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Global aspects of volcanism: the perspectives of "plate tectonics" and "volcanic systems"

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Abstract

The concept of plate tectonics provides a general framework to fit the distribution and general characteristics of volcanoes. There remain, however, many details of volcanic activity that are difficult to explain solely by this paradigm. For example, plate tectonics predicts that volcanic activity should take place continuously along all convergent and divergent tectonic margins, where in fact we observe a point-like distribution of volcanic centers that does not fit the predictions. Also, it is observed that many volcanoes share common characteristics despite being located in different tectonic settings, while other volcanoes sharing the same tectonic setting display very different behavior. For instance, so far, there is no congruent explanation offered by plate tectonics about why in similar tectonic conditions volcanism is sometimes polygenetic and elsewhere monogenetic. On the other hand, volcanic activity on a global scale tends to define a series of rules that are independent of the tectonic setting, and therefore should reflect general processes that are not controlled directly by the plate tectonic engine. We show that, by concentrating on the relationship among processes (from the moment of magma generation to the moment of eruption) and by incorporating all of these processes as components of a single system, a coherent picture of volcanism on a global scale emerges and allows us to interpret better many otherwise puzzling aspects of volcanism such as those mentioned above, hence providing a general framework that fills the conceptual gap left between plate tectonics and most (if not all) of the characteristics of volcanic activity at a global scale.

Keywords: Volcanic systems; Plate tectonics; Global volcanism; Polygenetic volcanoes; Monogenetic volcanism

1. Introduction

The distribution of volcanoes around the world has long drawn the attention of earth scientists. Attempts were made to explain such distribution from the mid-19th century onward, but it was only until the second half of the 20th century that the unifying concept of plate tectonics provided a general theory capable of explaining volcanism and other global phenomena such as mountain building and earthquake distribution (e.g., Vine, 1966; Le Pichon, 1968; Wilson, 1973). In this theory, volcanoes are associated with one of three tectonic scenarios defined by the relative movement of the rigid plates that cover the Earth's surface (Perfit and Davidson, 2000), and therefore they are

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classified as either: (1) rift volcanoes (where two neighboring plates move away from each other); (2) subduction volcanoes (where two plates move towards each other); or (3) hot-spot volcanoes (far from plate margins). This three end-member classification is adequate to explain the general distribution of volcanoes around the world, but there are various aspects of volcanism at a global scale that are difficult to explain based solely on the theory of plate tectonics. For example, the occurrence of volcanic centers that do not fit very easily in any of the three mentioned tectonic settings (Francis, 1995), the occurrence of monogenetic and polygenetic volcanism regardless of tectonic scenario (Macdonald, 1972; Nakamura, 1977), the observed point-like distribution of volcanic centers along the tectonic margins where volcanic activity is expected, and the presence of gaps in volcanic activity along these margins (Shimozuru and Kubo, 1983; de Bremond d'Ars et al., 1995). All of these hard-to explain volcanic features, however, seem to be the result of common rules controlling volcanism at a global scale; rules that are not included explicitly on the paradigm of plate tectonics.

As contended in this paper, these general rules encompass processes that range from magma generation to the moment when magma reaches the surface, and can be incorporated in a general model of volcanism termed "Volcanic System" (which as shown latter includes the magmatic system as one of its components). Although at present the concept of a "volcanic system" is used in a general context by most volcanologists, the formal definition and treatment of such system is not common in the literature. Therefore, it is not surprising that the implications of this general concept within the still more general framework of plate tectonics have not been systematically examined until now. In this paper, we start by reviewing previous formulations of volcanic systems. By using key aspects of these models, we present a revised version of the formal definition of a volcanic system and use the updated model to explain general features of volcanism at a global scale. In particular, we concentrate in some of the troublesome aspects of volcanism mentioned above (the existence of polygenetic and monogenetic volcanism, the point-like distribution of volcanism, and the occurrence of volcanism at odd locations) while exploring some of the relationships that exist between the concept of "volcanic system" and the more general theory of "plate tectonics".

At the onset, it is important to point out that we have focused our attention in qualitative aspects as a necessary first step towards a better understanding of a very complex system. While we offer a synthesis of current understanding of physical processes relevant in volcanic systems, we also hope that this review can be used as a blueprint to focus future debate on some of the key questions raised.

2. Volcanic systems: previous models

Walker (1993b, 2000) formally defined a volcanic system to include: (1) the melting anomaly at the magma source; (2) the conduits through which magma rises to the surface; (3) the magma chambers (if any is formed); (4) the zones of geothermal influence; and (5) the volcanic edifice (Fig. 1). Using this definition of a volcanic system, Walker (1993b) grouped basaltic volcanoes into five distinctive system-types based on variations on the time-averaged magma supply rate and on the frequency of incoming magma batches (Fig. 2). In his classification scheme, lava shield volcanoes and flood basalt fields form at the highest time-averaged output rate (~ 1 km^3 of lava produced each year). The distinction between these two systems resides in the frequency of incoming magma batches,

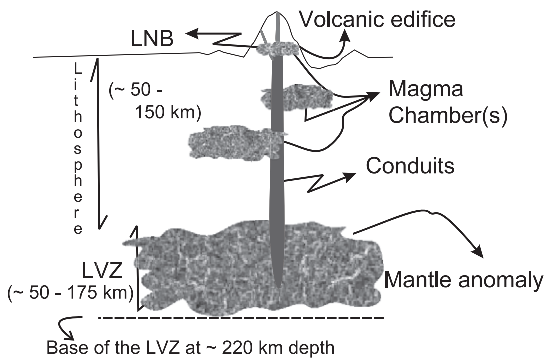

Fig. 1. Diagram showing the main components of any volcanic system (right column) and their relation with some important levels mentioned in the text (left column). The figure combines information provided by Walker (1993b), Sato and Ryan (1994), Bostock (1999) and Davies (1999). LNB = level of neutral buoyancy, LVZ = low velocity zone.

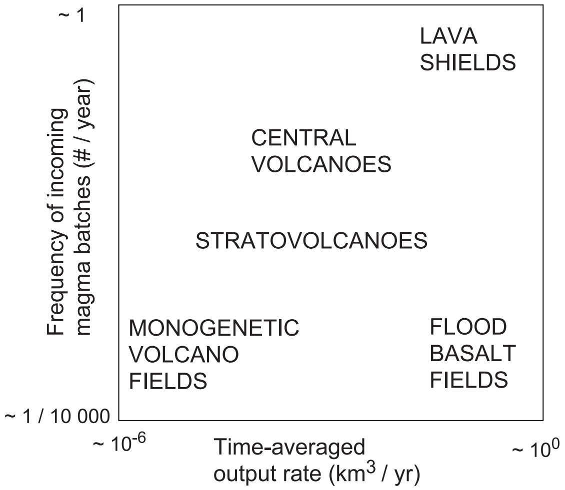

Fig. 2. Proposed relationship between the time-averaged output rate and the frequency of magma batches as the main factors controlling volcano morphology (after Walker, 1993b).

being much higher on lava shields ($\sim$ 1 episode/year) than on flood basalt fields ($\sim$ 1 episode every 10 000 years). Central volcanoes and stratovolcanoes form at intermediate time-averaged output rates ($\sim 10^{-3}$ km^3 of magma produced each year), but are distinguished from each other in terms of the frequency of magma batches (1 episode every 1000 years and 1 episode every 100–500 years, respectively). The fifth type of volcanic system in this scheme includes monogenetic volcanic fields, in which both the frequency of magma batches and the time-averaged magma supply rate are low ($\sim 10^{-6}$ km^3 of magma produced each year and 1 episode every few thousand years, respectively). Although the classification proposed by Walker (1993b) accounts for the most obvious differences in the morphology of basaltic volcanoes in the Earth, it does not explain the characteristics of the magma source or the processes responsible for the variation of the two main parameters considered (magma supply rate and frequency of incoming magma batches). Additionally, this model does not attempt to link the processes responsible for volcanic activity with a tectonic setting, or any other tectonic variable.

An independent formulation of the concept of a volcanic system was made by Gudmundsson (Gudmundsson, 1995). His model is based on his long-term observations of Icelandic volcanic activity (e.g., Gudmundsson, 1986, 1987, 1988, 1990a,b). According to Gudmundsson's model, the surface expression of a volcanic system in divergent plate boundaries consists of tension fractures, normal faults and volcanic fissures, regardless of the volcano-type formed (fissure swarm, central volcano or caldera; Fig. 3). The relative independence between the morphology of the edifice and the general surface expression of the system is explained on this model as the result of the implied progressive evolution of the system. Accordingly, the birth of a volcanic system is marked by the formation of a dome-shaped magma reservoir at the crust-upper mantle boundary. During early activity, the system lacks a shallow crustal magma chamber, so that all intrusions (dykes) and extrusions (eruptive

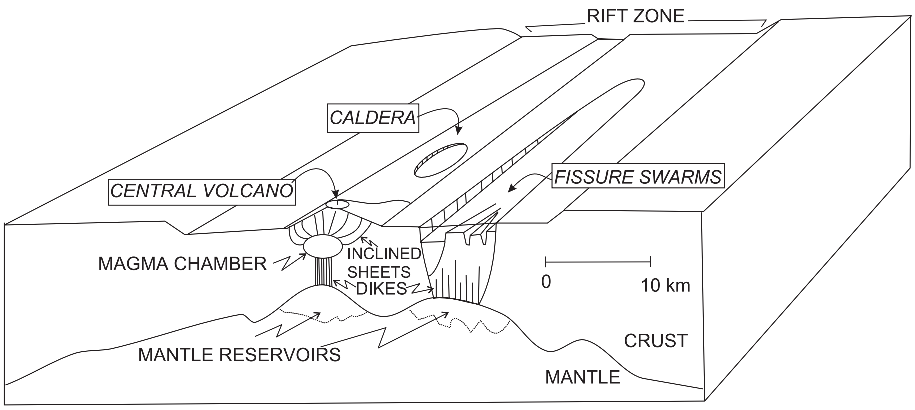

Fig. 3. Scheme showing the main components of volcanic systems in rift zones and the three morphologic types of volcanoes at the surface according to Gudmundsson (1995).

episodes) are fed from the deep reservoir, thus developing a regional dyke swarm. The rapid emplacement of many dykes may, in principle, shift the relative importance of horizontal and vertical stresses with time (Gudmundsson, 1986; Walker, 1993a). If the horizontal compressive stresses become higher than the vertical stress as the result of either tectonic or volcanic events, the emplacement of magma to form a sill will be favored and a crustal magma chamber might be formed eventually by the system. Upon formation of the shallow magma chamber, the conditions would be set for the system to evolve into a central volcano, and further evolution of the system would lead to the development of a caldera.

In analogy with Walker's model, in Gudmundsson's version of a volcanic system both the deep-seated magma reservoir and most of the plumbing of the system are common features of all the identified morphologically distinct volcano-types. Unlike Walker's model, Gudmundsson (1995) incorporates some variables of tectonic origin although the explicit relation between a volcanic system and plate tectonics still remains limited to rift scenarios. The combined effect of the relative timing of magma incursions into the upper crust and the influence of the regional stress-field, however, can be used to further explore the relationship existing between plate tectonics and volcanism at a global scale as shown in this paper.

Another aspect of particular importance is the recent shift concerning the concept of magma chambers. Various lines of seismic and petrologic evidence have led some authors to replace the view of isolated pools of magma beneath volcanoes and ocean ridges with a concept that portrays magma chambers as vertically distended mush columns (Sinton and Detrick, 1992; Marsh, 2000). This change in how we envisage the physical aspect of a magma chamber also imposes some constraints to the processes that are significant in a volcanic context. Consequently, the characteristics of the deep reservoir envisaged by Gudmundsson (1995) have to be modified to incorporate recent advancements in the understanding of magmatic systems. The consequences of incorporating key aspects of the magmatic system into the context of a volcanic system as envisaged by Walker (1993b) and Gudmundsson (1995) are far reaching, as will be shown in the remainder.

Although each of the two available versions of a volcanic system were inspired by volcanoes with specific characteristics (of basaltic composition in Walker's case and restricted to rift scenarios in Gudmundsson's case), both models offer a very convenient framework that can be used as a starting point of a more general version in which the volcanic system is inserted more clearly within the plate tectonics paradigm. In Section 3, we combine key aspects of both models (i.e., Walker's and Gudmundsson's) to develop an updated version of a volcanic system that is independent of chemical composition and that evidences the effects of the tectonically controlled variables.

3. Volcanic systems: a revised model

3.1. Mantle anomaly

3.1.1. Time and rheological constrains

One obvious condition that must be satisfied to have volcanic activity on a planetary surface is the presence of a liquid phase stored under a solid surface. On Earth, partial melting of the mantle is the main source (directly or indirectly) of the magmas erupted on the surface, and the conditions favoring mantle melting are directly related to plate tectonics. The time-scales of volcanic and tectonic processes, however, are very different from each other. Such difference in the time-scale involved is commonly overlooked, but it is a key factor in the understanding of volcanic activity at a global scale, and in clarifying the relationship that exists between plate tectonics and volcanism.

From the three ways that exist to melt a rock, the two dominant on Earth are a decrease in pressure and a change in the solidus temperature concomitant with a change in the chemical composition of the rock (Asimow, 2000). The former, and probably most important of these mechanisms is associated with the slow ascent of mantle rocks. For this reason, we concentrate on this section in the consequences of mantle melting as the result of vertical movements of mantle rocks. It is noted, however, that rock melting of shallower material (shallower mantle or crust) either by a change in chemical composition or by an increase in local temperature can become important at a late stage in the development of a volcanic system, although it will become apparent that the exact mechanism of melt production only plays a secondary role subordinated to the evolution marked by other parameters.

Ascent of mantle can take different forms in different tectonic settings. In divergent-plate boundaries vertical movements of mantle are likely to be related to large-scale convective motions (Davies, 1998), whereas in convergent-plate settings rising blobs of mantle material can be invoked as the result of a drop in the viscosity induced by the addition of small amounts of water accompanying subduction (Tatsumi et al., 1983; Asimow, 2000). For intra-plate settings rising diapirs (or mantle plumes) have been commonly invoked (Morgan, 1971), although the plume hypothesis involving deep-mantle processes has been recently questioned (Anderson, 2000; Foulger, 2002). In any case, despite the differences in the exact mechanism controlling mantle ascent in different tectonic settings, due to the large viscosities involved ($10^{17}-10^{21}$ Pa s (Lambeck and Johnston, 1998)) the rate of ascent of mantle rocks will be of the order of cm/year (Rubin, 1993), and can be considered as representative of the characteristic time-scales of mantle movement in all tectonic scenarios.

Not only the ascent of mantle rocks, but all other tectonic processes can be characterized by rates that are limited by the large viscosities involved. Melting production rates, however, are likely to occur at a much faster rate than that characteristic of tectonic processes. For instance, the rate of crystal dissolution in an existing magma has been estimated to be significant on the order of hours to months (Edwards and Russell, 1998), and the achievement of textural equilibrium between a crystalline solid and a liquid phase has been estimated to take place in just a few years (Cooper and Kohlstedt, 1986). Although these processes are not completely equivalent to the process of melt production from a completely solid phase, based on the available evidence it is reasonable to assume that the time required to melt mantle rocks will not be much greater than the characteristic times observed for crystal dissolution, or the achievement of textural equilibrium, once the P–T conditions for melting have been reached.

The characteristic rate of tectonic processes is also extremely slow when compared with the rate at which volcanic processes take place. In the absence of a different mechanism allowing rapid vertical movement of magma, volcanic manifestations on the surface would take place only over time periods in the order of 10^5-10^6 years (the time required for a parcel of mantle material to ascend a few kilometers). This contrasts with the time scale of a single volcanic eruption (days to tens of years), or even the lifetime of a large volcanic center ($\lesssim 10^6$ years). The difference in both time scales is so large indeed, that for most practical purposes in volcanology the mantle can be considered as essentially stationary (Rabinowicz et al., 2001). Therefore, there seems to be a contradiction in the fact that a basically stationary mantle, where the $P-T$ conditions change only very slowly, is directly responsible for processes that occur at much faster rates like melting or volcanic activity on the surface.

The process linking both extremes seems to be magma ascent through rock-fracturing because this process allows magma ascent rates that are more akin to the time-scale characteristic of volcanic activity (Rubin, 1993). Consequently, the birth of a volcanic system can be defined as the moment when hydraulic fracturing allows a significant amount of magma to be transported out of the source region. The time during which plate tectonic processes favored the initiation of melt production, but in which the conditions necessary for fracturing of the rock above the zone of melting had not been achieved, can therefore be considered as a gestation period of a volcanic system.

Note that these definitions of a gestation period and the birth of a volcanic system differ from that given by Gudmundsson (1995) in that the conditions required to initiate melting are not considered an integral part of the volcanic system, but rather are considered part of plate tectonics. It is also important to note that, based on these definitions, the actual conditions required to produce mantle melting are not limited to ascent of large mantle diapirs. Alternative mechanisms of mantle melting such as the lateral displacement of mantle regions that have a variable amount of fusable material, the presence of lateral temperature gradients within the mantle, or other effects of a self-organized plate-tectonic system (e.g., Anderson, 2000, 2001; Foulger, 2002) are entirely compatible with the definition of a volcanic system.

3.1.2. Deep magma chambers and the conditions for the birth of the system

It is now necessary to consider in more detail the conditions required for hydraulic fracturing of the rock above the melting zone. The pressure required to initiate hydrofracturing is determined by the density contrast between melt and solid rock, and by the tensile strength of the overlying rock. These two factors determine a threshold height of the column of melt required for initiation of hydrofracturing, and also impose some constraints in the position of that column and on the volume of melt that can be erupted. For instance, Stolper et al. (1981) have documented an apparent decrease of the density contrast with increasing depth so that it can become extremely small (< 0.1 g/cm^3) at depths of over 110 km, and even it might become negative (i.e., the melt becomes more dense than the surrounding solid) at depths ~ 250 km. These estimates impose a lower limit for the depth of melt that can ascend through hydrofracturing. On the other extreme, Nicolas (1990) has shown that in oceanic ridge settings the critical height of melt would range from 10 km (for a density difference of 0.5 g/cm^3 at ~ 70–50 km depth) to infinite (for a depth equal to the Moho, where the density difference is assumed to be zero) assuming a tensile strength of 50 MPa. Although strictly related to rift scenarios, considering that the uncertainties on the variation of melt density as a function of depth are large and that little is known about the exact tensile strength of the rocks involved in a particular location, it seems reasonable to accept the 10 km obtained by Nicolas (1990) as an estimate of the order of magnitude of the critical height required to initiate hydrofracturing in all tectonic settings. The critical height can then be used to assess the volume of melt required to initiate hydraulic fracturing. If all the melt was to concentrate in a single spherical pool before the initiation of hydrofacturing, the critical height would imply the presence of chambers with a volume at least 523 km^3 at 70-km depth. This volume is comparable with that estimated to have erupted in single fissure eruptions in some flood basalt provinces or in large explosive events (volcanic explosive index, or VEI ≥ 7), but it is much larger than the volume of melt erupted in more typical volcanic events (VEI ≤ 4; magma volumes ~ 0.3 km^3) (Hooper, 1997; Pyle, 2000). At the other extreme, the volume of magma needed before initiation of fracturing can be minimized if the "pool" of melt is assumed to have an elongated shape with a vertical long axis. A realistic lower limit can be found assuming that all the melt is contained in a vertical cylinder with length equal to the critical height (10 km) and 10 cm in diameter; the lower limit for the diameter of the cylinder corresponding to the typical dimension of veinlets observed in peridotite massifs (Nicolas, 1990). Such "magma chamber" geometry implies a melt volume $\sim 8 \times 10^{-8}$ km^3 that would generate a lava flow with a volume comparable to the carbonatite flows erupted at Oldoinyo Lengai volcano and that correspond to the smallest end of volcanic eruptions (Pyle, 2000).

There are some aspects of the melting process itself, however, that do not support the existence of chambers in which all the volume available is occupied by melted rock. In particular, it is well established that upon melting, connectivity of the melt phase will take place only through edges in which the interfacial energies between solid and melt satisfy precise stability criteria (Toramaru and Fujii, 1986). If melt interconnectivity allows even very small amounts of melt to rapidly move in a vertical direction away from its source region, the limited amount of melt ponding is likely to preclude the formation of the large hydraulic head required to fracture the overlying rock (on the order of 50 MPa as mentioned above). Even if we consider that all the ascent of melt takes place through percolation processes, we have to consider that as this melt raises it will find conditions that are unfavorable for its existence (Fig. 4). Due to the small melt volumes involved in this percolation process, the liquid rock will solidify very fast so that it is unlikely that a significant portion of melt can leave the zone of melting by this mechanism, although it can result in the slow ascent of the mantle anomaly in time scales of the order of 10^4–10^6 years (Stolper et al., 1981). These time periods are more akin to the gestation period of a volcanic system as defined above. Therefore, the melt produced within the mantle would remain effectively trapped at the source region unable to reach the surface in timescales comparable to those of volcanic activity,

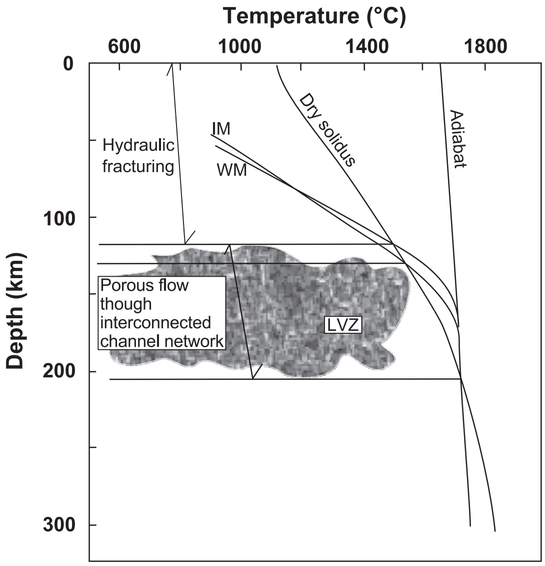

Fig. 4. Temperature profiles of two regions within the USA (IM = intermountain, WM = western margin) as inferred from seismic data, solidus and adiabat lines taken from Sato and Ryan (1994). The horizontal lines mark the vertical limits of the zone of melt production/storage or LVZ. Although the amount of melt inferred to exist at each region is not the same in both cases, above the LVZ the thermal structure of the mantle does not favor the presence of any melt (note that the upper end of the LVZ is different for each region). Any melt intruding the layer above the LVZ will experience cooling and eventual solidification, unless hydraulic fracturing allows its rapid ascent.

unless the interconnectivity of mineral grains within the source region allows the eventual formation of a vertical conduit entirely within the region of melting. The vertical dimensions of this conduit will have to be comparable to the critical height, although density variations with depth within the source region, and local variations on the strength of the overlying rock, are likely to be important factors controlling the exact value of the critical height required to initiate hydraulic fracturing.

The melting process of mantle rocks also imposes some constraints on the shape of the pools of magma that can be formed at depth. Observations made on mantle peridotites, where the melting process has been quenched at various stages (Nicolas, 1986, 1990), show that local melt segregation is preferentially parallel to the flow plane experienced by the slowly deforming rocks. If we consider that the deformation near the top of an ascending blob is horizontal both on the fluid around and within the blob (Griffiths, 1986; Cruden, 1990), and also take into consideration the general pattern of flow associated with mantle convection (Davies, 1999), it is reasonable to conclude that the mineral fabric of the mantle rocks will tend to be horizontal rather than vertical near the top of the asthenosphere and on the lower parts of the lithosphere. Additionally, if we consider that capillary forces and the slow deformation of the solid matrix can be up to 4 orders of magnitude more important than melt buoyancy in regions with only small amounts of melt present (Riley and Kohlstedt, 1991), the establishment of lateral rather than vertical connections seems more likely, despite the lower density of melt relative to the still solid rock. Consequently, magma collection within a vertically elongated pool within the region of melting seems rather implausible unless the amount of total melt is relatively large. Note that by limiting the development of melt interconnections in a vertical direction, the time required to achieve the conditions for hydrofracturing is extended until a significant change in the $P-T$ conditions of a particular region of mantle have changed significantly so that more melt is produced.

On the other hand, it is important to note that the melting process and the porous flow of melt do not preclude the slow accumulation of large pools of magma, although they impose some conditions preventing these magma chambers to form an integral part of a volcanic system. As the flow evolves at the interior of the region of melting, the melt will tend to be concentrated in larger pools as a direct consequence of the flow process itself (Rabinowicz et al., 2001), or as a consequence of chemical reactions (Daines and Kohlstedt, 1994). Thus, deep magma chambers can be formed on the top of the melting zone as the result of the evolution of porous flow. However, as long as there is not enough hydraulic head to fracture the overlying rock, the melt in these chambers will only ascend rather slowly together with the mantle anomaly, and will not feed any volcanic eruption, nor will contribute for the development of a volcanic system. Even if we consider that the melt trapped within the source region is likely to result in a gravitational instability that eventually might lead to the formation of a mantle diapir, the slow nature of the porous flow leading to the melt collection on these pools, and the slow ascent rates of mantle diapirs, still will result in time scales $\sim 10^6$ years (as estimated in Section 2) before the onset of hydraulic fracture of the overlying rock. Therefore, the chambers formed as the result of the evolution of deep porous flow of melt will belong to the incubation period of a volcanic system as defined above and, generally, will not form part of any volcanic system.

In summary, currently available evidence suggests that the presence of large cavities completely filled with melt at upper-mantle or lower-crust depths below a volcanic system is unlikely. Instead, the geometrical constrains imposed by the melting process, through the mineral channel network, suggest that melt remains trapped within the network extending for large distances horizontally. Evidence provided by the variation of seismic velocity as a function of both depth and lateral position around the planet is consistent with this hypothesis.

3.1.3. The roots of a volcanic system: seismic evidence

Through the development of modern computational capabilities, and with the establishment of a worldwide network of seismographs, great advances have been made in recent years in determining the mantle *P* and *S* velocity distribution on a global scale (Nolet et al., 1994). One important result of this type of seismic studies was to establish the presence of a low shear velocity zone (LVZ) embracing the upper 200–300 km of the mantle (Davies, 1999). Although the interpretation is still somewhat controversial, it is generally accepted that the LVZ might reflect the presence of small amounts of melt ($\sim$ 1–2%, in some cases up to 10%) (Matsushima, 1989; Sato and Ryan, 1994). Most importantly, from the point of view of a volcanic system, the LVZ is not uniformly distributed around the planet being more clearly defined in tectonically active regions, like present day plate margins (Grand, 1994). The global distribution of LVZs roughly coincides with the distribution of active volcanism on a global scale and is therefore consistent with the predictions of plate tectonics. Thus, we infer that the deepest part of any volcanic system extends laterally at depth farther than the area covered by volcanic vents on the surface, and is defined by the low velocity zone observed in seismic surveys of the mantle. These regions are characterized by the presence of melt within an interconnected network of narrow channels, and only in rare circumstances form large accumulations of melt. The depth and thickness of these zones is variable, and may be associated with a change from a more ordered crystalline array to a more isotropic medium below, as suggested by the Lehman discontinuity (Bostock, 1999), or by other tectonic variables including the geometry of mantle deformation in specific tectonic settings (e.g., the observed variable depth of the LVZ as a function of distance from the trench in Japan (Hasegawa and Zhao, 1994)). In any case, it is remarked that these deep "magma chambers" will become part of a volcanic system only when the conditions for hydraulic fracturing of the overlying rock are achieved.

3.2. Zone of lateral influence of a volcanic system

Factors such as mineralogy of the mantle source, the magma production rate, the regional stress field at different depths and the lithological contrasts found on the upper level rocks will define where hydraulic fracturing above the mantle instability takes place and the critical height of melt that is required to initiate the rupturing process. Considering storage of melt in a network of narrow channels, as discussed above, the critical height is likely to be achieved more easily by a very narrow channel ($\sim$ centimeters of diameter, close to the veinlet dimensions discussed in Section 2) that forms randomly within the region of melting as interconnectivity extends vertically. Just after the hydraulic fracturing of the overlying rock starts, lateral flow of the melt filling the (mainly horizontal) interconnected network surrounding the newly opened vertical path is initiated. Therefore, the horizontal porous flow will feed the just formed vertical conduit that begins to ascend through rock fracturing (Nicolas, 1986, 1990). By considering a spherical volume of rock with a diameter equal to the critical height, it is possible to explain the existence of $\sim$ 6–160 km^3 of melt ready to be tapped, depending on the percentage of rock that is melted (1–30% of the volume of the sphere, respectively; the upper limit being selected as a reasonable upper limit derived from basaltic plateau provinces (Kent et al., 1997)). These volumes are comparable to magma volumes erupted by even the larger typical volcanic eruptions on Earth (VEI up to 7), and therefore suggest that a typical scale for the horizontal separation of two adjacent volcanic systems is on the order of 10 km. It is noted, however, that such estimate does not preclude the possibility of magma traveling from still larger horizontal distances at depth, and therefore of feeding much larger volcanic eruptions.

The physical conditions for the initiation and sustained flow within fractures that are rooted in a porous zone have been considered in some detail by Sleep (1988). He showed that the depth reached by the upper end of the opening dike depends on the amount of melt that can be tapped without reducing the pressure at its lower end, and on the orientation of the stress as it rises through the lithosphere. As a result of this mechanism of magma extraction, the amount of magma that can be tapped after initiation of any fracturing event will depend strongly on the extent of lateral interconnectivity of melt, therefore defining the limits of the zone of lateral influence around the just formed hydraulic fracture. When this lateral flow is not enough to sustain the pressure at the lower end of the vertical fracture, the eruption will halt, leaving some melt trapped within the original network (Fig. 5). Therefore, the limitations encountered upon tapping determine the next stage in the evolution of the volcanic system, as explained next.

3.3. Time evolution of a volcanic system

Although in Section 2, it was shown that the range of melt available for tapping from the mantle melting process is broad enough to explain the typical range of volumes involved in single eruptions, it is emphasized that not all of the available melt must be tapped in a single event. Actually, it is likely that the surrounding matrix will be modified as the lateral porous flow evolves, following the opening of the vertical fracturing. Such rearrangement in the geometry of the matrix might result in the opening of other vertical fractures in rapid succession, or alternatively it may widen the already existing vertical channel, depending on the details of the deformation of the matrix itself, but also on the strength of the overlying rock. In any case, the vertical ascent of magma away from the melting source will be also influenced by the orientation of regional stress at shallower levels. The combination of

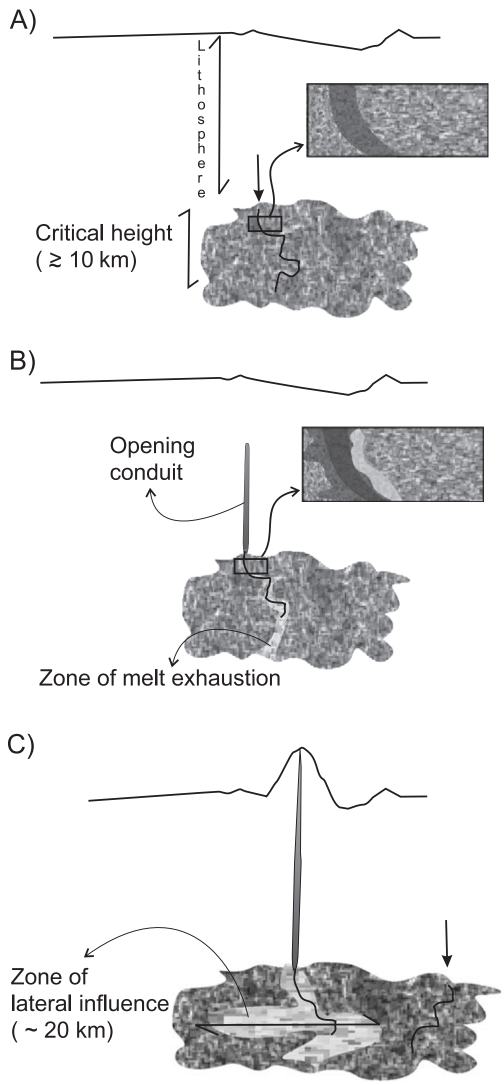

Fig. 5. Diagrams showing the evolution of the melt network upon the opening of the hydraulic fracturing. (A) shows the time immediately before a narrow channel reaches the critical height to initiate fracturing. The melt tapped from the LVZ is collected by the opening conduit shown in (B) while producing a zone of melt exhaustion on its vicinity, shown by lighter shades of gray. (C) shows a system in which the conduit was large enough to reach the surface and construct a volcanic edifice on the surface. The eruption stops when the melt network cannot supply the conduit with enough melt to sustain the pressure on the dike, even if there is enough melt available in the LVZ. The vertical arrows in (A) and (C) indicate the origin of the next fracturing event. The insets in (A) and (B) show in more detail the effects of a well-interconnected network of melt (left of the conduit) and of a limited degree of interconnectivity (right of the conduit).

all of these factors ultimately will control the evolution of the system. Due to the large number of variables involved, a qualitative treatment of these processes remains a monumental task. However, much insight can be gained by focusing on the most basic aspects of the problem. Actually, by using two constraints imposed by the presence of tabular intrusions in the geologic record, it is possible to construct a qualitative model that encompasses many aspects of volcanism as observed at various localities around the globe. One of these constraints is that buoyancy is not enough to drive magma ascent at all times. If this were the case there will be no reason for any magma to be trapped in the vertical conduit once opened, and consequently dikes would not exist (or would be less common than they are) in the geologic record. Although certainly the hydraulic head of magma is the factor responsible for the initiation of magma fracturing, despite the difference in densities, ascent of magma through a fissure will not continue unless there is enough pressure (and therefore continuity in magma flow) at the entrance (lower end of the dike). The second constraint imposed by tabular intrusions is that these bodies develop in such a way that the plane of the intrusion is normal to the least principal stress (e.g., Nakamura, 1977). The occurrence of large tabular intrusions with a horizontal attitude (Wilson and Wheeler, 2002) is therefore evidence that the least regional stress can sometimes become vertical. Therefore, in addition to buoyancy it is necessary to consider simultaneously the magma pressure at the lower end of the conduit and the orientation of the stress field along the path followed by the opening fracture.

By using these constraints within the framework of a volcanic system, we now explore three alternative end-member scenarios depending on the combination of the amount of lateral interconnectivity within the region of melting and the orientation of the regional stress in the overlying rock (Figs. 6–8). Although in these scenarios the conditions are fixed a priori, the discussed behavior of the system is the logical consequence of those conditions. Consequently, the better the resemblance of natural occurrences of volcanism with any of the inferred behaviors the more likely the existence of a particular set of conditions at a given time. It is clearly possible to find specific examples of volcanism that combine some aspects of two scenarios. The existence of such "hybrid" systems does not imply that the model does not work, but rather is a

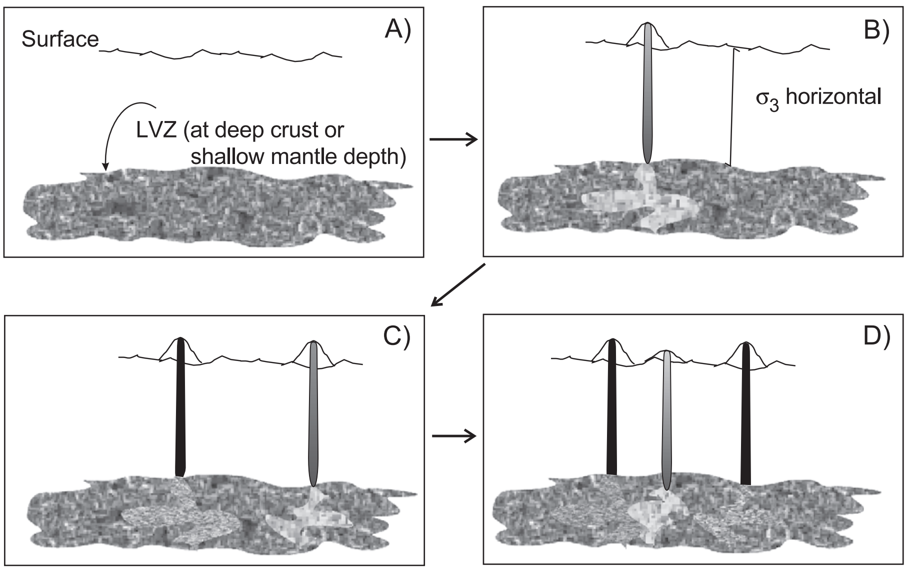

Fig. 6. Schematic evolution of a volcanic system from just before the end of the gestation period: scenario 1. Gray-scale conduits are active or very recent, whereas the black conduits mark older events in which magma has solidified. The significance of the color differences observed on the LVZ through time (A–D) are explained in Fig. 5. See text for details.

natural consequence of the complex nature of volcanism. Nevertheless, it is remarked that in practice it becomes easier to understand the complexities offered by many specific examples by making reference to a simple model with three end-members as that developed here.

In the first scenario, there is enough interconnected magma at the source region to keep pressure at the entrance of the conduit high enough to keep pumping magma to an ever increasing level and the minimum principal stress is always horizontal. Upon the initiation of hydraulic fracturing, these conditions will result on immediate surface manifestations. Therefore, these conditions apply to the formation of a volcano that issues magma directly from the depths of its source region (Fig. 6). The eruption continues until lateral flow of magma in the source region fails to sustain the required pressure at the entrance of the dike. Consequently, tapping of magma will temporarily exhaust the melt available in the source region immediately surrounding the dike, but melt will continue to exist regionally (Fig. 6B). The availability of interconnected melt at a regional scale implies that the conditions for hydraulic fracturing can take place at a slightly different place in a relatively short period of time (from months to a few hundreds of years) resulting in the formation of an independent volcano at the surface (Fig. 6C). Consequently, these conditions also explain the existence of monogenetic volcanic fields that are not associated to a larger structure. The continuing effects of regional melting, and the lateral flow of magma, will "recharge" the areas that were depleted by previous eruptions so that another conduit can be opened eventually very close to an existing conduit (Fig. 6D). In an extreme situation, if the recharge velocity is very high, the same conduit can be used at least partially for a second time or, more commonly, the new conduit will be parallel to the previous one. The whole process will continue to operate regionally as long as there is enough melting available at depth, resulting in the compositional similarity (but not uniformity) observed in many monogenetic volcanic fields. If the mantle anomaly survives long enough to ascend significantly, a compositional trend of the erupted products, and probably of the vent locations will be defined, as observed for example in the Chyulu volcanic field, Kenya (Novak et al., 1997).

In the second scenario, magma pressure is also large but the regional stress orientation changes at some depth so that the least principal stress becomes vertical (Fig. 7). These conditions imply that upon the

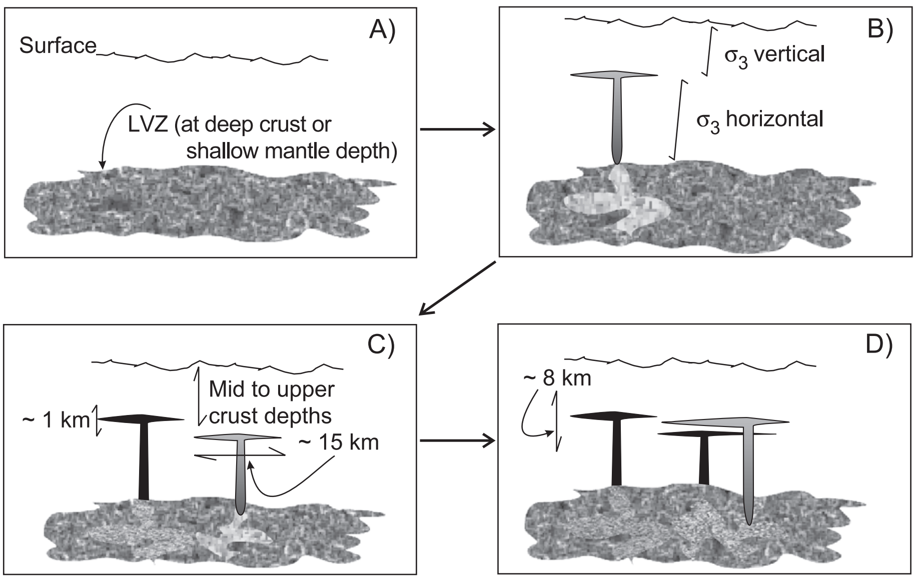

Fig. 7. Schematic evolution of a volcanic system from just before the end of the gestation period: scenario 2. Symbols as in Fig. 6. The vertical scales shown are based on seismic imaging of sill complexes by Wilson and Wheeler (2002).

initiation of hydraulic fracturing magma will never reach the surface, but rather that a sill will be emplaced at depth. The sequence described for the first scenario will be repeated, except that in this case there will be no surface eruptions. Instead, the melt solidifies as sills without reaching the surface. The depth where these sills form may not coincide with the depth at which the density of the magma equals the density of the surrounding rock, or level of neutral buoyancy (Ryan, 1987). Instead, the depth of sill emplacement can be primarily controlled by the level at which the stress of the country rock became horizontal. Other factors controlling the depth of intrusion include the depth reached by the vertical conduit before the occurrence of a significant pressure drop at their lower end and the brittle–ductile boundary on the crust, or the mantle–crust boundary (Wilson and Wheeler, 2002). In any case, despite the relatively large amounts of magma that are suddenly accumulated at some depth within the sill, the effects of cooling at the boundaries of the intrusion are likely to increase the difficulties experienced by the magma to rise still further. Such boundary effects will therefore promote the solidification of the complete magma pool as a single intrusion with a uniform appearance like that described by Marsh (2000). Interconnected dike and sill complexes also can be explained by this scenario.

In the third scenario, the amount of interconnected melt is not large enough to maintain the pressure at the entrance of the opening conduit regardless of the orientation of the regional stress. These conditions imply that the dike initiated at the onset of hydraulic fracturing will stop propagating upwards while still at depth, leaving a sizable amount of magma trapped in the conduit, but without forming a horizontal intrusion (Fig. 8). The limitation for the dike propagation is the poor interconnectivity of melt at the source rather than an intrinsic lack of a large volume of melt, and as in the first scenario, the conditions are adequate to produce the rapid succession of independent magma injections. In this case, however, none of the injections will reach the surface. Also, the depletion zone around the vertical conduit will be smaller than in the first scenario and, consequently, the combined effects of many small intrusions occurring in rapid succession may promote the convergence of fluid-filled cracks (Takada, 1994a,b). Additionally, the combined effect of the rapid succession of magma intrusions on the temperature of the surrounding rock at upper levels might promote the incubation of a region of partial melting at shallower depths (Hardee, 1982; Shaw, 1980). Consequently, the initial conditions for this case can adequately explain the formation of an upper level reservoir of magma. As already explained, the depth of formation of this shallow reservoir is not necessarily the level of neutral buoyancy, which explains the occurrence of chambers that seemed to be controlled rather by a rheological contrast in the country rock (Hooft and Detrick, 1993). The level of neutral buoyancy, however, will influence the evolution of the system over time. As the proximity to the level of neutral buoyancy of the original shallow reservoir reduces the initial hydrostatic head, the length of the dikes originated on the shallow level reservoir will be smaller on the average than those originating at the depth of magma production. Consequently, the onset of magma accumulation at still shallower depths will be facilitated, favoring with time the accumulation of magma at the level of neutral buoyancy. This process will facilitate the formation of a plexus of small magma reservoirs at various depths, or of a distended mush column (Marsh, 2000). Other factor that can influence in the development of such plexus of small magma chambers is the inhomogeneity of the composition of the shallower rocks (Edwards and Russell, 1998). Most of these inhomogeneities will be the result of the larger-scale tectonic processes (e.g., Bostock, 1999; Foulger, 2002), and therefore can result in melting zones that have characteristics apparently unrelated with a given tectonic setting. Also, it is noted that as the surface is gradually approached with this process, the interaction of various dikes can affect the mechanical properties of the rock in such a way that a central conduit, and consequently a polygenetic volcano might be established. The occurrence of an eruptive event through independent conduits (and hence of a monogenetic vent), however, is not precluded. Over time, the upper level reservoir(s) will serve to modulate the output of magma to the surface. On analogy with the processes described for the deeper level, the frequency of events of hydraulic fracturing from the upper level reservoir will be determined by the interconnectivity of melt at the shallow reservoir, the mechanical properties of the

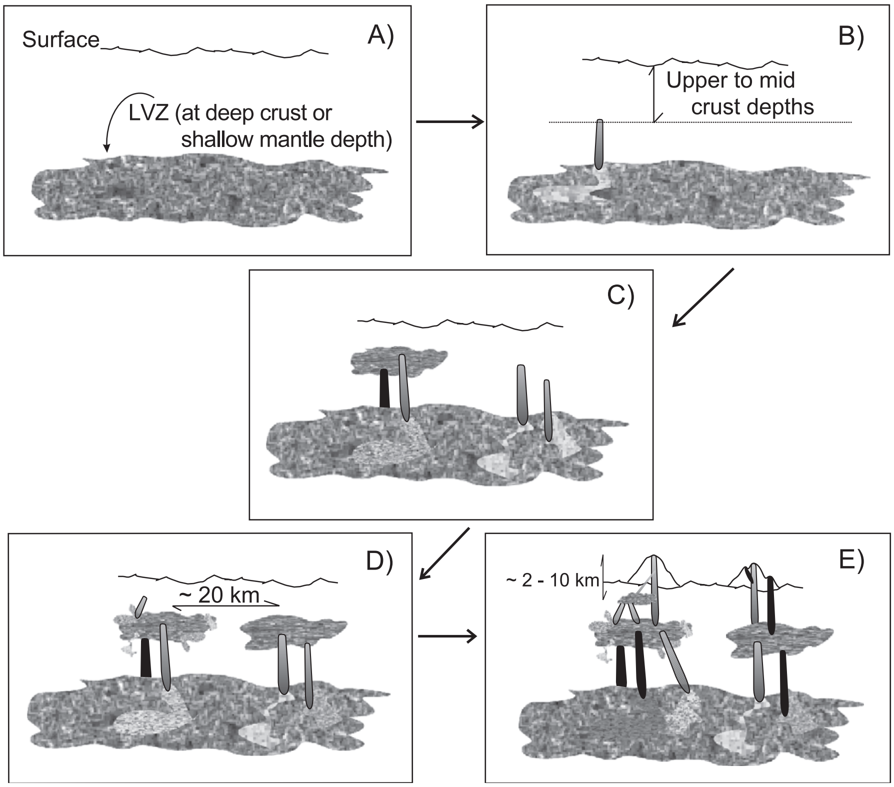

Fig. 8. Schematic evolution of a volcanic system from just before the end of the gestation period: scenario 3. Symbols as in Fig. 6. The influence of the shallow magma chamber on the local stress can result in hydraulic fracture events that depart notably from the more typical vertical direction. The horizontal scale in (C) is taken from the observed separation between volcanoes along the same tectonic margin (de Bremond d'Ars et al., 1995), and the vertical scale in (D) reflects the position of the level of neutral buoyancy (Ryan, 1987). Note the presence of monogenetic, parasitic cones in the volcanic edifices in (D).

rock surrounding the upper level reservoir, and by the rate of accumulation of pressure in it. Depending on the vertical position of the chamber and on the original volatile content of the magma, gas exsolution may be an additional factor important in producing the excess pressure required to fracture the surrounding rock and in controlling the development of the injection/eruptive event (Parfitt et al., 1993).

In summary, by focusing attention in the amount of interconnectivity present at the region of melting and to the orientation of the regional stress, it is possible to explain most of the observed features of volcanism around the world. As noted, not in every instance the initiation of hydraulic fracturing leads to a volcanic eruption. Indeed, the two latter scenarios just described result in a system that remains hidden at depth for a longer period of time after its birth. It is, therefore, by this process that a volcanic system may lead to the formation of volcanic structures at the surface other than monogenetic volcanoes. However, as the model implies, formation of monogenetic volcanoes is not exclusive of the conditions defined on the first scenario, but rather a consequence of the combination of several factors defining the critical height of the magma column that is required to initiate hydraulic fracturing at various depths. It is in the third scenario that melting as a consequence of factors other than ascent of rocks becomes important. The amount of melt produced at these shallow levels will therefore depend, to some extent, on the frequency and magnitude of the events directly rooted on the mantle source. Various degrees of crustal contamination can therefore be explained in this form.

4. Discussion

Instrumental in the formulation of the updated model developed in the previous sections, is the continuity that must exist between successive stages of a volcanic system. Dobran (2001) also has pointed out this continuity while discussing at length the mathematical aspects necessary to model processes occurring at specific stages. The multivariate nature of the problem, however, introduces large levels of complexity, making the establishment (and solution) of a quantitative model that encompasses all of the stages of the system difficult. Despite such complexity and unlike previous formulations (Walker, 1993b, 2000; Gudmundsson, 1995), our version of a volcanic system portrays a general picture of volcanism within a single, coherent model in which the role played by tectonic variables is easier to identify. Such identification is blurred when attention is focused only on isolated aspects of volcanism.

Despite its general character and qualitative nature, our model of a volcanic system allows us to explain many characteristics of volcanoes around the world that otherwise are hard to explain even qualitatively, and provides a general framework that allows us to forward simple explanations of aspects of volcanic activity that are not adequately explained by plate tectonics. We will focus the following discussion on three aspects of volcanism that have remained somewhat puzzling until now (volcano spacing along tectonically active regions, the occurrence of polygenetic and monogenetic volcanism and the occurrence of some volcanoes at odd locations around the world), and for which specific examples can be found very easily in at least two different tectonic settings. It is remarked, however, that the implications of the present model are not limited to these particular aspects of volcanism but extend to cover almost any volcanic feature on our planet.

4.1. Volcano spacing

According to plate tectonics the relative movement of the rigid plates that cover the Earth's surface define three specific scenarios where volcanic activity can take place. In two of these scenarios, partial melting of the mantle takes place (directly or indirectly) as the result of relative plate motions (Wylie, 1988). Despite the effect of local inhomogeneities, the relative motion of the plates predicts partial melting processes to occur along the whole length of convergent and divergent margins, and not to be focused at very specific localities leaving large portions of the same margin devoid of melt. Consequently, according to plate tectonics, we should expect to have volcanoes forming a continuous volcano along these margin types. What we actually observe is that although indeed nearly 90% of volcanic activity is concentrated on these zones (Perfit and Davidson, 2000), only on very special cases volcanic activity is continuous along plate margins. Rather, volcanic activity concentrates on very specific locations forming volcanic edifices or fields (two different types of systems) that are separated from neighboring centers of volcanic activity by distances ranging from 20 to over 200 km (Shimozuru and Kubo, 1983; de Bremond d'Ars et al., 1995). Most explanations for the point-like distribution of volcanism have invoked some form of mantle instability that results on the raising of discrete diapirs from a source that is continuous (Marsh and Carmichael, 1974; Kerr and Lister, 1988; Davies and Stevenson, 1992; de Bremond d'Ars et al., 1995). All of these models, however, fail in considering the time scales associated with volcanic activity on the surface and the ascent velocity of mantle diapirs at depth.

The extent and location of the LVZ around the globe coincides very nicely with the predictions of plate tectonics, forming an almost continuous belt underlying the tectonically active margins. This belt is the largest storage zone of melt that we can conceive, and lateral variations of the amount of seismic velocity attenuation can be interpreted as the result of variable amounts of melt, and of interconnectivity of such melt within that region. According to the model of a volcanic system developed above, the location of any individual volcano is determined by the place where fracturing first allows magma to move upwards more rapidly. The ensuing lateral movement of magma within the channel network at depth is controlled by the interconnectivity of melt in that region, and its extent defines the zone of lateral influence of that particular volcano. Depending on the local conditions of melt interconnectivity and lithospheric stress, the system can evolve differently along the same tectonic margin, therefore resulting in different styles of volcanism on the surface. In the case of an evolution leading to the formation of polygenetic volcanoes (scenario 3), the extent of the zone of lateral influence of a single volcanic edifice will coincide with that of a system. In this case, all of the magma produced within a large portion of the LVZ is funneled at depth to be eventually erupted as part of one single volcano. In some cases, the system can develop not one but two preferred conduits to lead most of the magma to the surface, resulting in a composite volcano, yet the zone of influence of the volcanic system remains more or less the same through time. Therefore, the observed separation of volcanic edifices along convergent margins can be interpreted as a measure of the extent of zones of lateral influence of volcanic systems. The observed separation suggests that lateral flow of magma within the zone of melting usually takes place for distances of less than 10 km, although in some cases it may be much larger.

In any case, the concept of volcanic systems offers us an explanation of the point-like distribution of volcanic centers that does not rely on the movement of mantle instabilities, and that therefore is compatible with the time scale of both volcanic and tectonic processes. Certainly, the ascent of mantle rocks during the gestation of a volcanic system might influence the exact location of volcanic centers at time scales of millions of years, but it is stressed that on shorter periods of time the most important factors controlling the distribution of volcanic centers at the surface are those described here.

4.2. Monogenetic vs. polygenetic volcanism

The general conditions required for the occurrence of monogenetic and polygenetic volcanism, as envisaged from the point of view of a volcanic system, have been outlined above (scenarios 1 and 3, respectively). Unlike previous models of polygenetic and monogenetic volcanism (Fedotov, 1981; Walker, 1993b), the explanation offered by volcanic systems does not rely on differences of magma generation rate to explain the differences on volcanic style. From a plate-tectonics point of view, rate of melting is controlled mainly by the rate of ascent of mantle rocks. As already explained, rates of ascent are slow in any tectonic setting. Consequently, based on plate tectonics, it is hard to explain why rates of melt are more rapid below a polygenetic volcano, whereas they are slower below a monogenetic volcanic field even when these features might coexist along the same tectonic setting (e.g., the Michoacán-Guanajuato volcanic field and the Popocatepetl and Colima stratovolcanoes along the Mexican Volcanic Belt, Alaniz-Alvarez et al., 1998).

From the point of view of a volcanic system, the most important difference between a polygenetic volcano and a monogenetic field with deep magma sources is not the rate of melting itself, but the degree of melt interconnectivity that prevails in each of the two volcanic regions. This is so because melt interconnectivity determines whether magma can reach the surface directly from its source region upon hydraulic fracturing of the overlying rock. Consequently, the occurrence of both polygenetic and monogenetic volcanism is compatible with variations on the local composition of the mantle that could be invoked to favor a more rapid melting process below a polygenetic volcano, but it also is compatible with scenarios in which there is a similar composition of mantle sources below two regions, each characterized by a different style of volcanic activity. Indeed, two mantle regions with differences in composition are very likely to have differences in their mineral fabric and/or on the degree of melt interconnectivity at the source region, whereas two places in the mantle with exactly the same mineralogical composition, and in which magma production rates are equal, can still have different mineral fabrics and melt interconnectivity due to differences in the local pattern of mantle deformation. Consequently, the type of volcanic system that develops at each of these two locations will be different in both cases.

The model of volcanic systems also evidences aspects of polygenetic and monogenetic volcanism that seem to be completely unrelated to plate tectonics, like for instance the volumes of magma associated with each style of volcanism. In particular, Connor and Conway (2000) have mentioned that many monogenetic fields have total volumes of magma comparable to those of polygenetic volcanoes. This is no surprising since total volumes of magma are controlled by large scale mantle motions, as implied by plate tectonics, and therefore it becomes immediate to accept that two volcanic provinces that are located in similar tectonic settings are likely to erupt equivalent melt volumes irrespective of the style of volcanism occurring on the surface. However, according to the model developed in this paper, the total volume of magma erupted by one monogenetic volcano depends on how long it is possible to sustain the pressure at the entrance of the conduit within the zone of melt storage rather than on total (regional) melt production. Due to the rapid nature of porous flow during the initial stages of an eruptive episode, the surrounding matrix is likely to experience large changes that, in general, will block the channel network that feeds the main conduit. Therefore, the volume of magma tapped by most monogenetic eruptions is likely to be relatively small, although the occurrence of larger volume eruptions remains a valid possibility. This contrasts with previous explanations in which the range of volumes erupted by individual monogenetic volcanoes, from the small cinder cones to the huge volumes of magma erupted through fissure vents, remained difficult to explain (Nakamura, 1977; Takada, 1994c).

The influence of tectonic stress in the style of volcanism also has been considered by previous models (Nakamura, 1977; Takada, 1994c) and deserves further examination to the light of volcanic systems. The observation that two equivalent tectonic scenarios can host contrasting styles of volcanism proves that tectonic parameters are not the only ones controlling the style of volcanic activity around the globe. Undoubtedly, tectonic activity is likely to influence the general location of volcanism by controlling the general location of mantle anomalies that remain fixed for periods of time $\sim 10^6$ years, and by setting the background state of stress that is found at a given locality. As remarked in our model, however, the differences on the lithospheric state of stress controlling whether monogenetic or polygenetic volcanism dominates the local style of activity are not necessarily associated to recent tectonic events. Therefore, from the point of view of a volcanic system, the style of volcanism is relatively independent of the tectonic scenario. This conclusion contrasts with that reached by Takada (1994c), who proposed a direct association of tectonic regime with style of volcanism. His model relies on the mechanical interaction between liquid-filled cracks (dikes) during magma ascent, but direct application of the crack-interaction theory (Takada, 1990, 1994a) to explain the occurrence of polygenetic volcanism is very limited in a geological context. This limitation is imposed by the time intervals elapsed between two successive injection events, time that might exceed the time required for magma to solidify in a dike as suggested by periods of repose of $\gtrsim 200$ years that have been documented both in polygenetic volcanoes and monogenetic fields. Additionally, the crack-interaction theory can not be used to explain why is that many monogenetic fields are formed either as isolated provinces or on top of a larger volcanic structure. On the contrary, all the shortcomings faced by the crack-interaction theory are avoided when applying the concept of volcanic systems. Accordingly, there are two distinctive situations in which a monogenetic field can be formed (scenarios 1 and 3). The monogenetic fields that are not related to any large structure are characterized by having magma sources at mantle or deep-crustal depths, whereas the monogenetic fields associated with polygenetic volcanoes will have, in general, magma sources comparable to those of the magmas erupted by the larger structure. These differences arise as the consequence of formation of polygenetic volcanoes envisaged as the long term result of the evolution of a volcanic system with specific conditions that promote ponding of magma at depth, but in which monogenetic eruptions are still possible given the exact combination of magma availability-overlying-rock-stress at any given time.

Moreover, the concept of volcanic system allows us to appreciate better the effects of a changing state of stress on polygenetic volcanic structures. Indeed, it is known that every magma excursion out of a magma reservoir induces a change of the local stress field that results from the opening of the dike transporting the magma (Rubin, 1993). This change will be accumulated through time, and depending on the frequency of the excursion events and on the mechanical properties of the country rock it can lead to important changes on the orientation of the stress field. Such changes form the basis of the formation of coherent dike complexes on large polygenetic volcanoes, ultimately controlling the development of rift zones on these volcanoes (Walker, 1993b). Under extreme circumstances, these changes also can result on a flipping of the minimum stress from the horizontal to the vertical, and back to the horizontal, that has been associated with the cyclic formation of radial dikes and inclined cone sheets on the same polygenetic volcanic system (Walker, 1993a).

In any case, it is noted that, as the database of geophysical, geochemical and chronological information of volcanic fields increases, it will be possible to develop a more deterministic understanding of the various features of volcanism than that outlined in this section. Nevertheless, although we still lack qualitative details of the various aspects distinctive between monogenetic and polygenetic volcanism, the concept of volcanic systems provides a simpler explanation for most of them than previous models.

4.3. Volcanoes at odd locations

Some volcanic centers do not fit very well in any of the three tectonic settings where volcanism is usually expected. For example, the volcanic centers of the Basin and Range province of western North America and the young volcanoes on the Tibetan Plateau present "stimulating problems" that should encourage "fresh insights into volcanism and tectonics" (Francis, 1995). To this category also belong volcanoes that form a secondary volcanic chain, ("back-arc" volcanoes) near to subduction zones (e.g., Hasegawa and Zhao, 1994), volcanic fields that are far enough from the plate margins to be considered part of the rift system and whose location is considered enigmatic (e.g., Novak et al., 1997), and many seamounts that are formed away from the mid-ocean ridge or hot-spot (e.g., Batiza, 1982; Wessel and Lyons, 1997). Further, when stringent criteria are used to evaluate the effects of mantle plumes, most hot-spots remain "unexplained" (Anderson, 2003), and consequently also belong to this category. All of these anomalous volcanic centers can be explained very easily by the concept of a volcanic system by making explicit two conditions that are indispensable for the occurrence of volcanism at the surface: there must be some melt produced and stored at depth, and there must be enough pressure to break the rock above the melt allowing it to reach the surface. Within the context of a volcanic system it is clear that both of these conditions must be met *simultaneously*, and it is possible to explore some of the consequences of their occurrence. Although in the long run plate tectonics determines whether these two conditions are satisfied at a given time in a given place, changes on the location of mantle anomalies (melt source regions) takes place at a different rate than on the local state of lithospheric stress for two reasons. First, the rheological contrast between asthenospheric and lithospheric materials results in the "healing" of old fractures and so the stress memory of the material is different. Second, the local stress of the lithosphere is more readily influenced by changes produced by the volcanic system itself. For these reasons, it is not surprising to find sites where the crust is weakened by fractures and over which there is volcanism next to sites that are equally fractured but over which there is no volcanism at all (Alaniz-Alvarez et al., 1999; Contreras and Gómez-Tuena, 1999; Siebe et al., 1999; Sutter, 1999).

Another consequence of the different time-scales associated with plate tectonics on the one hand and volcanic systems on the other is that volcanic activity can be observed in places where it is somewhat unexpected from the point of view of plate tectonics. For instance, an example of how the slow evolution of mantle anomalies might affect volcanism around a region where lithospheric stress remains constant over time is observed on the Chyulu volcanic field, Kenya. This volcanic field has formed away from the African rift, and can be divided in two groups of volcanoes, each group with distinctive characteristics (Novak et al., 1997). Most notably, the oldest lavas forming the northwestern part of the field were erupted from lower crustal depths (as suggested by the presence of xenoliths), whereas the younger lavas erupted on the southeastern part of the field are more consistent with a shallower source. The difference in estimated depths of the source regions of both groups of volcanoes is about 70 km, which, assuming a rate of ascent of mantle material of ~ 10 cm/year, would be achieved in about 700 000 years, in agreement with the overall ranges of activity observed in both groups. Thus, not only the overall location of the field away from the rift proper, but also the fine structure of the field can both be explained as the result of the slow evolution of the mantle anomaly when inserted in the global context of a volcanic system.

5. Concluding remarks

Throughout this paper, it has been stressed that volcanism above any zone where magma is produced will occur only if conditions for rock fracturing exist, therefore allowing rapid ascent of magma to the surface. In other words, the occurrence of volcanoes always signals the presence of regions of lithospheric weakness above sites of melt accumulation. Tectonic variables are held responsible for the occurrence of even small volumes of melt around the world, but it is noted that the processes controlling melt production depend both on the regional tectonic history and on local aspects that might include lithospheric and mantle inhomogeneities. These processes have a characteristic time-scale of the order of 10^6 years, which is much longer than the characteristic time-scale of volcanic processes. By acknowledging the disparity between the rates of tectonic and volcanic processes, it becomes easier to understand that the slow changes associated with plate tectonics and mantle movements result in volcanic manifestations only a long time after a notable change in the conditions of plate motions has occurred, therefore leading to "unexpected" volcanic activity in some occasions. However, to fill the gap left between the characteristic time-scales of either phenomena, it becomes necessary to consider the combined effects of processes that take place from the moment of magma generation to the moment of eruption. The model of "Volcanic Systems" provides a convenient framework that helps us to visualize the relations among these processes, and also helps us to better define the role played by plate tectonics in controlling global volcanism. The predictions of this model, as updated in this paper, allow us to reconcile apparently contradictory observations in a much easier way than it is trying to force a preconceived explanation of volcanism to the light of only three tectonic settings. We have illustrated this point discussing examples of how the simple relationships highlighted in our model can explain volcano spacing along two types of tectonic margins, differences in the style of volcanism and the occurrence of volcanism at "odd" locations. Although our discussion was restricted to these three aspects of volcanism, it is noted that many other features of volcanoes can be explained with this single, coherent model. For this reason, it is considered that "Volcanic Systems" is the blueprint that should be used to guide future "fresh insights" required to understand in a quantitative sense the global aspects of volcanism, and their relation with plate tectonics.

Acknowledgements

We thank the many questions asked by numerous students throughout the years, questions that, sometimes unwillingly, stimulated an integral view of volcanic systems. We also thank the scepticism of many colleagues that still remain unconvinced by the arguments developed in this paper, for that will be a good reason to continue our quest for a better understanding of volcanic phenomena. Last, but not least, we thank Jean de Bremond d'Ars and Martin Flower for their comments to a previous version, comments that undoubtedly resulted in the improvement of the presentation.

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