a, the volume concentration of gas bubbles suspended in the liquid, is given by

$$a = \frac{1}{V} \sum V_i \tag{16}$$

where V_i is the volume of the single *ith* bubble and V is the total volume of the liquid *[Batchelor,* 1967]. From (15), it can be seen that for concentrations of gas bubbles as low as 1% or 3% the coefficient c changes from 100 to 40, respectively. Assuming for Stromboli a mean magma viscosity of 10^3 Pa s *[Vergniolle and Jaupart,* 1986], the effective viscosity can increase up to 10^5 Pa s with only 3% of small gas bubbles suspended in the magma.

4.3. Ground Displacement Induced by Pressure Variation in Magma

Considering the influence of the gas on the magma physical properties we can write the general time source function for the pressure variation induced in the magma by gas bubble growth under forced coalescence as

$$P(t) = P_h t^m \exp\left(-\frac{K}{\mu_{ef}} t\right) H(t) \tag{17}$$

where K and μ_{ef} are bubbly magma bulk modulus and effective viscosity, respectively, and P_h is the maximum pressure change induced in the magma-gas medium by gas bubble forced coalescence. Forced coalescence process generates in the medium a pressure drop (equation (8)) which is 2 orders of magnitude higher than what has been calculated (equation (7)) for the free gas coalescence model. Forced coalescence of a gas bubble with 0.5-m radius induces a pressure drop of $\sim 10^4$ Pa in the magma, which is consistent with ground displacement of $\sim 10^{-5}$ m measured for the volcanic tremor at Stromboli. In natural eruptions, bubbles should grow faster because of expansion and mass transfer effect, becouse of the interaction between bubble growth and ascent *[Zhang et al*, 1997]. It means that accelerating decompression is more suitable than constant decompression. For this reason, as the first approximation we found it more convenient to set m=1 in our fhodel, under the assumption that the bubble growth rate can be considered proportional to *t* (Figure 9a). At distances r larger than the bubble radius R, our spherical source can be considered as a point source *[Jiang et al*, 1994] which produces a symmetric displacement in the surrounding solid half-space proportional to the first derivative of the pressure of the point source *[Aki and Richards*, 1980]. Therefore the pressure source function produced by the forced gas coalescence process of (17) should be derivated in order

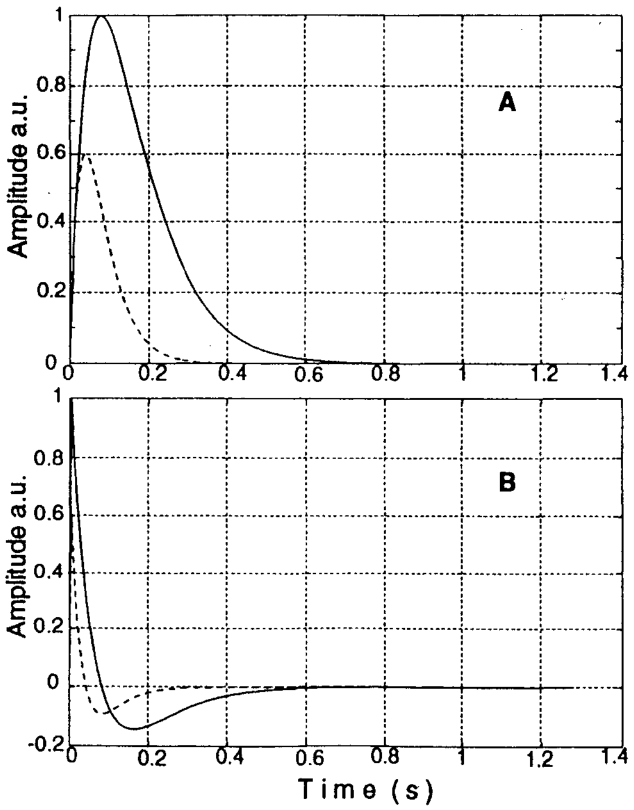

Figure 9. (a)Theoretical source function calculated from equation (9) for different time relaxation (curve 1) b=0.08 and curve 2) b=0.04) and same bubble growth parameter m=1. (b) First derivative (equation (18)) of the two pressure functions, assumed to be representative of ground displacement induced by the bubble growth in magma.

to be compared to the ground displacement produced by volcanic tremor

$$\left(1 \quad \frac{K}{\mu_{ef}} \right) \quad \left(\frac{K}{\mu_{ef}} \right) \tag{18}$$

where *S(t)* is the first derivative (Figure 9b) of the pressure function *P(t)*.

4.4. Spectral Characteristic of the Source Function

Assuming m=1, the cosine Fourier transform of the theoretical pressure time function for the forced coalescence process from (17), is given as

$$P(\omega) = P_h \frac{(b^{-2} - \omega^2)}{(b^{-2} + \omega^2)^2} \tag{19}$$

while for the first derivative of (18) S(w) be calculated as

$$S(\omega) = P_h \frac{2\omega^2}{}$$

where w=2πf. It can be demonstrated that in the high-frequency side of the theoretical density spectrum of (19), the asymptote decays as af^{-2}, while the low-frequency side has a constant asymptote. The first derivative density spectrum from (20) shows the same asymptotic decrease af^{-2} in the high-frequency side, while in the low-frequency side it has the asymptote proportional to f^2 (Figure 10). The theoretical source spectrum has then been compared to the displacement

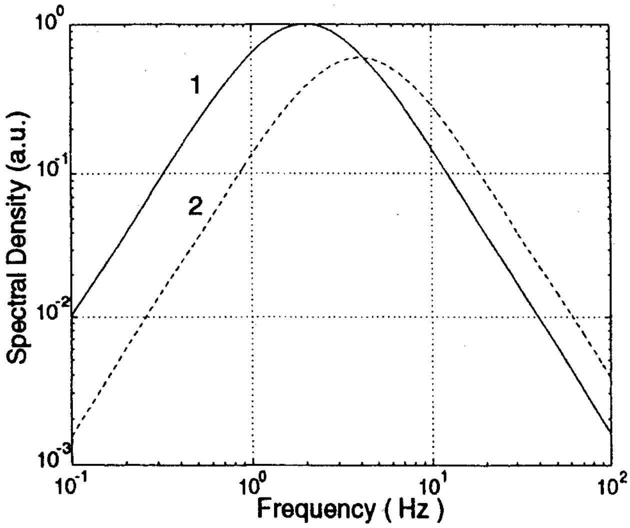

Figure 10. Theoretical spectra of ground displacement calculated for the two pressure functions of Figure 8b. The different frequency content depends on different *b* values: curve 1) b=0.08 and curve 2) b=0.04. Spectral amplitudes decay in the high- and low-frequency range as af^{-2}.

spectrum at Stromboli. Seismic records were first corrected for the instrumental response functions, and then the ground velocity was integrated. Spectra relative to ground displacement recorded at each of the five seismic stations of the Stromboli's network were calculated every hour for 2 days (Figure 11a). The 240 ground displacement spectra were stacked together to enhance spectral source characteristics and to reduce the path/site effects. This procedure allowed us to neglect the propagation effects of the medium in our theoretical treatment and to compare directly the tremor to our theoretical spectrum. The general trend of the stacked spectral functions does not depend on the transfer function (0.5-50 Hz) of the short-period

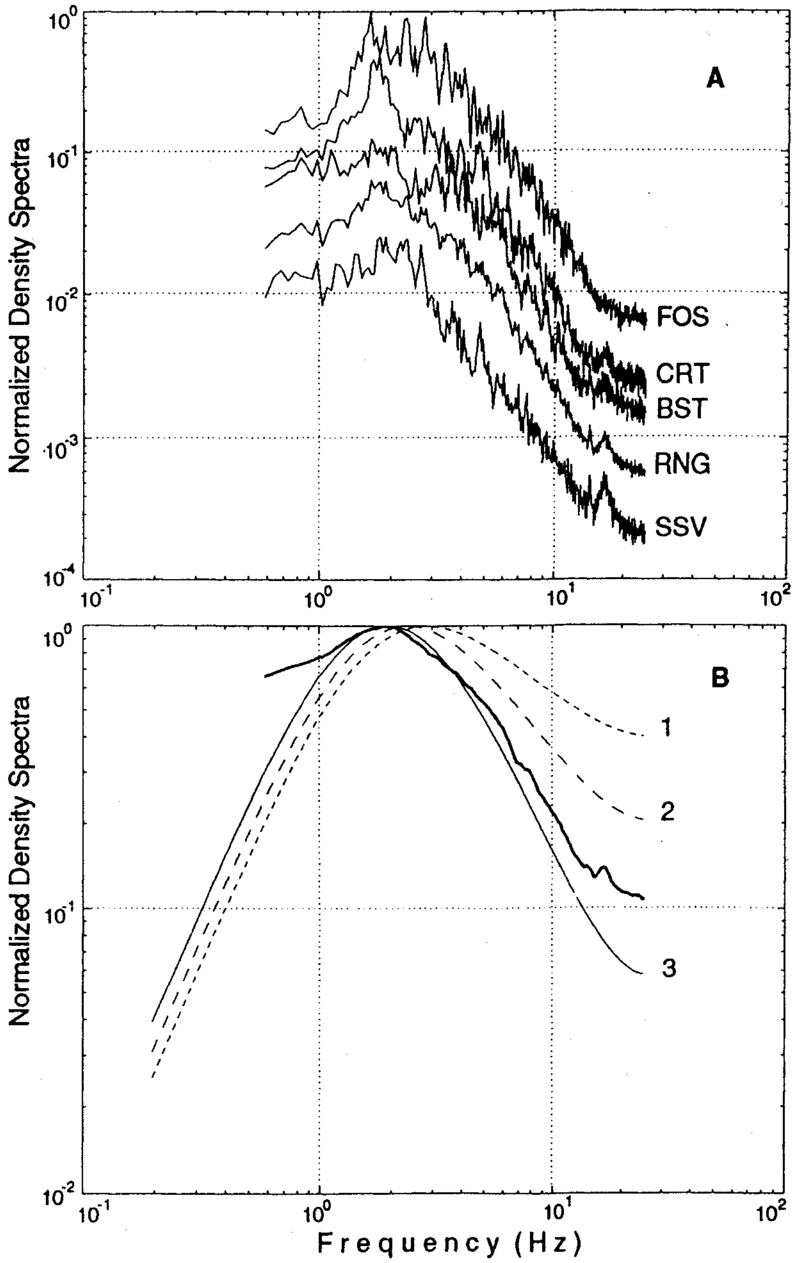

Figure 11. (a) Ground diplacement density spectra calculated for five seismic stations of the Stromboli's network. Each spectrum is the average of 48 spectra calculated every hour for 2 days. Spectral density has been normalized to different values, according to their distance from the craters in order to improve data representation, (b) Volcanic tremor ground displacement spectrum obtained by stacking the five spectra of Figure 9a. This mean spectrum is representative only for the tremor source spectral content. The spectrum is compared with three different theoretical source spectra calculated assuming the same relaxation time (b =0.076) and three different gas bubble growths in magma: curve 1) m=1/2, curve 2) m=2/3 and curve 3) m=1. Gas bubble growth under constant acceleration (m= 1) seems to be more suitable for volcanic tremor at Stromboli. The two asymptotes decay as af^2.following our theoretical assumptions and defining a "corner" frequency of 2.1 Hz.

(1 s) seismometers used. In fact, broadband experiments at Stromboli *[Neuberg et ai,* 1994] have revealed how even if volcanic activity irradiates seismic energy at frequency far below (~0.1 Hz) the typical frequency responce of short-period instruments (> 1 Hz), tremor shows small energy in the low-frequency band (< 1 Hz). Polarization analysis and wave propagation have revealed that the low-frequency band (0.1-0.4 Hz) of the tremor spectral content has no volcanic origin but is generated by ocean microseisms *[Braun et ai,* 1996]. Therefore we infer that ground displacement spectrum calculated as the average (Figure lib) of different spectra calculated from short-period recordings at stations with variable azimuth and distance to the source is a reliable representation of the tremor source. The averaged ground displacement spectrum shows the predicted asymptotic behavior we calculated assuming m=l (Figure lib). The intersection between the two asymptotes defines a frequency (F=l/27πb) which corresponds to the mean duration of the single pulse (Figure 9). According to our model, this frequency should depend on the viscoelastic properties of the magma. At Stromboli, it corresponds to ~2.1 Hz which is equivalent to 6=0.076. Considering that we have calculated the bulk modulus as $\sim 10^6$ Pa and the effective viscosity as $\sim 10^5$ Pa s, the *b* coefficent should be ~0.1, and consequently the predicted frequency for the volcanic tremor at Stromboli should be 1.6 Hz.

4.5. Wave of Bubbles and Source Time Rate

Volcanic tremor shows significant peaks in the spectrum which are consistent in time and space *[Schick,* 1988]. These spectral peaks have been often explained as source properties induced by resonating magma *[Aki et ai,* 1977; *Chouet,* 1985] or as path effects due to resonant scattering of media *[Gordeev,* 1993; *Correig and Vila,* 1993; *Kedar et ai,* 1996]. It is not our aim to discriminate between source and/or path effects in the tremor spectral contents at Stromboli. However, infrasonic records suggests that spectral peaks of volcanic tremor could be produced by a source which repeats in a quasi-stationary way in time. Infrasonic pulses are indicating that small gas bubbles are bursting at s rate which represents the source time signature and coincides with the formations rate of the gas bubbles in magma. Pressure pulses repeating at the same rate of s have been recorded inside the conduit of Old Faithful geyser by *Kedar et ai,* [1996]. The origin of the pressure pulses was explained as the collapse of small (~5 cm of radius) bubbles in the geyser hot water, while the delay time is produced by regular thermodymamic instability of the air-water two-phase system. In magma-gas systems, the pattern of the two-phase flow is controlled by the volume fraction of the gas phase. Bubble size distribution and coalescence process affect the hydrodynamics magma behavior which is closely related to the vesiculation process. Vesiculation can be characterized by several parameters, such as nucleation pressure, nucleation interval and total number of nucleated bubbles. *Toramaru* [1989] showed numerically that for basaltic magma rising with a velocity of 1 m s^{-1} , gas nucleation occurs at almost 2 km depth and with an interval of 2 m. Nucleation layers at 1-m interval had been, in fact, observed *[McMillan et ai,* 1987] in Columbia River basaltic lava flows. *Manga* [1996] showed how bubble interactions could lead to a gravitationally unstable bubble waves with 3 to 30 m wavelength in basaltic magma such as Stromboli. Wave instability could also be produced when gas bubbles rising in the magma are trapped beneath a high-viscosity contrast *[Thomas et ai,* 1993]. In this case, the wavelength of the bubble layers instability has been predicted to be in the range of 1 cm to 1 m *[Thomas et* al.,1993]. Assuming a magma velocity of cm s^{-1} *[Parfitt and Wilson,* 1995], layers of bubbles could rise with a minimum time interval of 1 s. This value is of the same order of magnitude of the mean delay time measured for the infrasonic pulses recorded at Stromboli. Hence we infer that infrasonic waves are produced by the degassing process which occurs at a rate proportional to the minimum wavelength of the bubble layers instability. We simulate the whole degassing process, taking into account also the way in which source time history repeats in time. This has been represented as a sequence of single pulses with a random distribution of amplitudes and phases. The general time function of the pressure fluctuations can be written as

$$P(t) = \Sigma_n P_n (t - \tau_n) \exp(-b(t - \tau_n)), \quad t \geq 0 \quad (21)$$

where P_n is the pressure changes induced in the magma by bubble growth, 6 is the magma-gas time relaxation, and r_n is the delay time of the *nth* single pulse. Random distribution of the source pulse amplitudes (pressure drop) and time delays r_n (nucleation interval) (Figure 12a) produces peaks randomly distributed in the spectrum (Figure 12c). If, however, the delay times follow a regularly equispaced distribution (Figure 12b), the resulting spectra will have stable spectral peaks (Figure 12d) typical of the harmonic tremors. Spectral peaks will be stable in time and space, but their origin will not be due to resonance of the source but to the regular occurrence of an impulsive source. However, the general shape of the spectra will be always the same because it is controlled by the time relaxation *b* of the magma-gas medium (equation (20)), which depends on the bulk modulus and viscosity ratio. Since the duration of the single pulse in our theoretical consideration depends mainly on the parameter of the fluid (equation (18)), we conclude that the general shape of the volcanic tremor spectra is determined by viscoelastic characteristic of the magma.

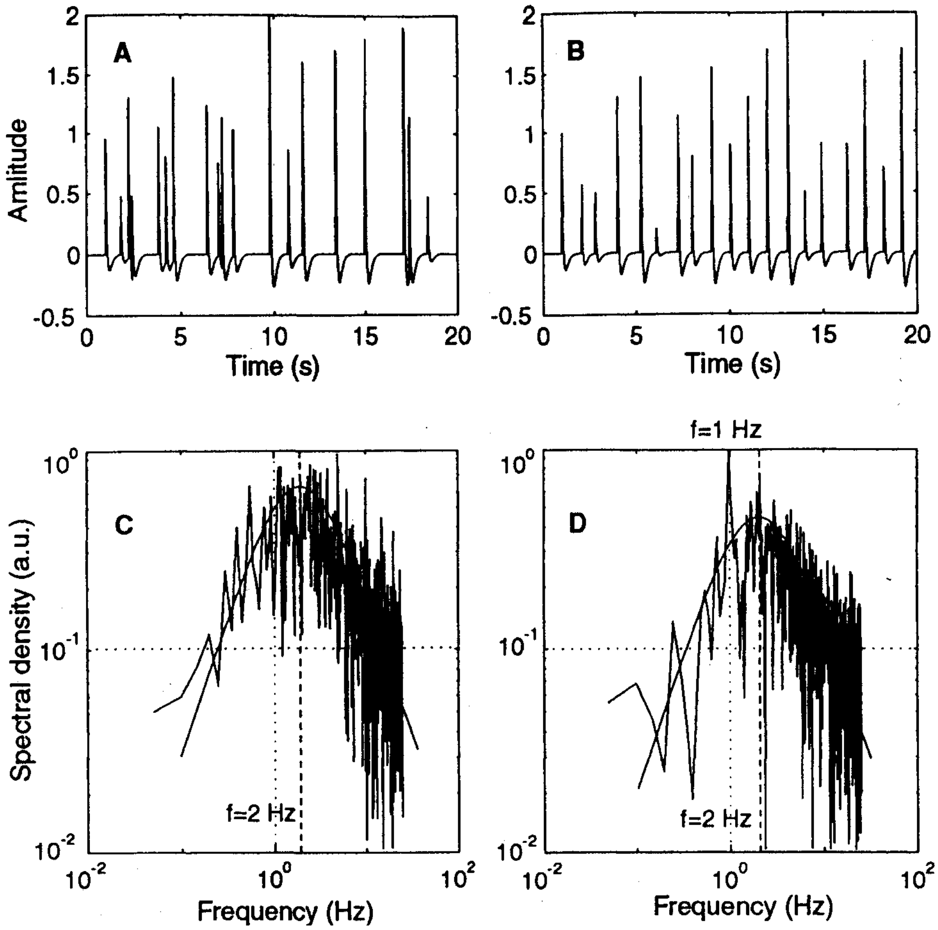

Figure 12. Source time history simulated as series of pressure drops (equation (19)) with (a) randomly distributed amplitudes and delay times and (b) normally distributed delay times around the mean value of 1 s. (c) and (d) the two spectra show the same asymptotic behavior and the same 2-Hz frequency at the asymptotes intersection. Distribution of peaks in the spectra reflects delay times between source events. This is more evident in Figure 12d where the strong peak at 1 Hz comes from the ~1-s delay time source distribution.

4.6. Comparison With Volcanic Tremor of Other Volcanoes

Tremor is a continuous signal common to all the active volcanoes and with the same spectral characteristic of low-frequency(1-3 Hz) content *[McNutt,* 1989]. Similar asymptotic behavior of the ground displacement spectrum has been found also for the seismic tremor recorded at different volcanoes (Figure 13). Therefore we infer that the same process as described for Stromboli is acting also on other volcanoes and that different viscoelastic magma parameters and bubble growth rate m are responsible for the different values of the "corner" frequency defined by the intersection of asymptotes. We cannot quantitatively demonstrate the relationship between volcanic tremor spectra and magma parameters because no accurate informations exists about the "in situ" characteristic of magma, unlike at Stromboli. However, a qualitative relationship seems to exist: Mount Hekla has very low viscosity magma (basaltic) and volcanic tremor spectrum shows a higher frequency (Figure 13c), while Mount Avaehinsky that has an high viscosity magma (andesitic) shows a lower frequency (Figure 13d). Our model emphasizes the role of gas dynamics and the importance of viscoelastic properties of magma compared to previous tremor models, and it does not need to include any geometrical constraint on the magmatic conduits to define the spectral characteristic of the source. Besides, rhythmicity of the infrasonic impulses recorded at Stromboli support a strong evidence of the instability rate of the two-phase system. This instability induces the source to repeat at a rate of s which could explain harmonic tremor recorded at several volcanoes. With this point of view, it appears as a strong coincidence that harmonic tremor at volcanoes with different magmatism such as Sakurajima, Japan *[Kamo et al,* 1977], Langila, New Guinea *[Mori et al,* 1989], Semeru, Indonesia *[Schlindwein et al,* 1995], and Arenal, Costa Rica *[Benoit and McNutt,* 1997] have harmonic spectral peaks around 1 s. According to our model, these harmonic frequencies should be related to a degassing process more regular than at Stromboli.

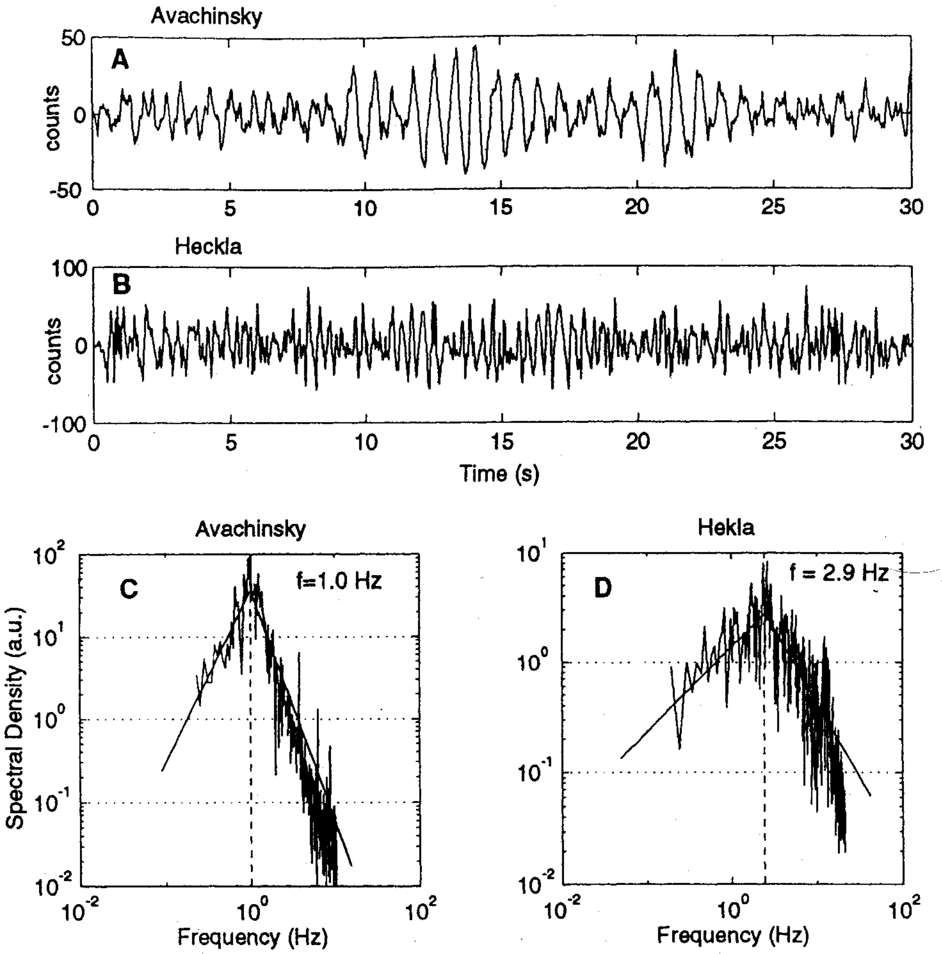

Figure 13. Example of volcanic tremor at (a) Mount Avachinsky (Kamchatka, 1991 eruption) and at (b) Mount Hekla (Iceland, 1991 eruption). Avachinsky is an andesitic volcano with a magma viscosity higher than Stromboli and as predicted shows (c) a "corner" frequency of Hz lower than at Stromboli. However, Hekla is a basaltic volcano with a magma viscosity lower than those of Avachinsky and Stromboli and has (d) a "corner" frequency (~4 Hz) higher than at Avachinsky and Stromboli.

5. Conclusions

We present a model for the volcanic tremor source at Stromboli which is strongly constrained by the recording of small infrasonic pulses. As previously inferred [*Ripepe et al.*, 1996], we demonstrated (equation (1)) that these small infrasonic pulses are generated by the bursting of small gas bubbles (~0.5 m) in one of the active vent. Long-term energy fluctuations have shown that volcanic tremor and infrasonic pulses are intimately linked to the same dynamical process. Seismic and infrasonic signals are coherent in time indicating that the two process are part of the same phenomenon which has two phases separated in time and space. We argue that the dynamical process responsible for the seismic signal is gas coalescence which produces the bubbles that reaching the magma surface explode and generate infrasonic pulses. We propose two different ways for the gas coalescence process to occur in magma: a free coalescence and a forced coalescence model. Both models could really be applicable, and are supported by numerical and experimental simulations [*Jaupart and Vergniolle,* 1988; *Mader et al,* 1994; *Manga*, 1996; *Sahagian et al,* 1989; *Zhang et al.,* 1997]. We tried to estimate the pressure change produced by the two coalescence models. Pressure produced by simple free coalescence (equation (7)) is more than 2 orders of magnitude smaller than pressure drop produced by the forced coalescence model (equation (8)). Gas could be forced to coalesce [*Jaupart and Vergniolle,* 1988] in the magma at a structural barrier such as a dike. When gas volume concentration reaches the critical value of 0.7, a large bubble will form and will start to flow in the conduit. Absolute pressure drop will be proportional to the bubble size (equation (8)), while the pressure time history in magma will depend on the bubble growth rate which we assumed as proportional to t. We propose that the volcanic tremor source function could be the viscoelastic reaction of the magma to the sudden pressure decrease induced by the gas bubble growth rate under constant depressurization (equation (18)) *[Mader et al,* 1994; *Zhang et al,* 1997]. A bubble in magma grows under two conditions: diffusion and decompression. When a gas bubble forms at depth, it slowly grows mainly by diffusion, and the size of the bubble remains as small as 1 cm. Only approaching the surface (100-200 m) does the decompression effect exceed the diffusion effect *[Sparks,* 1978], and the bubble reaches larger size. As a consequence, the source of the volcanic tremor should be shallow, and it should be confined in the last 100-200 m *[Chouet et al,* 1997]. Therefore the structural barrier, where small gas bubbles are forced to coalesce in a larger bubble, could be located between the surface and 200-m depth. Correlation between the seismic and infrasonic MSA functions (Figure 4b) indicates that the time elapsed during bubble growth and burst is very small, and it cannot exceed the MSA time window of 30 s. Considering that the gas bubble moves along the conduit with a velocity of ~2 m s^{-1} *[Batchelor,* 1967; *Sparks,* 1978; *Vergniolle and Jaupart,* 1986] we could calculate that the magma column above the coalescence level could be no more that 60 m. If we assume that a gas bubble is stable and it moves upward in the conduit with constant velocity, then no seismic signal will be irradiated by the bubble during its uprising. As a consequence, the magma column above the forced coalescence level will not contribute to tremor generation. This seems to be confirmed by laboratory experiments *[Delia Schiava et al,* 1996] where oscillations induced by gas bubble growth stop as soon as the bubble moves in the liquid at constant velocity. However, we cannot exclude that length of magmatic column or geometry of the structural barrier does not contribute to generate part of the tremor wave field. At Stromboli, tremor is dominated by *SH* waves *[Chouet et al,* 1997] which strongly support this hypothesis. Our model evidences the importance of gas bubble growth in bubbly magma as the origin of the pressure change in the magmatic column and the contribution of the viscoelastic reaction of magma as a potential source of the tremor frequency content. Moreover, infrasonic recordings indicate that the source repeats in time at a rate of almost 1 s supporting the evidence that magma is degassing through repeated bursting of bubbles which represents a reasonable mechanism of the generation of sustained tremor. We indicate that the delay time between two successive infrasonic pulses could reflect cyclic gas bubble formation. Layers of bubbles could occur cyclically under a wave-like instability *[Manga,* 1996]. A viscosity contrast in the shallow magma portion could produce instability with a wavelenght of some meters to few centimeters *[Thomas et al,* 1993]. Therefore, we assume that gas bubbles (~0.5 m) are formed by coalescence of a layer of small bubbles (~1 cm) at a rate of ~1-2 s. This quasi-cyclic behavior does not change the spectral characteristic of the volcanic tremor but introduces local frequency peaks according to the delay times between the growth of successive bubbles. The ground displacement spectrum of volcanic tremor recorded at Stromboli shows the same asymptotic decay as the first derivative of our theoretical source function. We have demonstrated that the low-frequency asymptote and the high-frequency asymptote define a frequency (2.1 Hz) that we named "corner" frequency, which depends on the time duration of the source function. We conclude that the volcanic tremor spectrum at Stromboli is controlled by the viscoelastic reaction of magma to the sudden decompression produced by gas bubble flow in the magmatic conduit.

Acknowledgments. We are grateful to Pasquale Poggi from Istifcuto Nazionale di Otfcica - Firenze for realizing the infrasonic sensor and for the valuable technical support during the years of this work. Special thanks to Giuseppe Soda, Nicola Luise, Saverio Bitossi, Thomas Braun, and Massimo Delia Schiava for the constant help during field experiments. The manuscript has benefited greatly from critical readings of Sharon Kedar. Many tanks go also to an anonymous referee and Associate Editor who have helped us to clarify text and minds. This research was supported by GNV-CNR grants, and E. G. had a NATO-CNR fellowship during his stay in Italy.

References

Aki, K., and P. Richards, *Quantitative Seismology.* 932 pp., W. H. Freeman, New York, 1980.

Aki, K., M. Fehler, and S. Das, Source mechanism of volcanic tremor: Fluid-driven crack models and their application to the 1963 Kilauea eruption, *J. Volcanol Geotherm. Res., 2*, 259-287, 1977.

Barclay, J., D.S. Riley, and R.S.J. Sparks, Analytical models for bubble growth during decompression of high viscosity magmas, *Bull. Volcanol, 57,* 422-431, 1995.

Batchelor, G.K., *An Introduction to Fluid Dynamics*, 615 pp., Cambridge Univ. Press, New York, 1967.

Benoit, J.P., and S.R. McNutt, New constraints on source processes of volcanic tremor at Arenal volcano, Costa Rica, using broadband seismic data, *Geophys. Res. Letters, 21* 449-452, 1997.

Blackburn, E.A., L. Wilson, and R.S.J. Sparks, Mechanism and dynamics of Strombolian activity, *J. Geol Soc. London, 132*, 429-440, 1976.

Bland, D.R., *The Theory of Linear Viscoelasticity.* 157 pp., Pergamon, Tarrytown N.Y., 1960.

Braun, T., and M. Ripepe, Interaction of seismic and air waves as recorded at Stromboli volcano, *Geophys. Res. Lett., 20,*, 65-68, 1993.

Braun, T., J. Neuberg, and M. Ripepe, On the origin of the long-period tremor recorded at Stromboli volcano (Italy), *Ann. Geophy., 39*, 311-326, 1996.

Buckingham, M.J., and M.A. Garces, Canonical model of volcano acoustics, *J. Geophys. Res., 101*, 8129-8151, 1996.

Chouet, B., N. Hamisevicz, and T.R. McGetchin, Photoballistic of volcanic jet activity at Stromboli, Italy, *J. Geophys. Res., 79*, 4961-4976, 1974.

Chouet, B., Excitation of a buried magmatic pipe: A seismic source model for volcanic tremor, *J. Geophys. Res., 90,* 1881-1893, 1985.

Chouet, B., G. Saccorotti, M. Martini, P. Dawson, G. De Luca, G. Milana, and R. Scarpa, Source and path effects in the wave fields of tremor and explosions at Stromboli Volcano, Italy, *J. Geophys. Res.*, *102*, 15,129-15,150, 1997.

Correig A.M., and J. Vila, On the frequency contents of local events: Source or path effect?, *Geophys. J. Int.*, *115*, 863-877, 1993.

Della Schiava, M., S. Ciliberto, C. Laroche, and M. Ripepe, Explosive source mechanism at Stromboli deduced by laboratory experiment, *Atti Convegno Annuale G.N.G.T.S.*, *15*, 355-367, 1996.

Ferrazzini, V., K. Aki, and B. Chouet, Characteristics of seismic waves composing Hawaiian volcanic tremor and gas-piston events observed by a near-source array, *J. Geophys. Res.*, *96*, 6199-6209, 1991.

Francalanci, L., P. Manetti, and A. Peccerillo, Volcanological and magmatological evolution of Stromboli volcano (Aeolian islands): The role of fractional cristallization, magma mixing, crustal contamination and source heterogeneity, *Bull. Volocanol.*, *51*, 355-378, 1989.

Gordeev, E.I., Modeling of volcanic tremor as explosive point sources in a singled-layered, elastic half-space, *J. Geophys. Res*, *98*, 19,687-19,703, 1993.

Gordeev, E.I., V.A. Saltykov, V.I. Sinitsyn, and V.N. Chebrov, Temporal and spatial characteristics of volcanic tremor wave fields, *J. Volcanol. Geotherm. Res.*, *40*, 89-101, 1990.

Jaupart, C., and S. Vergniolle, Laboratory models of Hawaiian and Strombolian eruptions, *Nature*, *331*, 58-60, 1988.

Jiang, J., G. R. Baird, and D.P. Blair, Dynamic responce of a half-space to a buried spherical source, *Geophys. J. Int.*, *119*, 753-765, 1994.

Kamo, K., T. Furuzawa, and J. Akamatsu, Some natures of volcanic micro-tremors at Sakurajima Volcano, Japan, *Bull. Volcanol. Soc. Jpn.*, *22*, 41-48, 1977.

Kanamori, H., J.W. Given, and T. Lay, Analysis of seismic body waves excited by Mount St. Helens eruption of May 18, 1980, *J. Geophys. Res*,. *89*, 1856-1866, 1984.

Kedar, S., B. Sturtevant, and H. Kanamori, The origin of harmoric tremor at Old Faithful Geyser, *Nature*, *379*, 708-711, 1996.

Kubotera, A., Volcanic tremors at Aso volcano, in *Physical Volcanology*,edited by L. Civetta et al., pp. 29-47, Elsevier, New York, 1974.

Lighthill, M.J., On sound generated aerodinamically, I, General theory, *Proc. R. Soc. London A*, *211*, 564-587, 1978.

Lu, N.Q., H.N. Oguz, and A. Prosperetti, The oscillations of a small floating bubble, *Phys. Fluids A*, *2*, 252-260, 1989.

Mader, H.M., Y. Zhang, J.C. Philips, R.S.J. Sparks, B. Sturtevant, and E.M. Stolper, Experimental simulations of explosive degassing of magma, *Nature*, *372*, 85-88, 1994.

Manga, M., Waves of bubbles in basaltic magmas and lavas, *J. Geophys. Res.*, *101*, 17,457-17,465, 1996.

McMillan, D.K., R.W. Cross, and P.E. Long, Two-stage vesiculation in the Cohassett flow of the Grande Ronde basalt, south-central Washington, *Geology*, *15*, 809-812, 1987.

McNutt, S., Volcanic tremor from around the world, *Bull. N. M. Bur. Mines Miner. Resour.*, 131-183, 1989.

Mori, J., H. Patia, C. McKee, I. Itikari, P. Lowenstain, P. De-Saint-Ours, and B. Talai, Seismicity associated with the eruptive activity at Langila volcano, Papua New Guinea, *J. Volcanol. Geotherm. Res.*, *38*, 245-255, 1989.

Napoleone, G., M. Ripepe, G. Saccorotti, L. Totti, and A. Albianelli, The Strombolian activity presently recorded at Stromboli, *Acta Volcanol.*, *3*, 163-171, 1993.

Neuberg, J., R. Luckett, M. Ripepe, and T. Braun, Highlights from a seismic broad-band array on Stromboli volcano, *Geophys. Res. Lett.*, *21*, 749-752, 1994.

Okada, H., Y. Nishimura, H. Miyamachi, H. Mori, and K. Ishihara, Geophysical significance of the 1988-1989 explosive of Mt. Tokachi, Hokkaido, *Japan*, *Bull. Volcanol. Soc. Jpn.*, *35*, 175-203, 1990.

Parfitt, E.A., and L. Wilson, Explosive volcanic eruption, IX, The transition between Hawaiian-style lava fountaining and Strombolian explosive activity, *Geophys. J. Int.*, *121*, 226-232, 1995.

Proussevitch, A.A., D.L. Sahagian, and A.T. Anderson, Dynamics of diffusive bubble growth in magmas: isothermal case, *J. Geophys. Res.*, *98*, 22,283-22,307, 1993.

Ripepe, M., Evidence for gas influence on volcanic seismic signals recorded at Stromboli, *J. Volcanol. Geotherm. Res.*, *70*, 221-233, 1996.

Ripepe, M., P. Poggi, T. Braun, and E. Gordeev, Infrasonic waves and volcanic tremor at Stromboli, *Geophys. Res. Lett.*, *23*, 181-184, 1996.

Sahagian, D.L., A.T. Anderson, and B. Ward, Bubble coalescence in basalt flows: Comparison of a numerical model with natural examples, *Bull. Volcanol.*, *52*, 49-56, 1989.

Schick, R., Volcanic tremor-source mechanisms and correlation with eruptive activity, *Nat. Hazards*, *1*, 125-144, 1988.

Schlindwein V., J. Wassermann, and F. Scherbaum, Spectral analysis of harmonic tremor signals at Mt. Semeru volcano, Indonesia, *Geophys. Res. Lett.*, *22*, 1685-1688, 1995.

Scriven, L.E., On the dynamics of phase growth, *Chem. Eng. Sci.*, *10*, 1-13, 1959.

Seidl, D., R. Schick, and M. Riuscetti, Volcanic tremor at Etna: A model for hydraulic origin, *Bull. Volcanol.*, *44*, 43-55, 1981.

Sparks, R.S.J., The dynamics of bubble formation in magmas: A review and analysis, *J. Volcanol. Geotherm. Res.*, *3*, 1-37, 1978.

Sparks, R.S.J., and S. Brazier, New evidence for degassing processes during explosive eruptions, *Nature*, *295*, 21-23, 1982.

Sugioka, I., and M. Bursik, Explosive fragmentation of erupting magma, *Nature*, *373*, 689-692, 1995.

Thomas, N., S. Tait, and T. Koyaguchi, Mixing of stratified liquids by the motion of gas bubble: Application to magma mixing, *Earth Planet. Sci. Lett.*., *115*, 161-175, 1993.

Thomas, N., C. Jaupart, and S. Vergniolle, On the vesicularity of pumice, *J. Geophys. Res.*, *99*, 15,633-15,644, 1994.

Toramaru, A., Vesiculation process and bubble size distributions in ascending magmas with constant velocities, *J. Geophys. Res.*, *94*, 17,523-17,542, 1989.

Uhira, K., and M. Takeo, The source of explosive eruptions of Sakurajima volcano, Japan, *J. Geophys. Res.*, *99*, 17,775-17,789, 1994.

Vergniolle, S., and G. Brandeis, Origin of the sound generated by Strombolian explosions, *Geophys. Res. Lett.*, *21*, 1959-1962, 1994.

Vergniolle, S., and G. Brandeis, Strombolian explosions, 1, A large bubble breaking at the surface of a lava column as a source of sound, *J. Geophys. Res.*, *101*, 20,433-20,447, 1996.

Vergniolle, S., and C. Jaupart, Separated two-phase flow and basaltic eruptions, *J. Geophys. Res.*, *91*, 12,842-12,860, 1986.

Vergniolle, S., G. Brandeis, and J.-C. Mareschal, Strombolian explosions, 2, Eruption dynamics determined from acoustic measurements, *J. Geophys. Res.*, *101*, 20,449-20,466, 1996.

Zhang, Y., B. Sturtevant, and E.M. Stolper, Dynamics of gas-driven eruptions: Experimental simulations using C02-H20 polymer system, *J. Geophys. Res., 102*, 3077-3096, 1997.

M. Ripepe Dipartimento di Scienze della Terra, Universita' di Firenze, Via G. La Pira,4 50121 Florence, Italy (e-mail: maurizio@ibogfs.cineca.it)

E. Gordeev Geophysical Service, Russian Accademy of Science, 683006 Petropavlovsk-Kamchatsky, Russia (e-mail: gord@omsp .kamch atka. su)

(Received October 28, 1996; revised July 15, 1998; accepted August 13, 1998.)