# Classification and quantification of volcanic eruptions

R. Scandone, L. Giacomelli, F. Fattori Speranza and W. Plastino

*Dipartimento di Fisica, Università degli Studi "Roma Tre", Roma, Italy*

(Received: March 5, 2008; accepted: October 15, 2008)

***ABSTRACT*** The eruptive style of volcanoes is the result of a complex process, which begins at depth and ends with the arrival of magma with variable amount of volatiles at the Earth's surface. The amount of gas, the magma composition, and the supply rate are the main factors controlling the either effusive or explosive character of an eruption. Degassing during the ascent favors a non-explosive eruptive style. Magma may reside in crustal magma chambers for a long time and may be subject to cooling processes with differentiation and separation of batches having different physico-chemical properties. Eruptions may be the result either of a direct connection between a crustal magma chamber and the surface or the ascent of single batches detaching themselves from a larger reservoir, each with its own history of ascent and composition. The eruptive style is the result of the complex mechanism of ascent of magma from its generation to its arrival at the surface. In this paper, we revise earlier schemes of eruption classification. We propose a conceptual grid of classification of the style of volcanic eruptions comprising the main descriptive terms that have been used in the past. The scheme is based on the interplay of a limited number of controlling factors (composition, velocity of ascent and mechanism of supply) that are thought to affect the way in which a volcano may erupt more relevantly. We further provide a more rigorous definition of the magnitude and intensity of an eruption that can be used for the quantitative comparison of eruptions of different volcanoes.

## 1. Introduction

Volcanism is the result of the convective heat loss from the interior of the Earth. The magma which is generated within the mantle or upper crust, rises towards the surface essentially because of a density contrast; during its rise it differentiates and cools. The peculiarities of the ascent are responsible for the nature and character of volcanic activity.

The activity of volcanoes displays a wide range of eruptive styles and sizes of eruption. A single eruption may emit volumes of magma ranging over different orders of magnitude, with a similar variation in the magma discharge rate. Effusive eruptions are characterized by small erupted volumes and discharge rates, when compared with explosive eruptions. Occasionally, effusive events with volumes of different cubic kilometers, may occur. The style of an eruption is mainly controlled by the amount of volatiles coupled with magma reaching the surface, which, in turn, is controlled by the rate of ascent (Scandone *et al.*, 2007).

Perret (1924) was among the first to recognize the role of volatiles during explosive eruptions when describing the eruption of the Vesuvius in 1906: "Gas is the active agent and magma is the vehicle”. Sparks (1978) first quantified the mechanism of gas exsolution and bubble formation suggesting that fragmentation of magma occurs because of gas pressure growth inside closed packed bubbles during ascent and decompression of magma. This paper has permitted the development of a first generation of models of magma ascent in conduits taking into account gas exsolution and bubble growth (e.g. Wilson and Head, 1981) by assuming degassing under closed conditions (gas not lost to surroundings, but trapped in the bubbles). Under these conditions, the different eruptive styles (explosive versus effusive), even during the same eruption or eruptive period, were ascribed to different original gas contents feeding the eruption from within the magma chamber (e.g. Scandone and Malone, 1985).

Eichelbergher *et al*. (1986) suggested that the formation of a permeable foam allows the degassing to occur from the walls of the conduit, thus permitting non-explosive silicic eruptions even with an originally water-rich magma. Recent experiments have shown that homogeneous bubble nucleation in silicic magmas requires very large supersaturations ($\Delta P \sim 150$ MPa) regardless of the decompression rate (Mangan and Sisson, 2000). Moreover, preserved vesicle textures in silicic pumice (even those with abundant phenocrysts) appear to require homogeneous bubble nucleation, at least during a rapid magma ascent that accompanies an explosive activity (Klug and Cashman, 1994; Cashman, 2004).

On the basis of these ideas, Scandone *et al.* (2007) infer that the eruptive style is the result of the ascending history of individual magma batches. This model can thus show that:

1 - degassing processes may be inhibited if bubble nucleation and growth require high supersaturation. In this case, the magma arrives at the surface with high supersaturation and volatile content thus favouring an explosive style of eruption;

2 - exsolution processes with low supersaturation favour bubble nucleation and growth forming a permeable foam and favouring volatile loss during ascent. In these processes, effusive eruptions will be favoured;

3 - mechanism of ascent and the time required to reach the surface may affect the amount of degassing and the style of an eruption;

4 - a high discharge rate, during explosive eruptions, is the result of the disruption of magma in the conduit that drastically decreases the viscosity of the eruptive mixture.

The scope of this paper is to review the different styles of eruptions with the aim of providing a scheme of classification of eruptions based on a limited number of controlling parameters. Based on this scheme, we suggest the most useful parameter that can permit a better quantification of the “size” of eruptions.

## 2. Factors controlling the style of eruptions

In the past, it was assumed that the chemical composition of magma was the main factor controlling the explosivity of an eruption (e.g. McBirney, 1973). However, we observe both end members of volcanic eruptions (effusive or explosive) irrespective of the silica abundance of magma. It is more appropriate to say that magma which is richer in silica has a higher probability of being erupted explosively than basaltic magmas. Such property, related to the rehology of magma, is in turn controlled by external factors that affect the way in which the magma reaches the surface and the time spent at different depths below the Earth’s surface. In the following, we will examine these external factors in more detail.

### *2.1. The role of velocity of ascent*

Scandone *et al*. (2007) have shown that the style of silicic eruptions is basically controlled by the mechanism of ascent and the time spent above saturation depth. Magmas with the same initial volatile content may erupt either explosively or with lava effusion also during a basaltic eruption (Marianelli *et al*., 2005). High supersaturation required for homogeneous bubble nucleation in silicic magmas (Mangan and Sisson, 2000) implies that the magma ascends, for much of its path, to the surface as a liquid without bubbles. Foam formation is likely to occur in the final 2-3 km from the surface. Explosive fragmentation of magma or diffusive degassing of the foam depends mostly on the time spent within this zone. An explosive eruption is likely to occur if the ascent, after the foam formation, is fast, not allowing efficient degassing, otherwise an effusive eruption may occur. Average velocities of the order of tens of cm/s are required to start explosive eruptions of silicic magmas (Scandone *et al.*, 2007). Much higher peak velocities are likely in the last 2 km of ascent to the surface. Similar mechanisms of eruption are put in evidence by the explosive eruptions of Vesuvius which occurred during the last period of persistent activity between 1631 and 1944 (Scandone *et al.*, 2008). Scandone *et al*. (2007) estimate that, after a few hours (5-10) at a depth shallower than the saturation depth, the magma may exsolve, segregate and lose much of its volatiles and explosive potential unless sudden decompression like flank failure or dome collapse takes place.

### *2.2. The role of magma supply*

Small-to medium-sized eruptions commonly comprise a succession of eruptive pulses separated by quiescent periods lasting up to weeks or months, with each eruptive pulse characterized by its own precursory phenomena, textural character of products, and water content.

The ascent of isolated magma batches may explain the episodic explosive eruptions lasting tens of minutes (Fig. 1a) (Scandone *et al.*, 2007). In these cases, the Magma Discharge Rate (MDR) is a measure of the velocity of fragmentation and depends basically on the physical magma properties (e.g.: the small explosive eruptions of St. Helens in the summer of 1980 and March 1982, the eruptions of Mt. Spurr in 1992, the first phase of the Redoubt eruption in 1989, etc.). The conditions leading to sustained explosive eruptions, lasting hours-days, require that a large magma reservoir be directly connected to the surface by a continuous conduit and that the fragmentation surface propagates down to the reservoir thus allowing a fast ascent of the magma (Fig. 1b). In these cases, the MDR is controlled basically by the overpressure of the deep feeding reservoir and the eruption ends either when all the magma is used up or when the pressure of the reservoir determines the collapse of the chamber walls and shuts off the eruption (Scandone and Malone, 1985). Alternatively, the downward propagation of the fragmentation front reaches a depth where the process is no longer allowed. The collapse of the chamber is likely during major, explosive eruptions that are fed by a central conduit (e.g. St. Helens 1980, Pinatubo 1991). The aspect ratio of the depth/width of the chamber roof controls the possibility of opening new conduits and may result in a caldera collapse along annular faults thus permitting the continuation of the eruption also in the presence of a collapse of the chamber walls (Scandone and Acocella, 2007).

Effusive eruptions require longer residence times at shallower depths. The longer times not only permit the diffusive degassing, but may enhance the mechanism of bubble formation by heterogeneous nucleation. Extensive decompression crystallization occurs if sufficient time is spent at shallow depths (Blundy and Cashman, 2001) with the growth of oxide microcrystals that act as nuclei for heterogeneous growth of bubbles. Such process is favoured in hotter and less viscous basalt with faster growth of microlites. The process substantially lowers the super-saturation required for bubble formation thus favoring an extensive degassing of magma. Slow ascent velocities are also put in evidence by the alteration of hornblende rims during the dome eruption of St. Helens between 1980 and 1987 (Rutherford and Hill, 1993); similarly, extensive syn-ascent crystallization characterizes the extruded dome lobes (Cashman, 1992).

A bimodal distribution of crystallinity and gas content also characterizes the eruptions of Stromboli and Vesuvius (Bertagnini *et al*., 2003; Marianelli *et al*., 2005; Scandone *et al*., 2008). High crystallinity, water-depleted magmas are erupted during an effusive eruption, and crystal-poor, water-rich ones during violent Strombolian events. In both cases, it is inferred that the magma erupted in an effusive style and resided for periods lasting months-years at depths shallower than 3 km. Conversely, magma that erupted during explosive events ascended with higher velocity from a deeper reservoir.

A slow ascent of isolated magma batches (Fig. 1c) is the likely cause of effusive eruptions of short duration. These eruptions generally produce single channel lava flows or single lobes or spines, and, depending on the residual amount of volatiles, eruptions of mafic magmas may have a relatively high eruption rate and moderate Strombolian activity or lava fountains. Some eruptions may be characterized by the arrival of different pulses at intervals of days with episodic increases of eruption rate.

Effusive eruptions, lasting several months-years with low Strombolian activity require continuous feeding possibly due to the establishment of an accumulation zone at shallow depth where magma degasses (Fig. 1d). The direct connection of this reservoir to the surface, and the establishment of a pseudo stationary condition where the magma output is approximately equal or lower than the episodic input into the shallow reservoir is likely. A direct connection to a deeper reservoir is unlikely, as it would permit higher eruption rates and a different style of eruption. Episodic increases of the eruption rate may result from new arrivals in the shallow reservoir. In these cases, a small volume of the shallow chamber, not big enough to buffer the arrival of new ascending batches is likely.

## 3. A classification scheme of eruptions

The lower solubility of water of mafic magmas increases the depth of the degassing zone and, at the same time, the lower viscosities of this magma may permit the aggregation of bubbles and formation of gas slugs that may easily escape from the liquid. As a consequence, explosive eruptions of mafic magmas are less frequent since they require a higher ascent velocity of magma batches.

Taking into account this limitation we propose a simple classification scheme based on the composition of erupted magma, velocity of ascent, and mode of magma supply. This simple

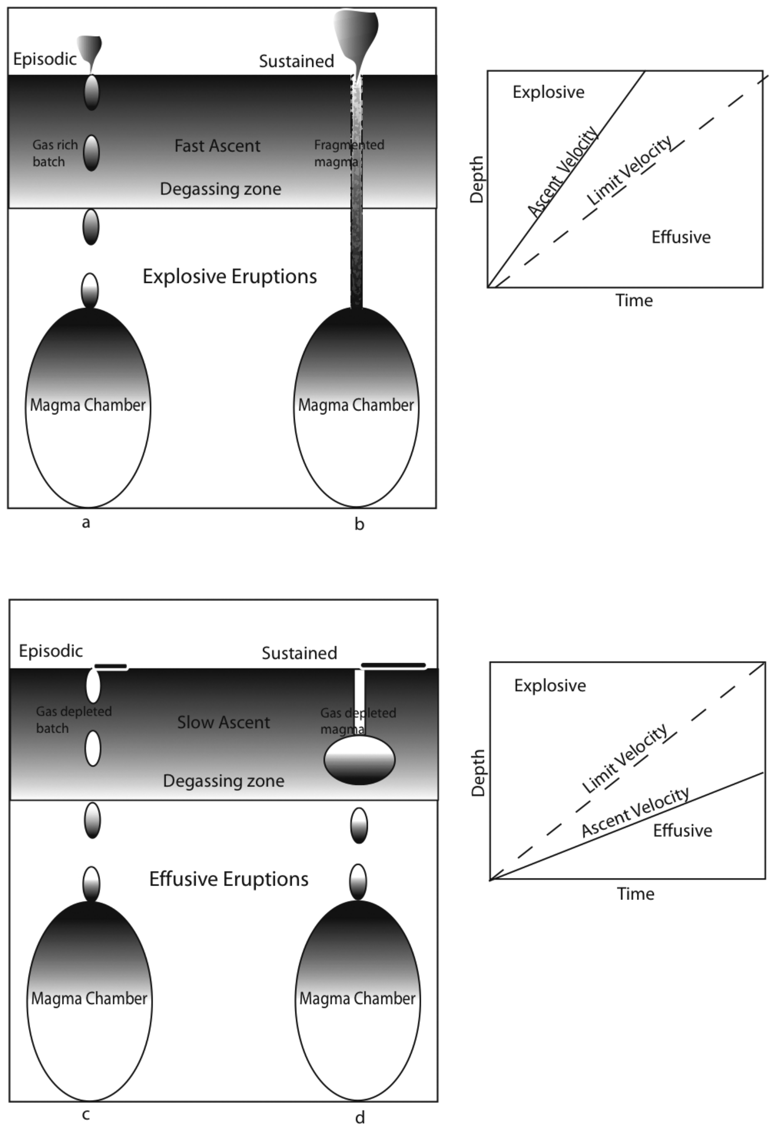


Fig. 1- Scheme of magma ascent and eruptive mechanisms. The right-hand side insets show the average ascent velocity given by the slope of the straight line. The case above shows the velocity of ascent which is higher than the minimum velocity thus allowing degassing; below, the velocity is lower than the limiting velocity: a) ascent of fast, isolated magma batches resulting in explosive episodic eruptions; b) connected conduit permitting the eruption of a fragmented magma mixture and the occurrence of a sustained explosive eruption; c) ascent of slow isolated magma batches resulting in effusive episodic eruptions; d) conduit connected to a shallow reservoir permitting the eruption of a liquid magma and the occurrence of a sustained effusive eruption.

scheme (Fig. 2) explains the great variety of different eruptions observed on Earth.

The combination of three different factors determines 8 different eruption styles, which represent the end members of a continuous range of activity. A transition from one style to another even in the course of the same eruption is likely.

The eruptions have been subdivided first on the base of magma composition, as it strongly affects the water solubility, and the degassing mechanism. Silicic volcanism leads mostly to a mixed style of activity often making volcanoes with steep flanks decapitated by caldera-forming eruptions or flank failures. Basaltic volcanoes, on the other hand, have a predominantly effusive style of eruption, which determines edifices with gentle slopes and rarer explosive eruptions.

The average velocity of ascent, is the second factor to take into account, which mostly controls the effusive versus explosive style of activity. We infer that the predominant effusive eruptions of basaltic volcanoes are due to the low viscosity, lower solubility of volatiles and easier degassing during the ascent so that the required velocity of ascent for explosive eruptions is higher than that for silicic magmas and these events are relatively less common with mafic magmas.

The supply mode is the last controlling factor. The statistics of volcanic eruptions (Simkin and Siebert, 1994) point out that low-volume eruptions are the predominant style of activity. Scandone *et al.* (2007) infer that this is the result of the ascent of isolated magma batches (quanta) which detach themselves from a crustal reservoir and ascend towards the surface. The individual history of each batch (velocity) affects the eruptive style. The connection of a crustal reservoir with the surface is likely to cause sustained eruptions. This process is relatively easier if the reservoir is closer to the surface, and more difficult if it is deeper. The closeness to the surface in turn affects the degassing of magma so that sustained effusive eruptions are relatively more frequent than sustained explosive eruptions.

The interaction with external factors may determine important changes in the style of an eruption. Water magma interaction and the failure of domes or volcanic edifices are two of the most common factors that contribute to increasing the explosive potential of ascending magmas.

## 4. Quantification of the "size" and "power" of eruptions

In the past years, considerable efforts have been made to define an index that quantifies the size and style of an eruption.

The Tsuya scale (Tsuya, 1955), based on the volume of erupted products, was not widely used probably because the scale interval was based on multiples of $10^2$ $m^3$.

Hedervary (1963) proposed a scale based on the energy released during eruptions. This scale does not differ very much from the Tsuya scale, taking into account that most of the energy released during an eruption is under the form of thermal energy, which is proportional to the volume of the erupted products (Yokoyama, 1956).

Newhall and Self (1982) proposed the Volcanic Explosivity Index (VEI) as an indicator of the relevance of an explosive eruption. The VEI is a qualitative index based on a number of different parameters like the total volume of ejecta, the column height, the description of the eruption, the stratospheric injection of dust, etc., that are commonly used to characterize an eruption. This index is a useful tool for the comparison of different explosive eruptions, and to subdivide them into a few general classes. Unfortunately, it fails when dealing with effusive eruptions because it was not devised for this kind of activity. Further on, a discrepancy between the VEI for eruptions which have not been observed and those with visual observations may arise. In fact, for the former, the parameter that weights more in assigning the VEI, is the total volume of ejecta; on the contrary, the VEI of eruptions that occurred in the last decade is assigned mostly on the height of the eruption plume. This parameter has been estimated only for a limited sample of eruptions that occurred in the past, based on detailed field investigations. Even with these strong limitations, the VEI proved useful in defining a scale law for eruptions that occurred during the past two decades (Simkin and Siebert, 1994), thus suggesting an implicit quantitative character.

Walker (1980) proposed to characterize explosive eruptions with three different parameters: the magnitude (the total volume of erupted products), the intensity (the discharge rate), and the

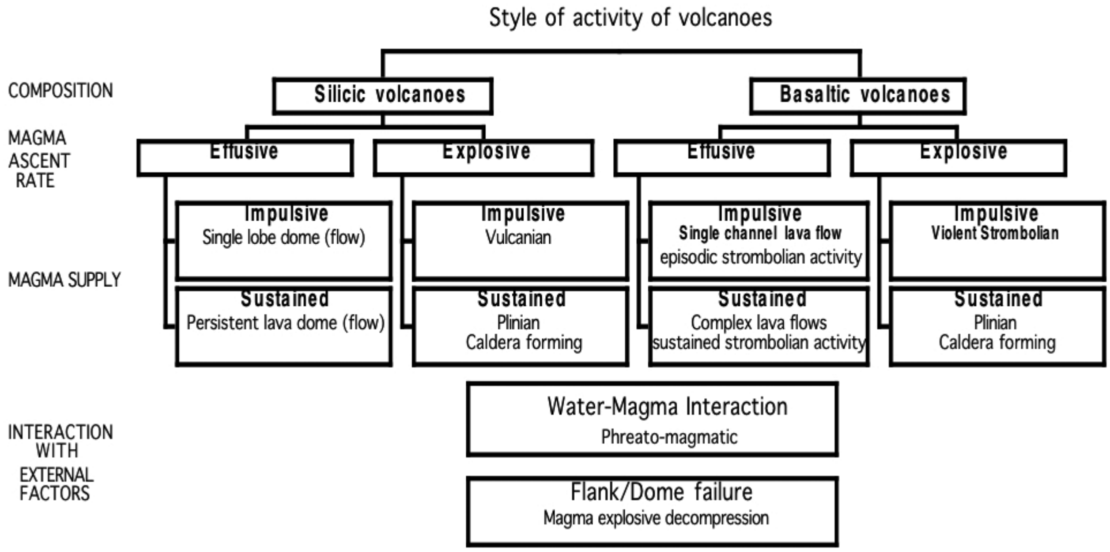


Fig. 2- Scheme of classification of volcanic activity based on the controlling parameters shown on the left. The first row shows the effect of the composition. The second row illustrates the role of ascent velocity, with high ascent rates resulting in explosive activity, and a low ascent rate resulting in an effusive activity. The third and fourth rows show the combinations of the different factors with episodic magma supply resulting in impulsive eruptions and connected magma supply resulting in sustained eruptions. The lower rows illustrate the effect of interaction of magma with external factors.

violence (the modality with which pyroclastic flow deposits are emplaced).

Carey and Sigurdsson (1989) estimated the intensity of a number of Plinian eruptions occurring in the past by evaluating the height of the eruptive column and showing a rough correlation between intensity and magnitude.

The VEI (Newhall and Self, 1982) strongly relies on two main factors: the volume of erupted products, and the height of the volcanic plume. The former parameter has been basically used to assign the VEI for eruptions that occurred in the past or of effusive type [see for example the catalogue "Volcanoes of the World" by the Smithsonian Institution; Simkin and Siebert (1994)]. The latter is widely used to assign a VEI to current, observed volcanic eruptions.

These two parameters measure the size of the eruption (the volume), and, under particular conditions, its power (energy/time) (the plume height). The plume height is proportional to the fourth root of the emitted power for a sustained eruption plume (Settle, 1978; Wilson *et al.*, 1978). Unfortunately, the plume height does not give a true estimate of the output when the column collapses and produces pyroclastic flows (Sparks and Wilson, 1976). A VEI assigned merely based on column height as in these cases, would underestimate the eruption explosivity.

In order to gain a better insight into the significance of the eruption plume height, we recall its relationship with to emitted power (Wilson *et al.*, 1978):

$$H_t = 8.2 \cdot \dot{Q}^{1/4} \tag{1}$$

where $H_t$ is the maximum height of the column and $\dot{Q}$ is the stationary rate of energy (in Watt). It can also be written so:

$$\dot{Q} = \rho v \pi r^2 C_p (T - T_0) F \tag{2}$$

where $\rho$, $v$, $C_p$, $T$ are respectively the density, velocity, specific heat, and temperature of the erupted mixture, $r$ the radius of the conduit (the term $\rho v \pi r^2$ is the mass flux) and $T_o$ the cooling temperature of the mixture. $F$ is an efficiency factor varying between 0.7 and 1 (Wilson *et al.*, 1978).

Fig. 3 shows the height of a sustained eruption plume as a function of the MDR, using a density of 2500 kg $m^{-3}$, a specific heat of 900 J $kg^{-1}K^{-1}$, an initial temperature of the eruptive mixture of 1150 K and final temperature of 293 K, and an efficiency factor $F$=1. The figure also shows, for comparison, the effective plume heights measured for recently observed explosive eruptions (half-filled square) and those evaluated by Carey e Sigurdsson (1989) (open dots).

The VEI, assigned by plume height, is based on a range of heights of the plume (Table 1). We may calculate the corresponding ranges of magma discharge using Eqs. (1) and (2).

This simple calculation in Fig. 3 shows the irregularity of the scale. For example, the VEI=1 covers four orders of magnitude of discharge rate; the VEI= 2 almost three orders, VEI= 3 two and half, and so on. The different range of VEI may possibly prevent a correct scaling of eruptions.

A second point that must be emphasized is the collapse of the eruption column and the generation of pyroclastic flows. Theoretical models (Sparks and Wilson, 1976; Sparks *et al.*, 1978), field evidence (Walker, 1985), and the observation of current eruptions (Carey *et al.*, 1990) seems to indicate that a massive production of pyroclastic flows occurs when the MDR becomes so high that the turbulence of the column is unable to mix sufficient air. The efficiency ($E$) of the eruption [the ratio between the potential energy over the thermal energy; Scandone and Giacomelli (1998)] measures the ratio between the energy transformed into mechanical energy (potential) and the energy available under the form of thermal energy:

$$\begin{aligned} E &= \frac{M_h g h}{M_{tot}(C_p \Delta T + L)} = \frac{x g h}{(C_p \Delta T + L)} \\ x &= \frac{M_h}{M_{tot}}, \end{aligned} \tag{3}$$

where $M_h$ is the total mass of product raised in the plume to the height $h$, $M_{tot}$ is the total mass of product, $C_p$ is the specific heat, $\Delta T$ the temperature decrease, and $L$ the latent heat. The typical efficiency of a Plinian eruption with a 30 km high plume is of the order of 20-25%. If, for example, the column collapses and produces pyroclastic flows from an ash fountain 5 km high (the jet portion of the column), the efficiency drops to about 4% (without considering the co-ignimbrite plume). The decrease of efficiency is not due to a decrease of mass discharge, but to a poor mixing of the eruptive mixture with ambient air. In this case, we hypothesize that

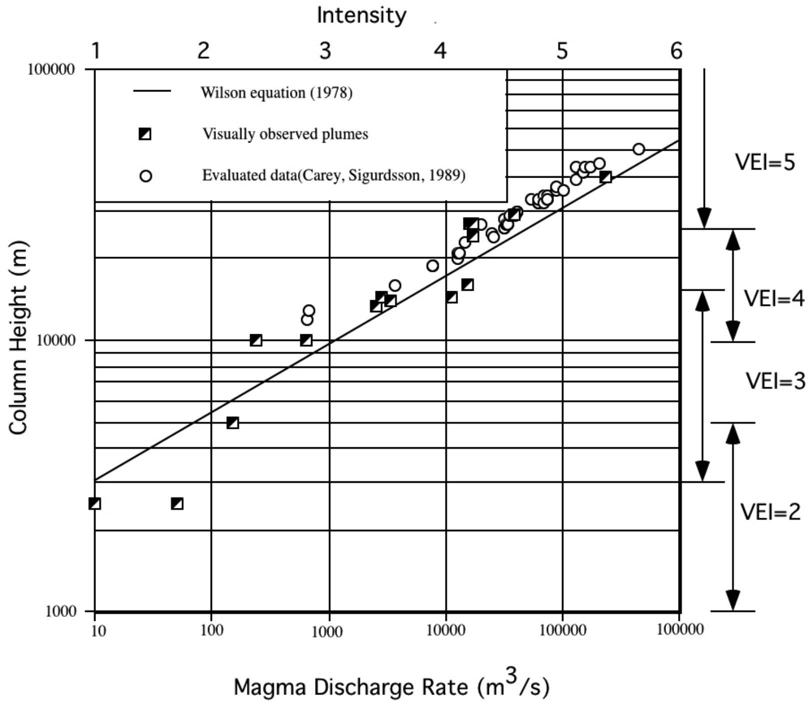


Fig. 3 - Relationship between the MDR and the height of the eruption plume. The straight line is based on the equation of Wilson *et al.* (1978) for sustained eruptions. On the right-hand side the VEI assigned according to the definition of Newhall and Self (1982) based on the plume heights is shown. The top scale is the intensity evaluated as the logarithm of the MDR.

pyroclastic flows are produced by a simple collapse of an eruption column; different efficiency estimates arise if they are generated by an overpressured jet.

In any case, the VEI, assigned on the basis of the eruption plume height, would fail to account for the true intensity of the eruption.

### *4.1. The intensity measured by MDR*

Table 2 shows the ranges of MDR observed for different volcanoes in the course of their past or current activity. The main peculiarities shown in the table are that effusive eruptions have a generally lower discharge rate with respect to explosive ones. Although this table accounts mostly

Table 1 - VEI based on plume height.

| VEI | Plume Height km | MDR ($m^3/s$) |
|---|---|---|
| 1 | 0.1-1 | $1.1 \times 10^{-5}$-0.12 |
| 2 | 1-5 | 0.12-72 |
| 3 | 3-15 | 9.3-5807 |
| 4 | 10-25 | 1147-44810 |
| 5 | >25 | >44810 |

Table 2 - MDR ranges for different volcanoes.

| VOLCANO | MDR ($m^3$/sec) | Intensity | Style of Activity | Reference |
|---|---|---|---|---|
| Etna (1614-1975) | 0.26-119 | -0.59-2.08 | Effusive | Wadge, 1981 |
| Mauna Loa (1843-1975) | 2.5- 332 | 0.4-2.52 | " | " |
| Kilauea (1840-1977) | 0.06-432 | -1.22-2.64 | " | " |
| Tolbachik (1975-1976) | ~11 | 1.04 | " | Fedotov and Markhinin, 1983 |
| St.Helens (1980)(*) | 3.8-82 | 0.58-1.91 | " | Lipman and Mullineaux, 1981 |
| Hekla (1947) | 8-76 | 0.90-1.88 | " | Wadge, 1982 |
| Vesuvius (1944)* | 50 | 1.7 | " | Imbo', 1949 |
| Laki 1783-84 | 60 (average), 7000 (peak discharge) | | | Thordarson and Self, 2003 |
| Ngaururhoe (1974) | 10 | 1 | Explosive | Wilson *et al.*, 1978 |
| Tolbachik (1976) | 50 | 1.7 | " | Fedotov and Markhinin, 1983 |
| Vesuvius (1944)** | 70-150 | 1.85-2.18 | " | Scandone *et al.*, 1986 |
| Ngauruhoe (1975) | 370 | 2.57 | " | Nairn and Self, 1978 |
| Fuego (1971) | 640 | 2.81 | " | Wilson *et al.*, 1978 |
| Usu (1977)(**) | 883 | 2.95 | " | Katsui *et al.*, 1978 |
| St.Helens (1980) (**) | 2000-16000 | 3.3-4.2 | " | Scandone and Malone, 1985; Criswell, 1987 |
| Hekla (1947,1970) | 3333-17000 | 3.5-4.23 | " | Wilson *et al.*, 1978 |
| El Chichon (1982) | 5000-8000 | 3.7-3.9 | " | Varekamp *et al.*, 1984 |
| Soufrière (1902) | 11000-15500 | 4.04-4.19 | " | Wilson *et al.*, 1978 |
| Santa Maria (1902) | 17000-38000 | 4.23-4.58 | " | " |
| Katmai (1912) | ~100000 | 5.0 | " | Ledbetter and Sparks, 1979 |
| Pinatubo (1991) | 154000-185000 | 5.18-5.26 | | Self *et al.*, 1996 |
| Bezymianny (1956) | 230000 | 5.36 | " | Wilson *et al.*, 1978 |
| Toba (70000 abp) | >1000000(***) | 6 | " | Ninkovich *et al.*, 1978 |
| *during dome emission, or lava effusion **explosive phase *** estimated by oceanic sediments | | | | |

for central volcanoes, the data relative to the Laki eruption (1783-1784) show that even fissural eruptions, have a low average discharge rate and the peak discharge rate does not attain the discharge of a small-medium sized explosive eruption. These last have on average an MDR that is higher by at least one or two orders of magnitude than the corresponding effusive eruptions with a similar volume of erupted products. For example, the Crater Peak eruptions of Mt. Spurr during 1992, erupted about $10^7$ $m^3$ in a matter of tens of minutes. Effusive eruptions of Mt. Spurr emit the same volume in a matter of several days or weeks. Such peculiarities suggest that the wide variations of MDR are related to different mechanisms of eruption and not only on the size and geometry of the feeding conduit.

The ascent of the magma is basically controlled by the density contrast with the surrounding rocks so that the order of magnitude change of discharge rate cannot be explained only because of the ascent velocity.

The fragmentation of magma, drastically decreasing the viscosity of the eruptive mixture is by far the most important factor that causes the high eruption rates of explosive eruptions. During this process the fragmentation surface migrates downwards at a velocity higher than the magma ascent (Scandone and Malone, 1985) and the eruption ends either when the ascending batch is completely used up, or when a pressure decrease causes a closure of the conduit. Alternatively, a deepening of the fragmentation surface attains a depth where the mechanism of fragmentation is prevented.

The MDR or intensity is a true quantitative power index, provided that we define an appropriate scale. In order to devise an intensity scale that may be comparable with the VEI scale because of its present use and simplicity, we suggest using the volumetric MDR measured in $m^3s^{-1}$. In practice, we propose the decimal logarithm of the volumetric MDR measured in $m^3s^{-1}$ of dense rock equivalent (DRE) for the intensity scale.

This scale has the great advantage of being logarithmic (like for example the Richter scale) thus covering over seven orders of magnitude, simple, and roughly equivalent to the VEI scale for the range 2-5 (the majority of observed eruptions). It has the further advantage of representing the same physical parameter (MDR) both for explosive and effusive eruptions.

The use of the height of the eruption plumes to assign the intensity of explosive eruptions can be maintained when there is no collapse of the column, provided that we transform the plume height into MDR using the simplified relations of Wilson *et al.* (1978) with the appropriate numeric values:

$$H\ (\mathrm{km}) = 1.718\ Q^{1/4}, \tag{4}$$

where $Q$ is the volumetric flux measured in $m^3s^{-1}$ of DRE, $H$ is the height of the eruptive column in kilometers. According to this definition of intensity ($I$):

$$I=\log\ [Q\ (\mathrm{m^3s^{-1}})], \tag{5}$$

we can rearrange the previous equation:

$$I=\log Q= \log(H/1.718)^4= 4 \log H- 0.94. \tag{6}$$

and we get the corresponding column height for each degree of the scale

$I$ = 1 -> $H$ = 3.0 km;
$I$ = 2 -> $H$ = 5.4 km;
$I$ = 3 -> $H$ = 9.7 km;

$I = 4 \rightarrow H = 17.2$ km;
$I = 5 \rightarrow H = 30.5$ km;
$I = 6 \rightarrow H = 54.3$ km.

These values must be considered as top estimates in cases of complete convection of the eruption plume without any collapse-generating pyroclastic flows. The average value should be used in order to have a uniformity between data obtained from current eruptions and those of the past where only a gross estimate can be made on the peak height of the eruption plumes and corresponding MDR.

The advantage of using the intensity scale is that it is simple, it is based on the MDR which, in turn, is the result of the multiple processes that control the style of an eruption; it is quantitative and may be used both for current eruptions and for past ones.

## 5. Conclusions

Volcanic eruptions and resulting volcanic edifices are the result of the complex ascent of magma towards the surface. We have shown that a limited number of parameters permit the classification of eruptions. Three factors (composition, velocity of ascent, and mechanism of supply) are able to characterize the end members of an eruptive style completely so permitting a better understanding of the nature of volcanism.

The size and power of an eruption are quantitatively described by the magnitude, and intensity (measured as the logarithm of the volume and the discharge rate). These values can be used to compare eruptions at different volcanoes, and to make statistics of volcanic activity using a quantitative index instead of a qualitative one (e.g. VEI). The magnitude and intensity, in turn, depend on the history of the magma ascent.

We are confident that the use of such a simple scheme can help to better define the activity of volcanoes and the comparison among different volcanoes and to better understand the role of the different parameters that control the style of volcanic eruptions.

**Acknowledgements**. The authors wish to thank Dario Slejko for invitating them to present the results of their research to the 26th GNGTS General Assembly in Rome, and M. Rosi and R. Petrini for a useful review of an earlier draft of this paper.

## REFERENCES


Bertagnini A., Metrich N., Landi P. and Rosi M.; 2003: *Stromboli volcano (Aeolian Archipelago, Italy): an open window on the deep-feeding system of a steady state basaltic volcano.* J. Geophys. Res., **108**, DOI: 10.1029/2002JB002146.

Blundy J. and Cashman K.V.; 2001: *Ascent-driven crystallisation of dacite magmas at Mount St. Helens, 1980–1986.* Contr. Mineral. Petrol, **140**, 631-651.

Carey S.N. and Sigurdsson H.; 1989: *The intensity of Plinian eruptions.* Bull. Volcanol., **51**, 28-40

Carey S., Sigurdsson H., Gardner J.E. and Criswell C.W.; 1990: *Variation in column height and magma discharge during the May 18, 1980 eruption of Mt St.Helens*. J. Volcanol. Geoth. Res., **43**, 99-112.

Cashman K.V.; 1992: *Groundmass crystallization of Mount St. Helens dacite, 1980–1986: a tool for interpreting shallow magmatic processes.* Contributions to Mineralogy and Petrology, **109**, 431–449.

Cashman K.V.; 2004: *Volatile controls on magma ascent and eruption.* In: Sparks R.S.J. and Hawkesworth (eds), The State of the Planet: Frontiers and Challenges in Geophysics, American Geophysical Union Monograph **150**, pp. 109-124.

Criswell C.W.; 1987: *Chronology and pyroclastic stratigraphy of the May 18, 1980 eruption of Mount St. Helens, Washington.* J. Geoph. Res., **92**, 10237-10266.

Eichelberger J.C., Carrigan C.R., Westrich H.R. and Price R.H.; 1986: *Non explosive silicic volcanism.* Nature , **323**, 598-605.

Fedotov S.A. and Markhinin Ye.K.; 1983: *The Great Tolbachik fissure eruption: geological and geophysical data*. Cambridge Un. Press, Cambridge, 341 pp.

Hedervari P.; 1963: *On the energy and magnitude of volcanic eruption.* Bull. Volcanol., **25**, 373-386.

Imbò G.; 1949: *L'attivita' eruttiva vesuviana e relative osservazioni nel corso dell'intervallo intereruttivo 1906-1944 ed in particolare del parossismo del Marzo 1944*. Annali Osservatorio Vesuviano, V serie, volume unico, 185-380.

Katsui Y., Oba Y., Onuma K., Suzuki T., Kondo Y., Watanabe T., Niida K., Uda T., Hagiwara S., Naga T., Nishikawa J., Yamamoto M., Ikeda Y., Katagawa H., Tsuchiya N., Shirahase M., Nemoto S., Yokoyama S., Soya T., Fujita T., Inaba K. and Koide K.; 1978: *Preliminary report of the 1977 eruption of Usu Volcano.* Journ. Fac. Sci. Hokkaido Univ., **18**, 385-408

Klug C. and Cashman K.V.; 1994: *Vesiculation of May 18, 1980, Mt St Helens magma.* Geology, **22**, 468-472.

Ledbetter M.T. and Sparks R.S.J.; 1979: *Duration of large magnitude explosive eruptions deduced from graded bedding in deep-sea ash layers*. Geology, **7**, 240-244.

Lipman P.W. and Mullineaux D.R. (eds); 1981: *The 1980 eruptions of Mt St Helens.* USGS Professional Paper 1250, 844 pp.

Mangan M. and Sisson T.; 2000: *Delayed, disequilibrium degassing in rhyolite magma: decompression experiments and implications for explosive volcanism.* Ea. Pl. Sci. Lett., **183**, 441-455.

Marianelli P., Sbrana A., Metrich N. and Cecchetti A.; 2005: *The deep feeding system of Vesuvius involved in recent violent strombolian eruptions.* Geoph. Res. Lett., **32**, doi:10.1029/2004GL021667.

McBirney A.R.; 1973: *Factors governing the intensity of explosive andesitic eruptions.* Bull. Volcanol., **37**, 443-453.

Nairn I.A. and Self S.; 1978: *Explosive eruptions and pyroclastic avalanches from Ngauruhoe in February 1975.* J. Volcanol. Geoth. Res., **3**, 39-60.

Newhall C.G. and Self S.; 1982: *The Volcanic Explosivity Index (VEI): an estimate of the explosive magnitude for historical volcanism.* J. Geophys. Res., **87,** 1231-1238.

Ninkovich D., Sparks R.S.J. and Ledbetter M.T.; 1978: *The exceptional magnitude and intensity of the Toba eruption, Sumatra: an example of the deep-sea tephra layers as a geological tool.* Bull. Volcanol., **41**, 1-13.

Perret F.A.; 1924: *The Vesuvius eruption of 1906. Study of a volcanic cycle.* Carnegie Institution, Washington, 151 pp.

Rutherford M.J. and Hill P.M.; 1993: *Magma ascent rates from amphibole breakdown: an experimental study applied to the 1980 – 1986 Mount St. Helens eruptions*. J. Geophys. Res., **98**, 19667 – 19685.

Scandone R. and Malone S.D.; 1985: *Magma supply, magma discharge and readjustment of the feeding system of Mount St. Helens during 1980.* J. Volcanol. Geoth. Res, **23**, 239-262.

Scandone R. and Giacomelli L.; 1998: *Vulcanologia: principi fisici e metodi di indagine.* Liguori, Napoli, 649 pp.

Scandone R. and Acocella V.; 2007: *Control of the aspect ratio of the chamber on caldera formation during silicic eruptions.* Geophys. Res. Lett., **34**, L22307, DOI:10.1007/s00445-007-0191-9.

Scandone R., Iannone F. and Mastrolorenzo G.; 1986: *Stima dei parametri dinamici dell'eruzione del 1944 del Vesuvio.* Bollettino Gruppo Nazionale di Vulcanologia, **2,** 487-512.

Scandone R., Kashman K.V. and Malone S.D.; 2007: *Magma supply, magma ascent, and the styles of volcanic eruptions.* Ea. Pl. Sci. Lett., **253**, 513-529, doi:10.1016/j.epsl.2006.11.016.

Scandone R., Giacomelli L. and Fattori Speranza F.; 2008: *Persistent activity and violent Strombolian eruptions at Vesuvius between 1631 and 1944.* J. Volcanol. Geoth. Res., **17**, DOI:10.1016/j.jvolgeores.2007.11.018

Self S., Zhao J.X., Holasek R.E., Torres R.C. and King A.J.; 1996: *The atmospheric impact of the 1991 Mount Pinatubo eruption.* In: Newhall C.G. and Punongbayan R.S. (eds), Fire and Mud, Univ. of Wash. Press, Seattle, pp. 1089-1115.

Settle M.; 1978: *Volcanic eruption clouds and the thermal output of explosive eruptions*. J. Volcanol. Geoth. Res., **3**, 309-324.

Simkin T. and Siebert L.; 1994: *Volcanoes of the world. 2nd edition.* Smithsonian Institution. Geoscience Press, Tucson, 349 pp.

Sparks R.S.J. and Wilson L.; 1976: *A model for the formation of Ignimbrites by gravitational column collaps.* J. Geol. Soc. Lond., **132**, 441-452.

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

Sparks R.S.J., Wilson L. and Hulme G.; 1978: *Theoretical modelling of the generation, movement and emplacement of pyroclastic flows by column collapse*. J. Geophys. Res., **83**, 1727-1739.

Thordarson T. and Self S.; 2003: *Atmospheric and environmental effects of the 1783-1784 Laki eruption: a review and reassessment.* J. Geophys. Res., **108**, D1 4011,DOI:10.1029/2001JD002042.

Tsuya H.; 1955: *Geological and petrological studies of volcano Fuji V. 5 On the1707 eruption of Volcano Fuji.* Bull. Earth. Res. Inst., **33**, 341-382

Varekamp J.C., Luhr J.F. and Prestegaard K.L.; 1984: *The 1982 eruptions of El Chicon volcano (Chipas, Mexico): character of the eruptions, ash-fall deposits, and gas phase.* J. Volcanol. Geoth. Res., **23**, 39-68.

Wadge G.; 1981: *The variation of magma discharge during basaltic eruptions.* J. Volcanol. Geoth. Res., **11**, 139-168.

Wadge G.; 1982: *Steady state volcanism: evidence from eruption histories of polygenetic volcanoes.* J. Geophys. Res., **87**, 4035-4049.

Walker G.P.L.; 1980: *The Taupo pumice: product of the most powerful known ultraplinian eruption?* J. Volcanol. Geoth. Res., **8**, 69-94.

Walker G.P.L.; 1985: *Origin of coarse lithic breccias near ignimbrite source vents.* J. Volcanol. Geoth. Res., **25**, 157-171.

Wilson L. and Head J.W.; 1981: *Ascent and emplacement of basaltic magmas on the Earth and Moon*. J. Geophys. Res., **86**, 2971-3001.

Wilson L., Sparks R.S.J., Huang T.C. and Watkins N.D.; 1978: *The control of eruption column heights by eruption energetics and dynamics.* J. Geophys. Res., **83**, 1829-1836.

Yokoyama I.; 1956: *Energetics in active volcanoes: activities of Volcano Mihara, Oshima Island during 1953-54.* Bull. Earth. Res. Inst., **34**, 185-195.

*Corresponding author:* Roberto Scandone
Dipartimento di Fisica
Università "Roma Tre"
Via Vasca Navale 84, 00146 Roma (Italy)
phone: +39 06 57337250; fax: +39 06 57337102; e-mail: scandone@uniroma3.it