Geofísica Internacional
ISSN: 0016-7169
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Universidad Nacional Autónoma de México
México

De la Cruz-Reyna, Servando; Yokoyama, Izumi
A geophysical characterization of monogenetic volcanism
Geofísica Internacional, vol. 50, núm. 4, octubre-diciembre, 2011, pp. 465-484
Universidad Nacional Autónoma de México
Distrito Federal, México

GEOFÍSICA INTERNACIONAL (2011) 50-4: 465-484

REVIEW PAPER

# A geophysical characterization of monogenetic volcanism

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Received: August 5, 2011; accepted: August 12, 2011; published on line: September 30, 2011


## Resumen

En contraste con los volcanes poligenéticos, las erupciones asociadas al nacimiento de volcanes monogenéticos son mucho menos frecuentes y por tanto más difíciles de observar y estudiar. En el mundo, sólo unos pocos casos de este tipo de eventos se han registrado en tiempos históricos. Entre ellos, Jorullo (1759~1766), Paricutín (1943~1952), el maar Ukinrek (1977) y la actividad en East-Izu, Japón (1930~1989) representan algunos de los eventos que fueron reportados y estudiados de acuerdo al nivel científico de cada época. Las dos primeras erupciones formaron conos de escoria de tamaño considerable y grandes flujos de lava a lo largo de varios años de actividad, representando el nacimiento de esos volcanes. El maar se formó como consecuencia de una serie de explosiones freatomagmáticas, con escasa producción de magmas basálticos, y East-Izu fue un evento que causó enjambres sísmicos intermitentes desde 1930 y concluyó con una pequeña erupción submarina en 1989, de magnitud insuficiente para formar un cono escoriáceo. En el presente trabajo se discuten los procesos físicos que originaron esos volcanes monogenéticos, principalmente con base a los datos sísmicos disponibles y otros parámetros reportados que revelan las similitudes y diferencias entre esas erupciones. Sumadas a las similitudes que caracterizan la actividad monogenética, se distinguen diferentes tipos de volcanes. La principal característica común de todos los tipos es el origen profundo, sub-cortical, de sus fuentes magmáticas. Dependiendo de la capacidad de ascenso del magma, resultante de la sobrepresión en la fuente y en sus propiedades boyantes, la actividad eruptiva superficial puede variar desde prolongadas erupciones con alta productividad de magma formando conos cineríticos y extensos campos de lava, hasta erupciones de corta duración y baja productividad de magmas que tienden a permanecer en la corteza sin aflorar. Casos extremos de éste tipo pueden limitarse a actividad freática derivada del calor trasferido por conducción, o a enjambres sísmicos locales. Compartiendo algunas características con estos volcanes, pero en otra escala espacial y con una fuente magmática somera reconocemos otro tipo de vulcanismo monogenético, el de los conos parásitos asociados a volcanes poligenéticos que aquí se sugiere resultan del balance entre la presión magmática y la resistencia mecánica del edificio volcánico poligenético.


Palabras clave: monogenetic volcanism, Jorullo, Paricutín, East-Izu monogenetic volcano group, parasitic cones.

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## Abstract

In contrast with polygenetic volcanoes, eruptions giving birth to monogenetic volcanoes are much less frequent and thus more difficult to observe and study. Only a few events of this type have occurred worldwide in historical times. Among these, Jorullo (1759~1766), Paricutín (1943~1952), Ukinrek maar (1977), and the East-Izu activity in Japan (1930~1989) are among the events that were studied and reported according to the scientific level of each period. The first two eruptions lasted for several years and were actual births of cinder cones and large lava flows. The maar resulted from a series of phreatomagmatic explosions, with relatively small basaltic magma production, and East-Izu caused earthquake swarms intermittently since 1930 ending with a small submarine eruption in 1989, not large enough to form a scoria cone. Here, we discuss the physical processes that originated those monogenetic volcanoes, mostly from the available seismological data and other reported parameters that reveal similarities and differences among those eruptions. Notwithstanding the common features of monogenetic activity, different types of volcanoes may thus be recognized. The main common feature for all types is a deep seated primary magma source, below the crust. Depending on the ascending capacity of magma derived from the excess pressure at the source and on its buoyancy, the surface activity may range from long-duration, high magma productivity eruptions forming cinder cones and extensive lava fields, to short-duration, low productivity eruptions related to similar sources but with a lower magma ascent capacity, that tends to stall within the crust. End members of the latter type may be limited to phreatic activity related to conducted heat phenomena or to local swarm seismicity. Sharing characteristics of those volcano types, but in a different spatial scale and with a shallower magma source, we recognize another type of monogenetic volcanism, namely the parasitic cones associated to polygenetic volcanoes that seem to result from the balance between the magmatic pressure and the mechanical strength of the polygenetic volcanic edifice.


Key words: monogenetic volcanism; Jorullo; Paricutín; East-Izu monogenetic volcano group; parasitic cones.

## Introduction

Worldwide, eruptions giving birth to monogenetic volcanoes are less frequent than eruptions of polygenetic volcanoes. However, along the geological time scale, monogenetic volcanism has formed numerous large groups of volcanoes. We know a few examples of historical monogenetic eruptions, and for some of them significant qualitative and quantitative observational data are available, as described below. A critical interpretation of those data and descriptions should improve our understanding of the nature of monogenetic volcanoes, and particularly of the physical processes that characterize them. In the present paper, we discuss some cases that may help to improve our understanding of such physical processes: The cinder-cone and lava field forming eruptions of Jorullo (1759), and Paricutín (1943), both in Mexico, the 1977 maars-forming eruptions in Alaska, the 1930-1989 episodes of activity in the East-Izu volcano group in Japan. The first two are the birth of sizable monogenetic cones and lava flows observed and reported with the scientific criteria of their times; the maars are craters formed by phreatomagmatic explosions, in some cases with a transitional phase to Strombolian activity, and East-Izu is a small eruption deemed as representative of the final stage of a group of monogenetic volcanoes. We also refer some other cases interpreted as examples of an end member of the latter case, the “failed” eruptions in monogenetic fields, in which the ascending magma does not reach the surface, and only local seismic swarm activity, and in some cases minor phreatic explosions probably caused by conducted heat interacting with ground water reveal the existence of the process.

The 1759 birth eruption of Jorullo, (18°58’25” N, 101°43’03” W), located at the SW sector of the Trans-Mexican Volcanic Belt (TMVB) was reported and described by local governors and by the popular tradition at that time. Recently, country rocks, and pyroclastic deposits and lava flows from the eruption have been studied to establish the tectonic setting and previous volcanic activity in the region of Jorullo (Guilbaud *et al*., 2011), and to explain its plumbing system and the mechanisms driving the eruption (Johnson *et al*. 2008, 2009, 2010).

The 1943 eruption of Paricutín was instrumentally observed and studied with various geophysical methods. The seismicity was detected and recognized in the continuous recordings of the National Seismological Service of Mexico, mostly by the main station in Mexico City, 320 km distant from the volcano. At the field, Mexican and USA scientists carried out several studies on the eruption (e.g., UNAM-

IGEOL, 1945; Atl, 1950; Yokoyama and De la Cruz-Reyna, 1990; Luhr and Simkin, 1993). The location of Jorullo and Paricutín volcanoes is shown in Figure 1.

The formation of two small basaltic maars behind the Aleutian arc near Peulik volcano in late March and early April 1977, the Ukinrek maars, is one of several known eruptions of this particular, short duration, monogenetic type in historic time. This is among the best documented of the about eleven known maar eruptions in the 20th century (Kienle *et al*, 1980; Self, *et al*., 1980).

Another recent monogenetic eruption occurred in the NE part of the Izu Peninsula, Japan, though the eruption did not result in the formation of a new cone. In that region many monogenetic cones and vents are distributed on the land and in the sea. In 1930, earthquake swarms and crustal deformations, probably of volcanic origin, were detected for about four months, but no eruption occurred. In 1989, a shallower earthquake swarm was recorded in the same area, followed by a small submarine eruption. The geophysical evidence indicates that this eruption was probably derived from remnant magmas that originated from subduction volcanism and stalled in the crust. This eruption may represent a final stage of the monogenetic volcanism in the area. A limit case of this type of magmatism with low ascent power is the "failed" eruptions, with no external manifestation other than low level, spatially confined seismic swarm activity and/or minor phreatic activity.

Finally, the formation of monogenetic cones of another type, parasitic cones of polygenetic volcanoes, is interpreted in terms of the mechanical behavior of polygenetic volcanic structures.

## Growth patterns of scoria-cone, lava-field forming eruptions: Jorullo and Paricutín

Considering the date of the eruption and the relatively isolated conditions of the Michoacán area in the middle of the 18th century, the growth of Jorullo is reasonably well documented, except for the earliest stages. Jorullo is composed of a large scoria cone and several aligned satellite cones. During the eruption, a main cone and four or five smaller cones formed along a SSW-NNE fissure. Estimated erupted volumes of lava and tephra are in the ranges 0.36–0.5 km$^3$ and 0.54–1.5 km$^3$, respectively (Luhr and Carmichael, 1985; Rowland *et al*., 2009).

The Paricutín cone was fed through a main vent, and to a lesser extent through a second and third vents that eventually formed Sapicho, and Taqui, as shown in Figure 2a (after UNAM-IGEOL, 1945). This characteristic of fissure forming aligned volcanoes is typical of scoria cones, contrasting with lava-dome forming eruptions. Figure 2b shows the growth curve of the Showa-shinzan lava dome formed in 1944, Japan, after Yokoyama (2004) who analyzed the results of precise levels along the foot of the dome. While Paricutín grew both vertically and horizontally by the lava and scoria production of several aligned vents, Showa-shinzan was formed by upward thrust of magmas from a single vent.

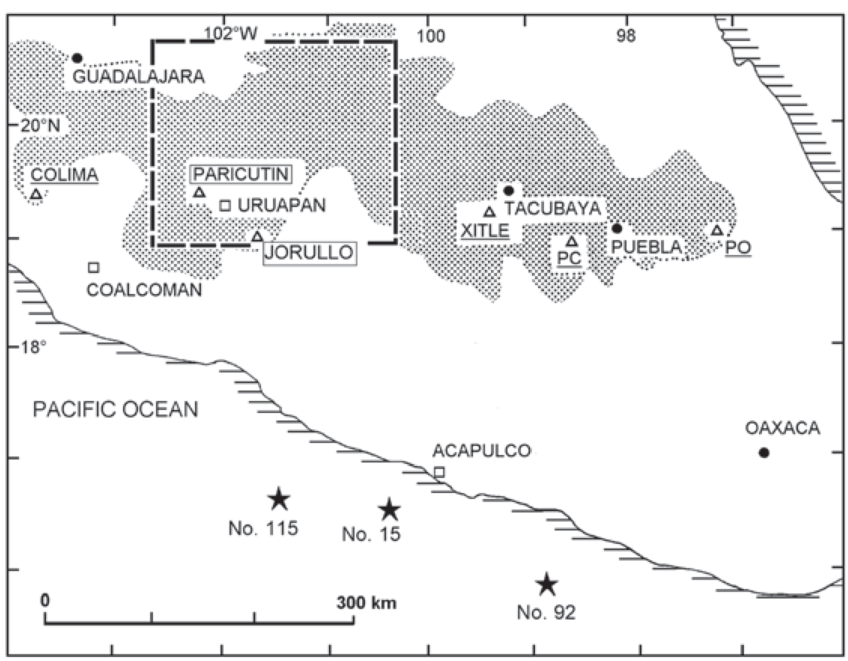



**Figure 1.** Location map of Jorullo and Paricutin volcanoes. Dashed rectangle: Michoacán-Guanajuato Volcanic Field; Dotted area: Mexican Volcanic Belt; PC: Popocatépetl, PO: Pico de Orizaba, Star marks: Epicenters of subduction earthquakes in February 1943. (Covarrubias, 1945b): No. 15 (Feb. 16, M 4.9), No. 92 (Feb. 20, M 6.0) and No. 115 (Feb. 22, M 7.7).

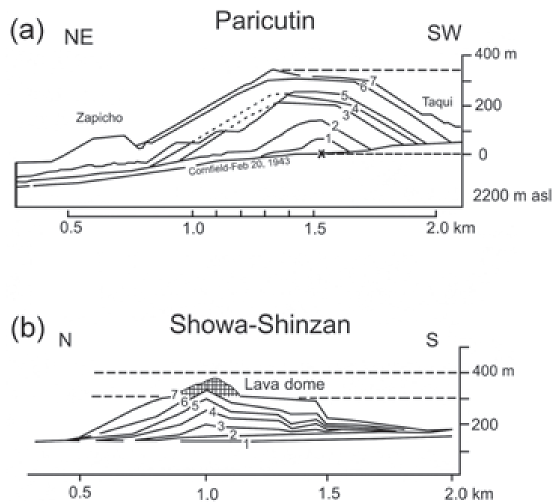



**Figure 2.** (a) Growth profile of Paricutín (UNAM-IGEOL. 1945, Appendix). (1): Feb. 23, 1943, (7): Feb. 20, 1944. (b) Growth profile of lava dome Showa-shinzan, Japan (Yokoyama, 2004), (1): Apr. 27, 1944, (7): May 18, 1945.

## Seismic activity related to scoria-cone volcano births

### *Seismicity at Jorullo*

Indeed seismographs did not exist in the 18th century, when Jorullo was formed in Michoacán, Mexico. Seismicity forerunning the eruption of Jorullo in 1759 can only be documented on the base of reports of felt activity. Around the birthplace of Jorullo, rumblings of the ground were heard since the end of June of 1759. Shocks were felt in July and gradually intensified. On Sept. 28, three or four sharp tremors were felt, and the following day, violent pyroclastic emissions began from a new vent, that continued for 3 days forming a cone that later became a satellite (Volcancitos) of a main vent. In this case, the first eruption began about 90 days after the first clear precursory seismic events. After that, we have no records or documents on the activity of the volcano until 1764 when the young volcano manifested the highest eruptive strength, and lava flows first appeared. According to Gadow (1930), most of the lavas were probably erupted between 1760~1766. After 1774, the activity ended.

### *Seismicity at Paricutín*

In the 1940's the Mexican National Seismic Network (SSN) operated using mainly seismographs of the Wiechert type. The closest station to Paricutín was about 180 km NW in Guadalajara. It was equipped with a seismograph of low magnification (80). The Main Central Station was located about 320 km to the E of the volcano in Tacubaya, Mexico City, with two Wiechert seismographs (17,000 and 1200 kg) with higher magnification (2000). Before and during the 1943 eruption of Paricutín, earthquakes with magnitudes *Ms* > 3 were clearly detected by the SSN Tacubaya station (Flores-Covarrubias, 1945a, 1945b; Yokoyama and De la Cruz-Reyna, 1990), while small precursory tremors were reported by the inhabitants. Flores-Covarrubias (1945b) and his group from UNAM installed a vertical Mintrop-Weichert seismograph (natural period 0.9 s), with optical recording in photographic film in a provisional dark cabinet built at 1077 m from the active crater. This instrument operated between March 5 and March 9, 1943. During that time no volcanotectonic events were detected, and only high frequency tremor signals with predominant periods grouped in two bands, 0.1 - 0.2, and 0.35 - 0.60 s were recorded.

The larger precursory volcanic earthquakes occurring at the Paricutín area were first recorded at the Tacubaya station about 45 days before the eruption marking the birth of the volcano. The volcanic earthquakes were all of the high-frequency type. Low-frequency events were not recorded at the SSN probably because their amplitudes decayed at that large epicentral distance. Unfortunately, most volcano-tectonic earthquakes were recorded by one, or at most by two stations, making a precise location of the hypocenters impossible. The seismicity of the precursory stage has been discussed by Yokoyama and De la Cruz-Reyna (1990), who determined the first major precursory earthquake to have occurred at 17 h 40 m, Jan. 7, 1943 (local time), with *Ms* 4.4 (No. 1 in Figure 3a). Considering its epicentral distance from Tacubaya, 322 km, it is very likely that lower-magnitude seismicity below the detection threshold of the SSN seismographs preceded this event.

According to Trask (1945), unrest among the population began in early February, 1943 when frequent earthquakes of volcanic origin which increased rapidly in number and intensity were felt at the site on Feb. 5, 1943. More than 300 tremors were felt the day before the Paricutín outburst occurred, 45 days after the first strong shock, and 15 days after the earthquake swarms were first felt, while the Jorullo eruption occurred 90 days after the initial rumblings of the ground. The precursory seismicity periods at both volcanoes were thus relatively long. Coincidentally, important subduction earthquakes occurred at the Pacific coast of Mexico before and after the outburst of Paricutín. Flores-Covarruvias (1945b) analyzed the seismograms of Tacubaya, Guadalajara, Manzanillo, Puebla, Oaxaca, Veracruz, and Chihuahua stations, and the epicenters of the largest three are shown in Figure 1. The chronology of the events after Feb. 16 is as follows (The earthquake numbers were assigned by the SSN at that time; all times local):

No. 15: Feb. 16, 18 h 36 m, *Ms* 4.9. Subduction earthquake.

No. 20 (of Paricutín earthquakes): Feb. 18, 18 h 39 m, an earthquake *Ms* 4.5 occurred at the Paricutín area. This was one of the large magnitude precursors of the Paricutín eruption.

Feb. 20, around 16 h, first Paricutín outburst.

No. 92: Feb. 20, 18 h 08 m, *Ms* 6.0, (most likely a subduction earthquake).

No. 115: Feb. 22, 03h 21m, subduction earthquake *Ms* 7.7, hypocentral depth 51.4 km. Epicentral distance from Tacubaya, 382 km. Seismic intensity VII (Mercalli) in Mexico City. This earthquake was followed by many aftershocks that made the identification of the Paricutín earthquakes difficult. Although triggering of volcanic eruptions by large tectonic earthquakes is possible, no evidence of this was found in the present case. In 1941 and 1942, before the outburst of Paricutín, earthquakes of magnitude 3.5 < *Ms* < 4 occurring in the area were also registered at Tacubaya. While these may not be directly related to the Paricutín activity, it is difficult to determine their origin.

Magnitudes of the precursory earthquakes *Ms* > 3 preceding the Paricutín eruption and cumulative energy release are shown in Figure 3a and b. In the precursory stage, the volcano released a total seismic energy $2\times10^{12}$ Joule corresponding to a single earthquake *Ms* 5. Yokoyama (1988) has empirically concluded that eruptions in andesitic volcanoes occurring after relatively long quiescence periods release a total cumulative seismic energy of order $10^{10\sim11}$ Joule. In particular, Paricutín earthquake No. 20 of Feb. 18 (marked in Figure 3a) which was also studied by Flores-Covarrubias (1945b) in seismograms of the Tacubaya and Guadalajara stations, and was assigned a hypocentral depth of 39.6 km, can provide valuable information about the monogenetic volcanism process.

Assuming a simple 2-layer crustal structure, with an upper layer of thickness *H* (40 km) and p-wave velocity *V1* (6.2 km /s) lying on a half-infinite layer with velocity *V2* (8.2 km /s), the refracted waves *Pn* would arrive at the observatory (320 km distant) earlier than the direct wave *Pd*, and the *(Pd - Pn)* values would depend on the depths of hypocenter, the larger the nearer to the crustal bottom. The (*Pd - Pn*) value for surface earthquakes is 4.4 sec while that for earthquakes of 40 km-deep is 8.5 sec. The (*Pd - Pn*) values of the 12 precursory earthquakes are found to be in the range between 4 and 7 sec, with some errors (Figure 3c). We may thus say that some of the precursory earthquakes were rather deep, near the crustal base, and that no systematic variation with time can be found, as the depths vary widely before the outburst.

Monthly number of volcanic earthquakes, monthly maximum magnitudes, and annual release of seismic energy for the period of 1943 to 1953 are shown in Figure 4. The figure shows that the seismic activity of Paricutín gradually

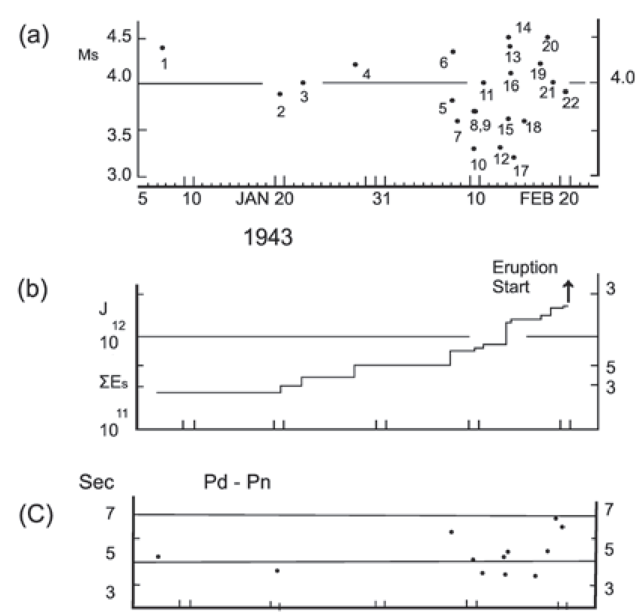



**Figure 3.** Precursory earthquakes of the 1943 eruption of Paricutin: (a) Earthquakes with magnitude $M_s$ > 3 preceding the onset of the eruption. (b) Cumulative earthquake energy before the outburst, (c) Temporal change of seismic phases ($P_d$ - $P_n$).

decreased after 1947. However, according to Wilcox (1954), from June to November 1951, the effusion of lava was large, and during January and February 1952, the effusion of lava was moderate to large, ceasing altogether on Feb. 25, 1952. The annual release of lavas from Paricutín is $40\times10^6$ m$^3$ and $43\times10^6$ m$^3$ for 1950 and 1951, respectively. Some unexpected phenomena occurred in the final stage of the eruption: The eruption rate in 1951 increased and exceeded that of 1950 and 5 earthquakes with magnitudes near *Ms* 4 occurring in April 1952 were conspicuous. Two of them were felt in the Paricutín area. Since one may expect that volcanic seismicity would decrease as surface activity declines, the earthquakes in April of 1952 may reflect a readjusting of the crustal stress caused by lava effusions.

*Magnitude distribution of Paricutín earthquakes*

The distribution of earthquake magnitudes in a region can be described by the frequency-magnitude distribution based on the Ishimoto and Iida (1939) and Gutenberg-Richter (1941) laws:

$$log\ N(M) = a - bM \qquad (1)$$

where $N$ is the cumulative number of earthquakes with magnitude ≥ $M$, and $a$ and $b$ are constants that describe the power law decay of occurrence frequency with increasing magnitude. The $b$-value is the slope of the straight line that fits well the earthquake-magnitude distribution data above the magnitude of completeness $Mc$, defined as the lowest magnitude at which nearly 100% of the events are detected.

The $b$-value strongly depends on the physical properties of the seismic medium and on the stress source causing the earthquakes. Areas where seismicity is caused by extensive regional stresses usually show $b$-values near or under 1.0 (Frohlich and Davis 1993). When stress concentrates causing clustering of seismicity and fracturing over distances of a few kilometers the value of $b$ may change significantly (Ogata and Katsura 1993); particularly, stress concentrations produced by sustained thermal gradients may increase $b$ to values as high as 2 or 3 (Warren and Latham 1970). The value of $b$ also provides important information related to the dimensions of the faults causing the earthquakes, since their magnitudes are directly related to the fault dimensions. Recording the $b$-value is thus equivalent to recording the mean magnitude, i.e. the mean crack length (Wiemer and McNutt 1997; Wiemer *et al*. 1998). The calculated $b$-value for the $Ms$ ≥ 3 Paricutín earthquakes detected in the period 1943~1951 at Tacubaya by the maximum likelihood method is about 0.75. Although this estimate is based on a data set clearly incomplete for magnitudes below 3, its low value is probably indicative of the absence of large thermal gradients extending over significant distances in the crust beneath the volcanic area and a relatively homogeneous crustal structure, suggesting the absence of a long-lived, large magmatic crustal reservoir prior to the eruption.

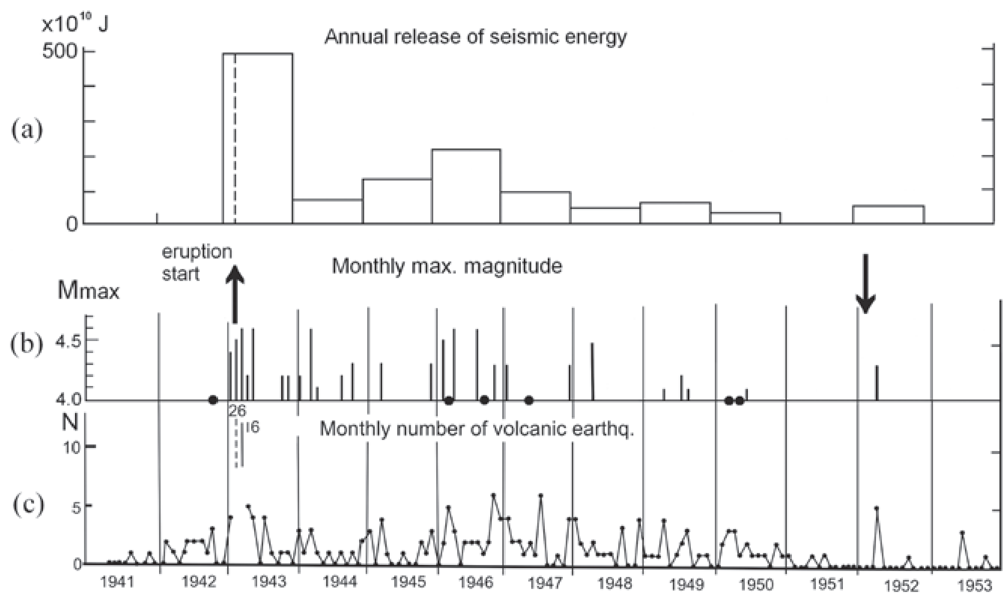



**Figure 4.** Characteristic of the seismicity associated to the 1943~1952 eruption of Paricutín: (a) Annual release of seismic energy, (b) Monthly largest magnitude of volcanic earthquakes with magnitude ($M_s$ > 4.0), (c) Monthly number of volcanic earthquakes ($M_s$ ≥ 3.0) recorded at Tacubaya, Mexico City ( $\Delta$ = 320 km).

*Subsequent seismic activity near Paricutín in 1997 and 2006*

More recently, other seismic swarms have been detected in the area. In March 1997, 230 earthquakes were detected between Paricutín and Tancitaro, a large quaternary stratovolcano about 10 km to the SW of Paricutín. Pacheco *et al*. (1999) studied this activity and concluded that the swarm was of tectonic origin and caused by local fracture systems related to the San Juanico-Buenavista fault, which is part of the Chapala-Oaxaca Fault system. Unfortunately, no *b*-value was reported for that swarm.

Between May and July 2006, another major seismic swarm was detected in the same area, with most of the epicenters located less than 10 km form the 1997 swarm. More than 700 earthquakes were studied by Gardine *et al*. (2011). Using a completeness magnitude around 2.6, Gardine *et al* (2011) calculated a *b*-value of 2.45, which they relate to a dike emplacement whose ascent stalled at a depth around 4-5 km, and did not reach the surface. This "failed" eruption was probably caused by a stalling of the vertical displacements of the magma, forming an inflating sill that acted as a small reservoir, inducing with this process a high $b$-value of the local seismicity.

## Magma ascent, and production rates of Jorullo and Paricutín

As stated above, the total volumes of lavas and pyroclastic material ejected at Paricutín for the period 1943 to 1951 were estimated at 1.9 km$^3$ (DRE) by Fries (1953) who determined the density of lava as $1.9\times10^3$ kg/m$^3$ and the density of tephra as $1.7\times10^3$ kg/m$^3$. The volumes of lava, tephra and DRE volume of magma emitted per year are depicted in Figure 5. The annual magma eruption rate decreased exponentially as shown by the diamond markers and by the best fit inclined solid line in Figure 5, both referred to the right-hand axis logarithmic scale.

Machado (1974) proposed a simple model to explain the exponential decay of the magma eruption rates reported at Vesuvius, Kilauea and Fayal. This model assumes a confined and pressurized magma reservoir connected to the surface by a conduit through which magma is transported in a Poiseuille regime. The discharge obtained from that model is:

$$Q = Q_0 \, exp\,(-\lambda t)\,, \qquad (2)$$

where $\lambda$ depends on the conduit geometry, the bulk modulus of the lava and the volume of the magma reservoir. Scandone (1979) estimated the overall value of $\lambda$ of Paricutín (the slope of the inclined solid line in Figure 5) as 0.31, and used this model to analyze the mean annual release rates of mass and energy during the eruption. He found two distinct regimes of the energy release revealed by the difference between the cumulative energy of the eruption and the annual energy release. After four years of the eruption onset, a reduction of the mean discharge rate indicated an increase of viscosity probably caused by differentiation of the magma and enhanced petrochemical reactions with the country rock. Such increased interactions with the crustal rocks probably resulted from a persistent decrease of the ascent velocity, as the primary magma source became depleted. We believe that the higher release of seismic energy from the volcano in 1946 (Figure 4a) may be related to that change, probably through the greater magma-host rock mechanical coupling caused by the increasing viscosity. Later, after 1948, the seismicity again decreased as the ascent velocity waned. We briefly discuss the reference value of the magma viscosity in the next section.

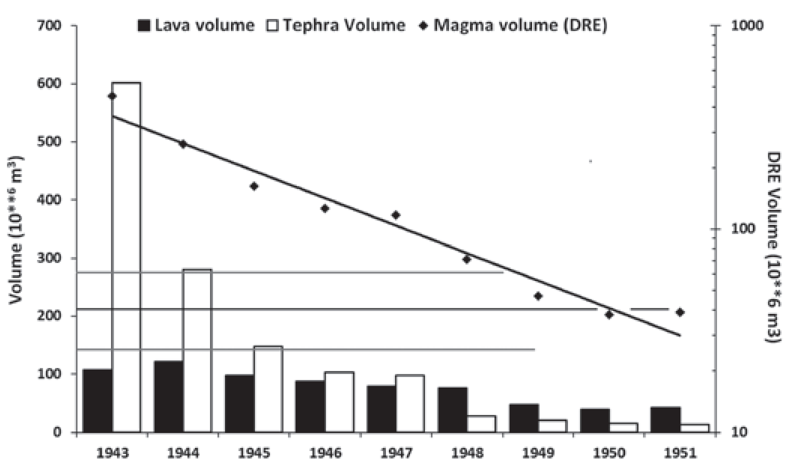



**Figure 5.** Annual volumetric release of lava (black columns), tephra (white columns) and magma (diamonds and inclined black line) of Paricutin, estimated with densities $1.9\times10^3$ kg/m$^3$ for lava, $1.7\times10^3$ kg/m$^3$ for tephra and 2.7 for DRE magma, after the original data of Fries (1953). The solid black horizontal line represents the 9 year average of the Paricutin annual production rate of magma, and the horizontal grey lines mark the range of annual rates estimated for 7 years of the Jorullo eruption.

The exponential decrease of magma eruption rate means that the magma source was not refilled by new magma after the eruption started. The magma eruption rate averaged for the period 1943 to 1951 is calculated as 1.9 $km^3$/9 yr = 211×$10^6$ $m^3$/yr = 6.7 $m^3$/s. This average is represented by the solid horizontal line in Figure 5. On the other hand, the short-term rate for the first year of the eruption, Feb. 20 to Dec 31, 1943, using Fries' data (1953, Tables 4 and 5) is estimated at 0.60$km^3$(DRE)/315days, that amounts to 0.69 $km^3$/y, or 22 $m^3$/s.

Although magma production estimates during the 1943 eruption of Paricutín are well determined, those of the 1759 eruption of Jorullo are not so well defined. According to Luhr and Carmichael (1985), the Jorullo eruption began in 1759, and the activity was strongest in 1764, continuing with lower intensity until 1774. Luhr and Carmichael estimated the total volume of ejected magma as 2 $km^3$ (DRE), from which lavas contributed 0.50 $km^3$. More recently, Rowland *et al*. (2009) estimated a total of 0.9 $km^3$ (0.7 $km^3$ magma DRE). We thus adopt the typical value of the magma produced at Jorullo in the range 1~2 $km^3$ in order to estimate the emission rate. Gadow (1930) reported that most of the lavas were probably erupted between 1760 and 1766. We may thus estimate the 7 year average of the yearly magma production of Jorullo to be in the range 1~2 $km^3$/7yr = 143~286×$10^6$ $m^3$/yr = 4.5~ 9 $m^3$/s. This range is represented by the grey horizontal lines in Figure 5, superimposed to the Paricutín data in order to compare with the rates of Paricutín. The averaged eruption rates of the two volcanoes are thus similar.

Table 1 shows the mean magma production rates of these and other monogenetic volcanoes described in a section below.

Hasenaka and Carmichael (1985) estimated the total magma production rate of the highest cone density portion of the Michoacán-Guanajuato volcanic field, comprising Paricutín volcano (centered at about 250 km from the trench, and with an extension of approximately 1000 $km^2$, Figure 1), as 0.12 $km^3$/1000 yr, i.e. 3.8 × $10^{-3}$ $m^3$/s. This area includes about 110 monogenetic volcanoes that may be assumed to have a common deep magma source and similar magma transport mechanisms.

*Viscosity of magma*

Viscosity strongly depends on the temperature, silica, and volatile and crystal contents of magma. In laboratory measurements, it is extremely difficult to control all of these factors in a single experiment. To explain some features of the monogenetic volcanoes, we use here the "macroscopic viscosity" proposed by Yokoyama (2005). He assumed in a first approximation that temperature, and volatile and crystal contents were similar in the lava domes, and correlated the magma eruption rates of 15 lava domes worldwide with their silica contents. He found that the viscosity depends on the silica contents according to the empirical relationship

$$\log \eta = 0.087 \times (\mathrm{SiO_2}\,\%), \tag{3}$$

The magmas of Jorullo and Paricutín contain 52 and 56 % of $SiO_2$, respectively. Then their viscosities may be estimated as 3.3 and 7.4 ×$10^4$ Pa·s, respectively. Krauskopf (1948) measured the viscosity of Paricutín lava flows during the winter of 1945~1946 applying the Jeffrey's formula, and obtained viscosity values between $10^4$ and $10^5$ Pa·s at temperatures in the range 1050 ~ 1070 °C. We thus may assume the viscosity of the Paricutín pristine magma is of order $10^5$ Pa·s.

**Table 1.** Mean magma production rates of some monogenetic volcanoes.

| Volcano | Ejecta | Duration | Volume $km^3$ (DRE) | Rate $m^3\ s^{-1}$ | Ref. |
|---|---|---|---|---|---|
| Monte Nuovo (1538) | P | 7 d. | 0.04 | 66 | Di Vito *et al.* (2009) |
| Wadalianchi (1719~21) | L, P | 3 yr. | 1.1 | 35 | Feng and Whitford-Stark (1986) |
| Jorullo (1759~66) | L, P | 7 yr. | 2 | 4.5-9 | |
| Paricutin (1943~52) | L, P | 9 yr. | 1.9 | 6.7 | |
| Waiowa (1943~44 ?) | Ash | 8 mos. | (small qty.) | | Baker (1946) |
| East-Izu (1989) | Pumice | 1 h. | (very small qty.) | | Oshima *et al*. (1991) |

**L: lava, P: pyroclastics**

*Source of magma*

The discussions on precursory seismicity and the absence of significant deformations in other sections supports the assumption that the main source of magma in volcanoes like Paricutín and Jorullo is deep, in the mantle, probably below the base of the crust, and that no crustal magma reservoirs are formed before, or in the early stages of the eruption. Wadge (1981) concludes that the long half-value period of the Paricutín lava effusion is compatible with a deep reservoir of large volume behaving as a closed system during the eruption. Also Luhr and Carmichael (1985) modeled the crystal fractionation event during the eruption of Jorullo, and concluded that the event must have occurred at lower crustal to upper-mantle pressures (800~1500 MPa or 30~60 km deep).

There are different possible causes to melt the mantle material, namely, the shear dissipation generated by a subduction process in the mantle wedge, the increased water (and other volatiles) content related to the materials carried by the subducting plate, and other more complex petrologic processes. These causes have been amply debated in the literature since the paradigm of plate tectonics was developed in the second half or the past century (see for example Best, 2003), and we thus limit our discussion to some physical aspects of the magma genesis relevant to the nature of monogenetic volcanism.

Hasenaka and Carmichael (1985) determined the ages of 78 volcanoes in the field as younger than 40,000 years by morphological methods calibrated by radiocarbon dating. They also estimated the overall cone density younger than 40,000 years B.P. as 2.5 cones /100 km$^2$ in the southern part of the Michoacán-Guanajuato volcanic field. The distance between Paricutín and Jorullo is about 80 km; both volcanoes can thus be included in an area of 100 km × 100 km. If we extrapolate the above cone density to this area as 250 cones/10$^4$ km$^2$, we obtain a frequency of monogenetic volcanisms in this area of 250 cones/40,000 yr = 0.0063/yr, or one cone-forming eruption every 160 years. This estimate is consistent with the 184 years interval between the eruptions of Jorullo and Paricutín. Hasenaka and Carmichael (1985) map of the volcanoes in the Michoacán-Guanajuato volcanic field, and their distances from the trench is reproduced in Figure 6. According to them, the distribution of volcanoes suggests that the Michoacán monogenetic volcano group has a direct relationship to the subduction process. Assuming that magma formed in this process tends to accumulate under the continental crust, a condition for the magma to ascend is that its pressure is high enough to open pathways through the crust. The pressure and temperature at which the phase transition occurs are governed by the Clausius-Clapeyron equation:

$$\frac{\Delta P}{\Delta T} = \frac{L}{T(V_l - V_S)} \tag{4}$$

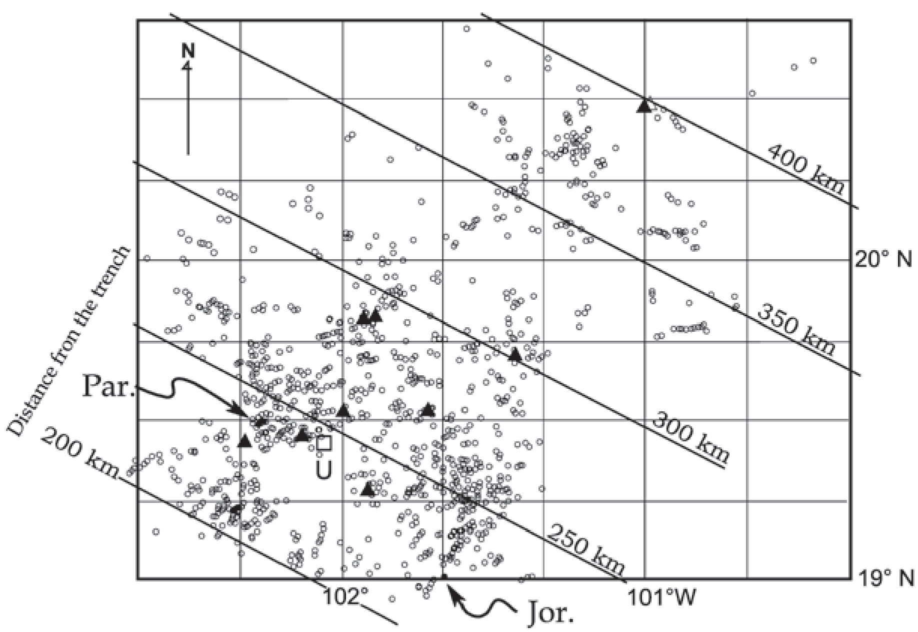



**Figure 6.** Distribution of volcanoes in the Michoacan-Guanajuato Volcanic Field (Hasenaka and Carmichael, 1985). Oblique lines indicate the distances from the trench. U: Uruapan City. Triangles indicate volcanoes with volumes larger than 10 km$^3$.

where $L$ is the heat of fusion, $T$ the melting point temperature, and $V_l$ and $V_S$ denote specific volumes of the liquid and solid phases respectively, and take as representative the values of 0.385 and 0.346 $cm^3$/gm respectively, based on published data (Dane, 1941; Gröschel-Becker *et al*., 1994, Ahrens, 1995).

Taking $\Delta P$ as the critical strength of the crustal rocks, which is of order $10^8$ Pa, using a typical value for the heat of fusion for MORB basalts $Lb$= 500 kJ/kg (Kojitani and Akaogi, 1995), and estimating $T$ as 1500°K, the required rise of melting point temperature is $\Delta T$ = 12°K. This means that magma with enough pressure to overcome the mechanical resistance of the crust may be formed if the mantle temperature is only about 12°K higher than the "normal" temperature.

Using again as a typical value for the heat of fusion 500 kJ/kg, the heat required to melt a cubic kilometer of mantle material is about $1.5\times10^{18}$ J. The amount of molten magma in the mantle that accumulates forming a layer at the base of the crust should be at least an order of magnitude larger than the magma reaching the surface. Thus for the typical volumes of 1~2 $km^3$ emitted by monogenetic eruptions like Paricutín and Jorullo, the heat required to melt a volume of say 20 $km^3$ is about $3\times10^{19}$ J. The additional heat flow required to raise the temperature of that region additional 12°K and to generate a pressure gradient high enough to make it reach the surface is

$$\Delta H = \rho C_p V \Delta T / A\tau , \qquad (5)$$

where $\rho$ ~ 3000 kg/$m^3$ is the density at depth, $C_p$ is the specific heat of magma, taken as 1.52 KJ/Kg°K (Bouhifd *et al*., 2007), $A$ is the area of the magma accumulation (~$10^4$ $km^2$) and $\tau$ the mean recurrence period of monogenetic eruptions (~ 200 yr). With these values, the excess heat flow necessary to increase the temperature by $\Delta T$ of the above volume of the material is $\Delta H$=17 $mW$/$m^2$, which is less that 20% of the mean global heat flow.

*Magma transport*

The way magma opens its paths to the surface is complex, and conduits probably form in intricate shapes and spatial patterns. The detailed description of such process is beyond the scope of this paper, and we limit our discussion to only two aspects of the process: the energy required and the time scale. As mentioned above, Yokoyama (1988) established that eruptions after long quiescence periods release a total cumulative seismic energy of order $10^{10\sim11}$ Joule, and interpreted this value as the seismic energy required to open magma pathways through the crust. The available evidence points that the seismic energy released before the Paricutín outburst exceed that range by an order of magnitude, suggesting a longer crustal path and thus a deeper magma source. The lower bound of the time scale required for such process may be inferred from the times at which significant precursory seismicity was reported or detected. From the available information in the cases of Jorullo and Paricutín this time scale may be of the order of weeks to months (see section 3.2). These pathways probably form an array of dikes that eventually evolve into a small number of conduits, or a single one, at least near the surface.

Referring to the annual release of the Paricutín ejecta DRE shown in Figure 5, Scandone (1979) estimated the half-value period $1/\lambda$ as about 3 years. In the figure, the small abovementioned deviation from an exponential decrease is visible around 1947. We remark that the increase in release of seismic energy in 1946 (Figure 4a) should reveal a relation between the magma and seismic energy release rates, probably reflecting the momentum transfer between the magma flow and the country rock.

In any case it is clear that the evolution of the magma eruption rate is quite compatible with the simple model of a deep magma reservoir without replenishment for the duration of the eruption, and a poiseuille flow between the reservoir and the surface driven by a pressure gradient caused by a phase change of mantle materials. To discuss some general characteristics of the magma flow through the duration of the eruption, in a first approximation we may consider two simple conduit shapes, a tubular vent, and a dike, both with uniform dimensions along their total lengths. The volume transported per unit time thus obeys the Hagen-Poiseuille's law for laminar flow:

$$Q = \frac{S}{k\eta}\left(\frac{\Delta p}{d}\right) \qquad (6)$$

where $Q$ denotes the discharge rate, $\Delta p$ the pressure increase at the base of the vent in excess of the lithostatic pressure, $\Delta p/d$ is the mean pressure gradient in excess of the lithostatic balance, $\eta$ the viscosity of the magma, and $d$ the vertical length of the vent. $S$ and $k$ are parameters related to the geometry of the conduit. For a tubular vent $S = r^4$, where $r$ is the radius of the vent and $k = 8/\pi$. For a dike, $S = h^3L$, where $h$ and $L$ are the horizontal width and length respectively, and $k$ = 12.

Applying this simple model to the eruption process, we may obtain at least some qualitative features of the flow, with the limitations derived from making some stern assumptions such as considering $S$ and $\eta$ constant or averaged values in Eq. 6. Based on a report by Trask (1944), in which he describes the vent formed in February 20 as a hole 3 m in diameter that formed during the first hours of "smoke" emissions, and afterwards (June 19) grew to 22 ~ 30 m in diameter. We may thus assume $r$ = 15 m as a representative value of the radius of the tubular vent along its total length. Similarly, 1943 reports by Foshag and Gonzalez-Reyna (1956) indicate that at the "Ahuan" stage of the eruption (November, 1944) they could see an active 0.5 m wide dyke near Taqui emitting lava. Assuming a dyke 1500 m long (the approximate length of the line joining Sapichu, Paricutín and Taqui), an alternative to the tubular vent is a dike of width in the order of a few meters, and about 1500 m long in the horizontal direction. For Jorullo, we are not aware of any description of the dimensions of emission centers. Only for simplicity, we adopt here the same dimensions as in the Paricutín case.

To estimate the pressure gradient in excess of the hydrostatic gradient $\Delta p/d$ that is compatible with the observed magma rates and measured viscosity we consider some possible values of the pressure gradient in the range 0.5 MPa/30 km to 10 MPa /30km (16.7 Pa/m to 333 Pa/m), and plot the magma eruption rate ($Q$) against the viscosity of magma ($\eta$). Figure 7 shows this plot, with the paramenter $\Delta p/d$ taking the values 0.5, 1, 5, and 10 MPa/30km. The horizontal black horizontal line represent the estimates of the mean rates for the active periods of Paricutín, and the gray horizontal lines the range of mean rates estimated for Jorullo. Considering $10^5$ Pa s as a likely value for the viscosity, as determined in Section 4.1, pressure gradients in the range 1~5 MPa/30km (about 30 to 170 Pa/m) are the most consistent, assuming a 15 m radius tubular conduit, or a 1500 m long, 3 to 6 m wide dike. The above discussion is not inductive and the results are highly non unique, but it may set some constrains to the physical process of magma transport at cinder cone volcanoes such as Jorullo and Paricutín.

*Earth-surface deformations related to eruptions*

The 1943~51 eruption of Paricutín ejected volcanic material amounting to 1.9 km$^3$ (DRE), but no deformations around the volcano were reported. Probably the eruption did not cause any remarkable depression. A contrasting example is the 1914 eruption of the polygenetic volcano Sakurajima, in which the effusion of 1.6 km$^3$ of magma (DRE), a volume similar to the magma ejected by Paricutín, caused noticeable depressions of the earth-surface at the nearby Aira caldera over an area of 40 km across. Subsidence values were as high as about 1.5 m. Volume of

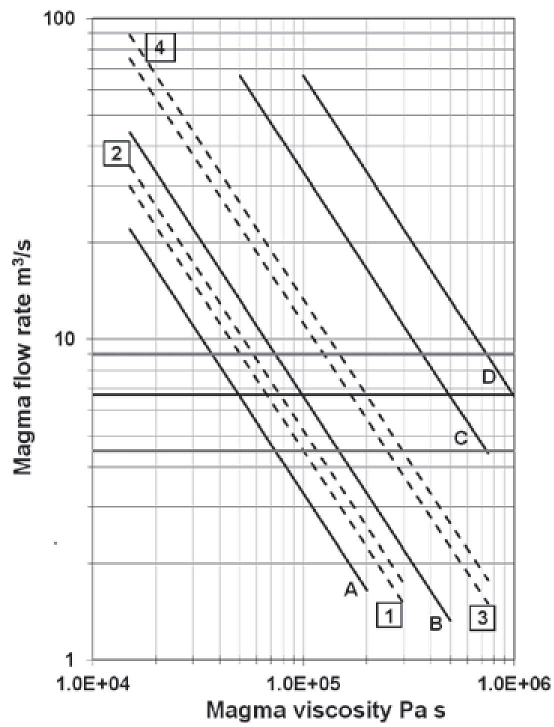



**Figure 7.** Magma flow rates as a function of magma viscosity, geometry of the conduit and likely pressure gradients according to Eq. 5. Black solid lines correspond to a cylindrical conduit with radius 15 m, and with pressure gradients, (A) 0.5 MPa/30km; (B) 1.0 MPa/30km; (C) 5.0 MPa/30km, and (D) 10.0 MPa/30km. Dashed lines represent the magma flow through dikes 1500 m long (horizontally), and (1) 6 m wide, with a pressure gradient 0.5 MPa/30 km; (2) 5 m wide, with a pressure gradient 1.0 MPa/30 km; (3) 3 m wide, with a pressure gradient 10 MPa/30 km, and (4) 4 m wide, with a pressure gradient 5 MPa/30 km. The solid black horizontal line represents the mean annual Paricutin magma production rate, and the horizontal grey lines represent the range of the mean annual rate of the Jorullo eruption.

the subsidence amounted to 0.24 km$^3$ (roughly 15 % of effusion volume), and has gradually recovered along a century. The origin of the magmatic pressure source was determined to be located at a depth of about 8 km by deformation analysis. This means that the magma source of Paricutín should be significantly deeper than the Sakurajima source.

## Maars, "small", and "failed" monogenetic eruptions

Maars are craters caused by predominantly phreatic and phreatomagmatic explosive episodes of short duration lasting from few hours to few days. The role of groundwater is crucial in this type of activity which otherwise may have not reached the surface considering the small volume of juvenile magmatic products ejected in these eruptions. A strong magmatic influence was observed in the Ukinrek maars in which the phreatomagmatic phase coexisted with transitional phases of incandescent lava pooling in one of the maar craters and producing Strombolian activity (Self, *et al*., 1980). Nevertheless, the small volume of juvenile material (totaling about 1.5 million $m^3$) and the short duration of the activity (11 days) suggest that the available magma had a small ascent capacity, and a trend to stall. In this case, heat conduction from a stalled magma source had an important role in the explosive events. Such may be the case of the monogenetic eruption observed in the NE part of the Izu Peninsula, Japan. In that region many monogenetic cones and vents are found on the land and in the sea (about 75 and 50, respectively, Figure 8 after Aramaki and Hamuro, 1977). They belong to the East-Izu monogenetic volcano group that has erupted during the last 40,000 years. In this region, most monogenetic cones are separated from the Quaternary polygenetic volcanoes, in contrast with the Michoacán area in which monogenetic cones are mixed with shield and stratovolcanoes. Almost all cones of the East-Izu group are basaltic and generally small in volume, amounting about 2.1 $km^3$ DRE in a total of 75 cones on land (Aramaki and Hamuro, 1977). The largest one, Oomuro-yama (OM in Figure 8) is about 0.8 $km^3$ DRE. The other volcanoes are even smaller in volume.

Repeated earthquake swarms and a small submarine eruption are the current features of volcanism in this area. In March~May 1930, earthquake swarms reported around Ito City amounted to 5000 felt earthquakes, the largest with magnitude $M_s$ 5.9. The epicenters concentrated at the sea, E of Ito City. The epicenters are plotted in Figure 9 (Nasu, 1935). The earthquake swarms were accompanied by land upheaval, SE of the city. The upheaval had the characteristics of a volcanic eruption precursor, but no eruption occurred in 1930. Kuno (1954) presented a model of this event, shown in Figure 10a, in which he assumed cracks reticulately distributed in this area, and interpreted that a magma branch moved upward causing earthquake swarms and upheavals, but stopped midway. This event is called the 1930 Ito Earthquake Swarm. The crustal structure around Izu Peninsula, obtained by explosion seismic observations by Asano *et al*. (1982) is shown in Figure 10b, where the hypocenters of earthquake swarms for 1977~1979 are also included. This suggests that the East-Izu monogenetic volcano group originates from magma trapped in the granitic layer, as inferred by Kuno (1954).

After the 1930 Ito Earthquake swarm, the North-Izu Earthquake ($M_s$ 7.3) followed in November of the same year, and both the 1978~1979 earthquake swarms and an earthquake $M_s$ 6.7 in 1980 were also detected to the E of the Izu Peninsula. In July 1989, earthquake swarms started again, with a maximum magnitude 5.5 on July 9. Along with volcanic tremors, a submarine eruption occurred on July 13 at the northernmost part of the group, about 3 km N of Ito City (Figure 8). The eruption site coincided with the epicentral area of the 1930 earthquake swarms.

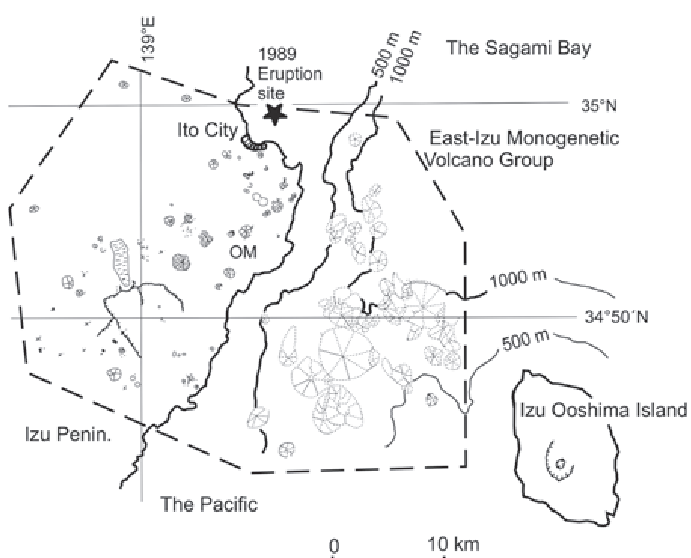



**Figure 8.** The East-Izu Monogenetic Volcano Group, Japan (Aramaki and Hamuro, 1977). It has about 75 volcanic cones on land and 50 under sea water.

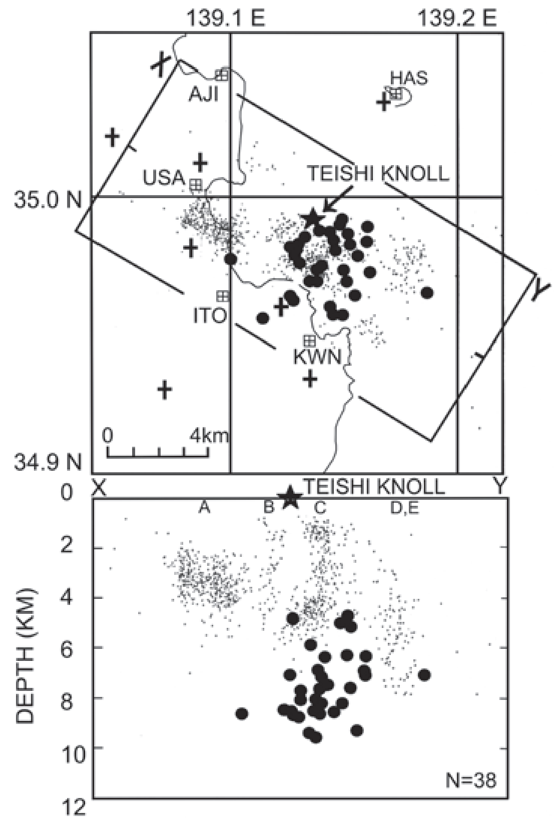



**Figure 9.** Distributions of earthquake foci of the 1930 and the 1989 earthquake swarms (Ueki, 1992). Solid circles: foci of the 1930 earthquake swarms. Dots: foci of the 1989 earthquake swarms. Squares with cross: Seismic observation points in 1930. Crosses: Seismic observation points in 1989.

The Hydrographic Department, the Maritime Safety Board of Japan (Oshima *et al*., 1991) repeated a topographic survey of the sea bottom using sonars, sparkers and Sea Beam before and after the explosion. After two weeks of precursory earthquake swarms, submarine explosions burst a few times within one hour in the afternoon of July 13 and ejected a small quantity of pumice. They succeeded in detecting the formation of a small knoll of about 25 m in swell, and 200 m in basal diameter, at a depth of about 100 m. According to Oshima *et al*. (1991), this was formed by upheaval of the sea floor, and it was named "Te-ishi knoll". This submarine eruption must be a culmination of internal magmatic activity developing since the 1930 earthquake swarm. It is very probable that the 1930 magma was re-activated in 1989 and scarcely reached the surface. This is obviously verified by the distribution of the hypocenters of both the earthquake swarms in 1930 and 1989 as shown in Figure 9 (after Ueki, 1992).

This is the first historic eruption in the East-Izu monogenetic volcano group, and at present, it may be interpreted in terms of a magma originated at depth, but stalled in the crust. This may be as well a characteristic of the "failed" eruptions that seem to have points in common with the cases of East Izu. If magma reaches a region of neutral buoyancy, the vertical migration of magma may stop, forming a sill as discussed by Gardine *et al.* (2011) regarding a dike emplacement in Michocacán, México, near Paricutín volcano. Depending on the depth of the sill, no external volcanic manifestation may occur as in the case of Michoacán, or heat conduction may overheat shallow groundwater causing phreatic explosions, without the emission of juvenile magmatic materials.—Furthermore, a decreasing ascending capability of magma manifested as the formation of sills below the eruption site may also clarify some aspects of the final stages of "major" monogenetic

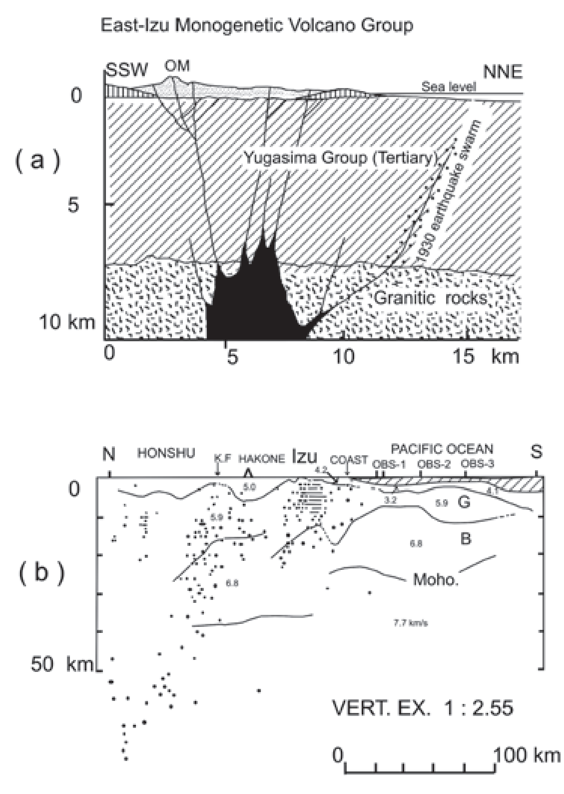



**Figure 10.** The East-Izu monogenetic volcano area. (a) A model of the 1930 earthquake swarm by Kuno (1953), (b) Crustal seismic velocity structure of Izu Peninsula obtained by observation of explosion earthquakes (Asano et al., 1982). The dots represent the cross sectional distribution of earthquake hypocenters during January 1977~November 1979.

eruptions. For instance, studying the subsurface structure of Paricutín, Wilcox (1954) interpreted a strong negative magnetic anomaly observed during an aeromagnetic survey in December 1947, centered at a point about 3 km NNW of the volcano, as a reservoir tapped on its flank at a depth of about 6 km. We are not aware of any previous magnetic study in the region, and there is no evidence that the magnetic anomaly was there before the onset of Paricutín activity. Assuming that the anomaly appeared in a late stage of the eruption, it may be explained in terms of an inflating sill resulting from the decreasing ascent ability of the magma caused by a decaying pressure gradient and decreasing buoyancy. The conditions for the arrest of a vertical propagating dike have been discussed by Taisne *et al*., (2011), and Moran *et al*. (2011), and a general discussion on maars and scoria cones has been recently published in the volume 201 (2011) of the Journal of Volcanology and Geothermal Research.

## Other historical cases of monogenetic eruptions

There are other examples of monogenetic volcano fields that have produced different types of activity in historical times. Without attempting to provide an exhaustive account, we briefly describe some of those events:

1) *Xitle*, Mexico City: 280 A. D.: It is a scoria cone rising 3120 m a. s. l., located on the S sector of Mexico City with a height of about 140 m above the surrounding ground and a basal diameter of 500 m. It belongs to the Chichinautzin monogenetic volcano field (Bloomfield, 1975). Although the initial datings by Arnold and Libby (1951), placed its formation 2422 BP± 250 yr BP, recent ages obtained from charcoal collected under the lava flows place the forming eruption at 1670±35 yr BP (Siebe, 2000). Its lavas flowed 15 km toward N and NE, covering about 72 km$^2$. The estimated volumes of lava and tephra from Xitle are 0.96 km$^3$ and 0.12 km$^3$ respectively (Cervantes and Molinero, 1995; Delgado *et al*. 1998).

2) *Cheju Island*, Korea, 1002 and 1007~1008 A.D.: An oval volcano island (75 km × 30 km) of basaltic rocks with more than 330 cones (scoria, tuff) and maars. At the center, the main cone Halla, raises 1950 m a. s. l. No precise reports on volcanic activities are available, and it is not clear which cones erupted in those dates.

3) *Monte Nuovo* in Campi Flegrei, Italy, 1538: An episode of several phreatomagmatic and magmatic explosions lasting 7 days, from Sept. 29 to 0ct. 6, formed a cone although no lava flows are reported (after Di Vito *et al*., 2009). The volume of the cone, estimated from a topographic sketch map in the Catalogue of the Active Volcanoes, Part XVIII (Imbò, 1965) is about 0.04 km3 (DRE). Then magma eruption rate was thus about 0.04 km$^3$/7 days = 66 m$^3$/s.

4) *Wadalianchi*, a monogenetic field in the NE part of China, 1719~1721. In an area 40 km × 25 km, there are 14 cones. Nangelaqiushan and Laoheishan cones formed in three years of activity and the lavas covered an area of 65 km$^2$. Feng and Whitford-Stark (1986) discussed the geology and petrography of these volcanoes. The magma eruption rate was about 1.1 km$^3$ DRE /3 yr = 35 m$^3$/s. These volcanoes are located at about 2000 km from the subduction zone of the Pacific coast.

5) *Canary Islands, Spain.* During the last 600 years, basaltic eruptions has been reported at the islands of La Palma (1430 ?), 1585, 1646, 1677, 1712, 1949, 1971), Tenerife (1492, 1704-05, 1706, 1798, 1909, this last one forming the Chinyero volcano) and Lanzarote (1730-1736 (Timanfaya), 1824) (Carracedo, 2007; Rodriguez-González, *et al*.; 2009; Sobradelo *et al*., 2011). All of those basaltic magma eruptions formed scoria cones and emplaced lava flows, having the typical characteristics of monogenetic volcanism.

6) *Waiowa*, 1943~1944: Located on the N flanks of Goropu Mountains, about 210 km E of Port Moresby, Papua. According to Baker (1946), on Dec. 27, 1943 this volcano ejected andesitic ash and steam to a height of about 4.5 km, and eruptions continued intermittently. On Aug. 31, 1944, ashes were observed at Port Moresby.

## A different kind of monogenetic activity: Parasitic cones on polygenetic volcanoes

Main cones of polygenetic volcanoes are generally fed through vents directly connected with shallow magma reservoirs. In their eruptions, magmas with high energy usually ascend through central vents that may remain open or sparsely closed after magmas stop moving or drain back to the depth. In the next eruption, if magmas have enough energy, they would ascend again through central vents, or otherwise, they would search weak points in lateral parts: these are the parasitic cones. After parasitic eruptions, magmas would solidify and fill up the cracks strengthening the crust around the secondary magma paths, and the magma of the next eruption would select new pathways to reach the surface. As far as the present authors know, parasitic volcanoes seldom erupt repeatedly.

Based on similarity principles, we propose here another type of monogenetic activity developing in a different scale as the monogenetic fields: the formation of parasitic cones and domes. In the latter case, it is important to clearly make a distinction with domes fed through main vents, which may be repeatedly emplaced and destroyed and is the case of Popocatépetl and Colima (Mexico), Tarumae (Japan) and many others. On the other hand, parasitic domes, such as Colle Umberto I (1895) on Vesuvius, Volcancito (1869-1878) on Colima and Showa-shinzan (1944) on Usu, Hokkaido are usually monogenetic, forming in a single cycle eruption. Magmas being fed through thin vents or cracks that are branches of main feeders would easily solidify with lavas after eruptions consolidating with the surrounding structures. Such parasitic eruptions may be considered as a different type of monogenetic volcanism, although comparatively lower in ejected volume when compared with scoria cone – lava field types such as Jorullo and Paricutín.

To illustrate these arguments, we present the example of Usu volcano in Japan, in which the vents formed in the 1910 eruptions were not reactivated in the 2000 eruption, and instead new vents formed. Figure 11 shows the distribution of parasitic vents and lava domes at the northern foot of Usu volcano. YS-hill is a mound formed in the 1910 eruption. On KP-hill, a simple mountain block, eight of the year 2000 phreatomagmatic or fumarolic vents opened just aside three of the 1910 phreatomagmatic vents, and did not reoccupy the older ones. KP-hill was injected by magmatic gases twice, in the 1910 and the 2000 eruptions: Tentatively we may say that the deeper parts of the 1910 conduits were closed and strengthened by the injected magmas (piling effect) and the 2000 magma could not reopen the 1910 conduits, while the KP-hill was not clogged by lavas. Such parasitic vents thus behave as a kind of monogenetic volcanism.

As shown in Figure 11, the parasitic vents are distributed roughly along a contour of 200 m above sea level or on a circle of about 2 km radius from the center of the volcano. When magma reaches beneath a volcano, it may be directed toward the weakest point, either the main crater above, or to lateral paths; the latter produces parasitic volcanoes or vents. Some mechanical arguments to explain this behavior, that may be applied the formation of parasitic craters in general, are briefly discussed next.

A criterion of fracture in material mechanics (Jaeger, 1964) for brittle materials, the "theory of maximum shearing stress", has proved to be of wide application in different experimental tests. In a plane stress configuration, the maximum shear stress equals the horizontal differential stress,

$$\tfrac{1}{2}\,(\sigma_x - \sigma_z)\,, \tag{7}$$

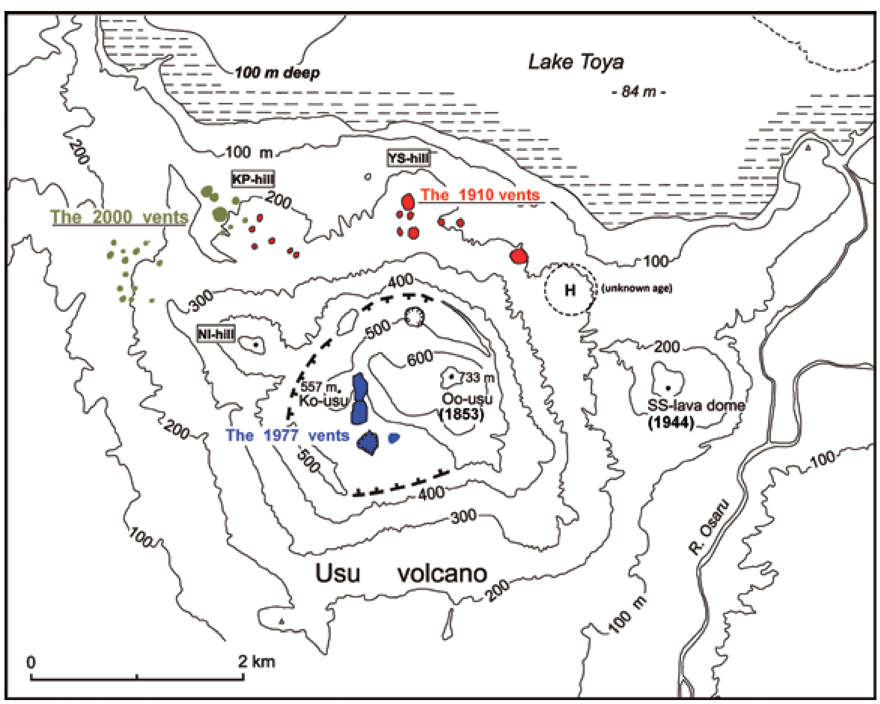



**Figure 11.** Parasitic vents at the foot of Usu volcano, Hokkaido: Red circular shapes: the 1910 vents. Green circular shapes: the 2000 vents. SS-lava dome: Showa-shinzan, formed in the 1944 eruption.

where $\sigma_x$ and $\sigma_z$ denote the principal stresses, i.e., the maximum shearing stress occurs across a plane whose normal bisects the angle between the largest and smallest principal stresses.

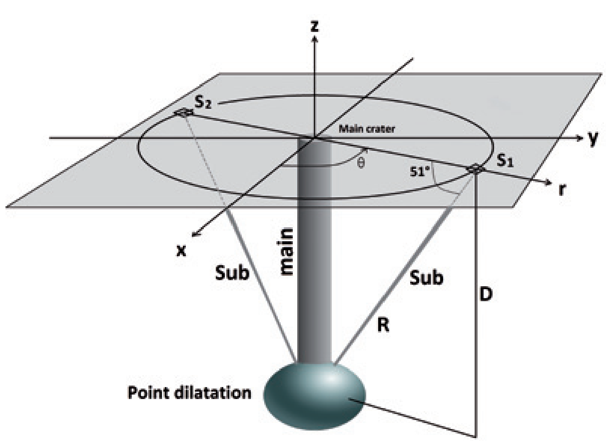
In the present discussion, we do not know how the principal stresses are distributed on each volcano. However, some general results may be obtained to explain the formation of parasitic vents.

We assume the simple model of a spherical pressure source of radius $a$, with its center at a depth $D$ in an elastic half-space. General solutions to the surface deformation when the pressure increment in the sphere is $P_0$ were first found by Sezawa (1931) for a general plastic medium. Mogi (1958) used his results to compare the deformation model with numerous field data of volcanic deformation. A simplified solution based on the assumption that $a/D$ <<1 is called the point center of dilatation and consists of a uniform expansion of a point-source. This expansion includes the quantity $a^3P_0$, a combination of two inseparable unknowns: the dimension of the sphere and the pressure increment, combination known as the strength (or intensity) of the singularity, since the stresses at the point dilatation are singular (McTigue, 1987).

The free surface displacements and the corresponding strain tensor components (using a combination of cylindrical and polar coordinates) are:

$$u_z = C_0 D / R^3, \text{ and } \quad \varepsilon_{zr} = -3C_0 D\, r / R^3 ,$$
$$u_r = C_0 r / R^3, \text{ and } \quad \varepsilon_{rr} = C_0\, (D^2 - 2\, r^2) / R^3,$$
$$u_\theta = 0 , \qquad \text{and} \quad \varepsilon_{\theta\theta} = C_0\, / R^3 , \qquad (8)$$

where $C_0 = (\lambda+2\mu)\, a^3P_0/2(\lambda+\mu)$, $r$ is the cylindrical radial distance on the surface from the epicenter of the sphere, $R = \sqrt{D^2+r^2}$ , and $\mu$ and $\lambda$ denote the Lame's parameters.




At the free surface, the vertical component of the stress tensor should vanish, $\sigma_{zz} = 0$.

Then, $(\sigma_x - \sigma_z)$ is expressed in polar coordinates $(r, \theta)$ as:

$$\sigma_{rr} - \sigma_{\theta\theta} = - 6\mu C_0 r^2 / R^5 , \qquad (9)$$

Independently of the value of $C_0$, equation (8) has maxima at:

$$r = \pm \sqrt{2/3}\;\; D \;\approx \pm\, 0.82\, D \qquad (10)$$

In other words, the medium undergoes a maximum horizontal differential stress at the radial distances $r = \pm$ 0.82 $D$, corresponding to a dip angle between the pressure source and the site of potential fracturing near the surface of 51°. This conformation is illustrated in Figure 12. At Usu volcano, assuming a horizontal distance $r$ = 2 km from the surface, we obtain the depth of the pressure source at $D \approx$ 2.4 km. This is a branch point of the lateral magma path, and in such cases, the magma pressure is not enough to break out at the summit crater through the main vent and may search weak points at lateral sides. Locations of parasitic cones thus depend on the position of the branch point from the main vent. Once parasitic cones have formed, the region around the subconduits is strengthened by piling modifying the principal stresses, and a new path may thus be formed.

Although the above results are approximate since in a real volcano the ground surface is not horizontal and the structure is heterogeneous, they prove to be another expression of the "isostatic surfaces produced by point dilatation beneath a free horizontal surface" of Anderson (1936, Figure 8). By this model he explained the formation of ring-dykes, sills, and cone-sheets.

## Discussion and concluding remarks

Why some volcanoes erupt repeatedly with repose periods of variable duration between eruptions, while others that have been named monogenetic show only one birth-eruption? The latter include an ample variety of magma compositions and eruptive types, such as basaltic scoria cones producing large lava flows, maars, and volcanoes so small that no permanent evidence is left on the surface. From the evidence of the

**Figure 12.** Geometrical distribution of new magma paths or subconduits (Sub) determined by the "maximum shear stress" induced by a "point dilatation" feeding the main crater, represented by the spheroidal shape only for illustration purposes. $S_1$ and $S_2$ are the points in which the formation of a new parasitic vent is more likely.

cases discussed in the present paper it seems that the primary cause of basaltic monogenetic volcanism is its deep subcrustal magma source, which has a limited volume and no capability of refilling, at least in the time scale of the duration of single monogenetic eruptions. The magma of such source directly reaches the surface through a complex conduit system that may include long series of pipes and cracks across all or most of the thickness of the crust; when the eruption ends, magmas filling the conduits and cracks would solidify increasing the strength (piling effect) of the crust beneath the volcano. Magmas of the next eruption would select different pathways to reach the surface. On the other hand, in polygenetic eruptions, magmas usually ascend from shallower magma reservoirs through shorter central vents that may remain open after magmas are ejected or drain back.

When the volume of the primary source involves a few tens of cubic kilometers, scoria cones and significant tephra and lava deposits are formed along several years of activity (e.g. Xitle, Jorullo, Paricutín). In the late stages of such eruptions, the pressure gradient driving the magma ascent wanes in the absence of replenishment sources and the magma may stall near the surface forming sills acting as a sort of relatively small short-life reservoirs. The longer residence time of the magma in such reservoirs may produce a compositional evolution of the erupted magma. Sometimes, the magma does not erupt at all, and the magma stalls in sills near the surface from the earliest stages of the process, leaving as a record only a characteristic seismicity as the main evidence of such "failed" eruptions. In some cases, surface activity may still be produced as weak eruptions from the sills, yielding minor volumes of magma (e.g., East-Izu volcanoes), or as phreatic explosions caused by overheating of water bodies or a combination of both (e.g., Ukinrek).

The ratio of the volumes of a dyke and a tubular pipe (such that they transport the same above-mentioned flow rate) is $V_{dyke}/V_{tube} \approx 10$, and the ratio of the lateral areas through which they lose heat to the country rock is $A_{dyke}/A_{tube} \approx 20$. Therefore, the dyke may contain a magma volume 10 times larger than the pipe, but its lateral area is almost 20 times larger, therefore dykes can cool faster larger volumes of magma favoring the stalling, cooling, and consequent strengthening of the crust. The stalling of deep-origin magmas in dykes and sills, the increased residence times, and the faster cooling leading to the closing and "sealing" or "healing" of the magma paths, that forces the magma at the base of the crust to open new paths is thus an equally important component of the process that makes a volcano "monogenetic".

A similar process associated to the shallow magma sources of polygenetic volcanoes can originate a different kind of monogenetic volcanism, the parasitic cones. In that case the main polygenetic vents are formed by direct, relatively short conduits connecting the crustal magma reservoir to the surface. When such main conduits clog, parasite vents may develop as thin sub-conduits formed as the result of fracturing caused by shear stress concentrations derived from the pressure in the main reservoir. Magmas associated with this monogenetic volcanism show evidence of long residence times in the crust and their composition is usually related to that of the associated polygenetic system (e.g., Monte-Nuovo–Campi Flegrei, Armienti, *et al*., 1983; Showa-shinzan –Usu, Yokoyama, *et al*., 1981; Tepexitl tuff ring – Los Humeros Caldera, Austin-Erickson, *et al*., 2011). Considering that the typical depth of a polygenetic magma reservoir is usually less than 12 km, parasites are commonly located at less than 10 km (i.e., $r \leq 0.82 \times 12$) from the main vents.

**Acknowledgements**

The authors have benefited by a book "Paricutín: The volcano born in a Mexican cornfield" compiled by Drs. J. F. Luhr and T. Simkin (1993). We thank Drs. M-N. Guilbaud, C. Siebe, V.-H. Garduño, G. Bocco and M. T. Ramírez and other organizers who spared no efforts to assemble the 250th anniversary of Volcán Jorullo's birth in Michoacán, México, held in September ~ October 2009. We are grateful to A. T. Mendoza-Rosas for her help with the figures.

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