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"Dynamics and structural evolution of collapse calderas: A comparison between field evidence, analogue and mathematical models"

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PART I:
INTRODUCTION

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1 VOLCANOES AND COLLAPSE CALDERAS

I.1.1 What is a volcano?

Volcanoes are merely points at the Earth's surface or on the sea floor where magma, and occasionally also non-magmatic material, flows out. This molten material is generated in the Earth's interior, in the upper mantle or crust. Generally, the accumulation of the extruded materials around the eruptive vent may lead to the construction of positive relieves of diverse morphologies (e.g. volcanic shields, stratocones, etc.).

I.1.2 Factors controlling eruptive cycles

Each volcanic eruption is the culmination of a complex series of geological processes controlled by diverse factors that will finally influence, and in some cases determine, the nature of an eruption at surface (Fig. 1.1). These processes may be grouped in (Figs. 1.1 and 1.2)(Marsh, 1989):

- **Magma genesis**
- **Magma transport**
- **Magma storage**
- **Magma chamber differentiation and eruption**

Therefore, although volcanoes represent very short time periods at geological scale or even at human scale (from days to few thousand years), they are the response or culmination of geological processes that last hundreds of thousands to millions of years (e.g. magma genesis) (Fig. 1.1). This phenomenon is also observable in the spatial scale. Then, whereas magma genesis takes place at Earth plate scales, magma transport to

shallow levels is restricted to a more regional level. Finally, the volcano and, consequently, the eruptive products, define the local scale.

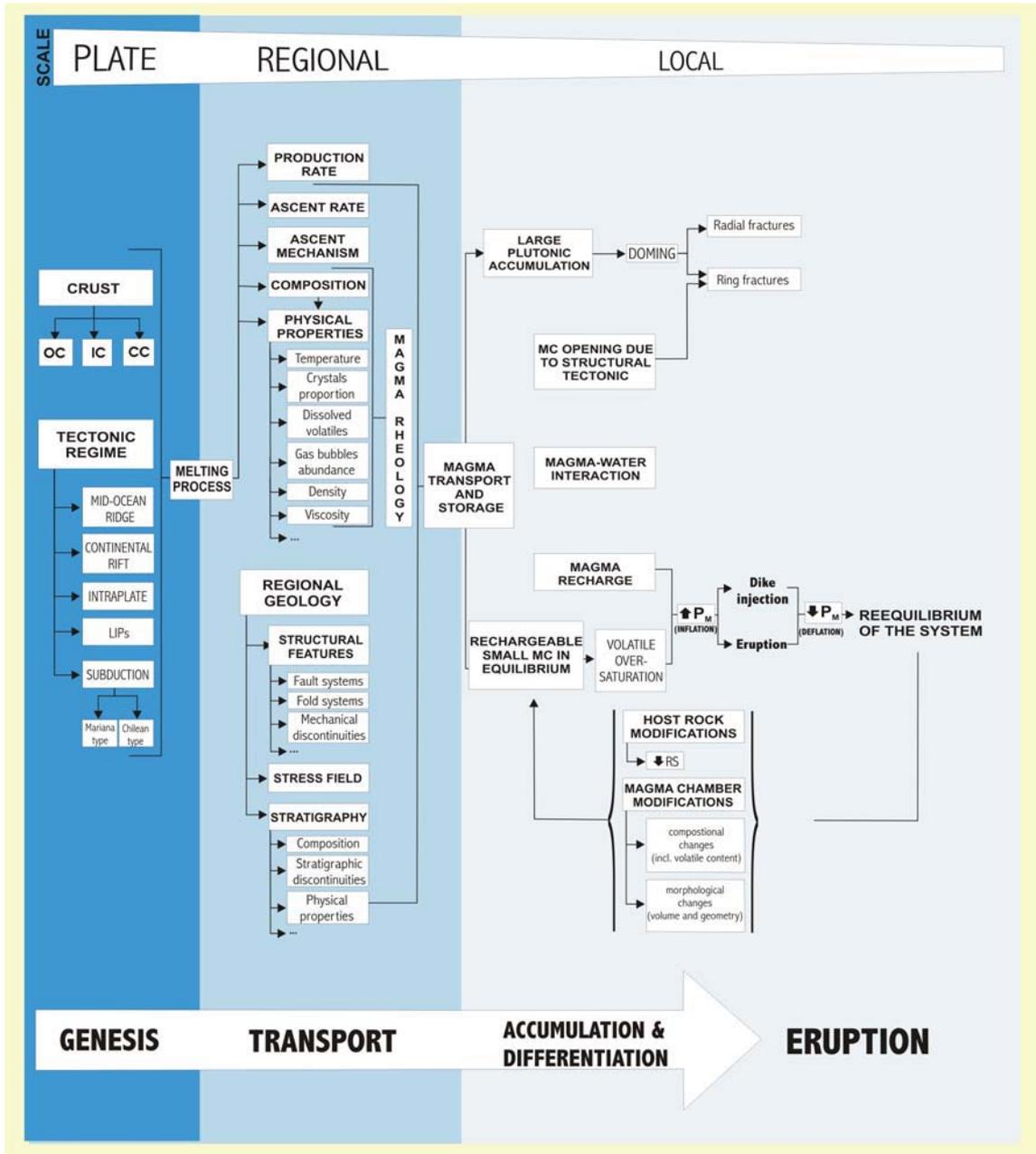


Fig. 1.1: Spatial scale and relationship between those geological processes that influence and determine the nature of an eruption at surface. These processes are controlled by specific factors also represented in the sketch. **OC** Oceanic crust; **IC** Intermediate/transitional crust; **CC** Continental crust; **LIPs** Large igneous provinces; **MC** Magma chamber; **P_M** Magmatic pressure; **RS** Rock resistance.

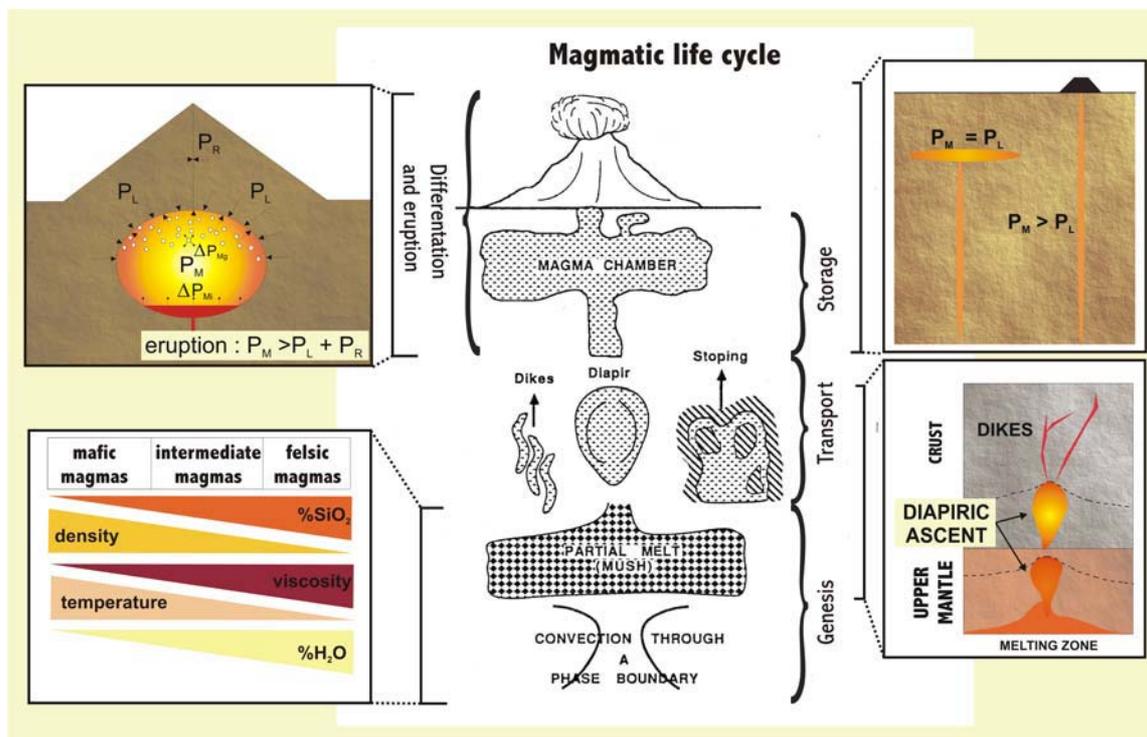


Fig. 1.2: Sketch of the processes involved in eruptive cycles. (Modified from Marsh, 1989) **(Bottom and left)** Influence of the composition on the physical properties of magmas. More felsic magmas with high SiO₂ content are normally more viscous and may have higher water content. By contrast, magma density and temperature are lower. **(Bottom and right)** Mechanisms of magma ascent: through diapirs when the surrounding rocks are able to deform plastically and through dikes when the host rock is brittle. **(Top and left)** Triggering mechanism of explosive eruptions. Pressure inside the magma chamber (P_M) may increase by an amount (ΔP_{Mg}) due to volatile oversaturation or to intrusion of fresh magma (ΔP_{Mi}). If the chamber pressure exceeds the lithostatic pressure (P_L) plus the tensile strength of the surrounding rocks (P_R), tensile vertical fractures may open allowing magma to escape from the chamber, sometimes reaching the surface to cause a volcanic eruption. (Modified from Marti and Folch, 2005) **(Top and right)** Mechanism of shallow magma chambers formation. Diapiric ascent is governed by the principle of buoyancy: the hot magma continues its ascent until it is less dense than the surrounding rock, when host rock and magma have the same density, the latter stops and begins to spread laterally and to accumulate. The magma ascent through fractures is controlled by pressure difference between the magmatic and the lithostatic pressure. Magma ascent continues until the magma pressure P_M equals the lithostatic pressure P_L .

❖ MAGMA GENESIS

The first step in an eruptive process is magma genesis. Almost all magma is generated by the partial melting of source rocks, either in the upper mantle or crust (e.g. Bateman, 1984; Petford et al., 2000). Although, magma genesis is more effective in areas where plates are created (divergent plate boundary) or destroyed (convergent plate boundary), there exists also intraplate volcanism (continental and oceanic), which demonstrates that a more localized rock melting is possible (Carey, 2005). Consequently, one of the most important variables in magma genesis is the tectonic setting (Fig. 1.1). Figure 1.3 summarizes the main tectonic contexts where volcanoes (i.e. calderas) may develop.

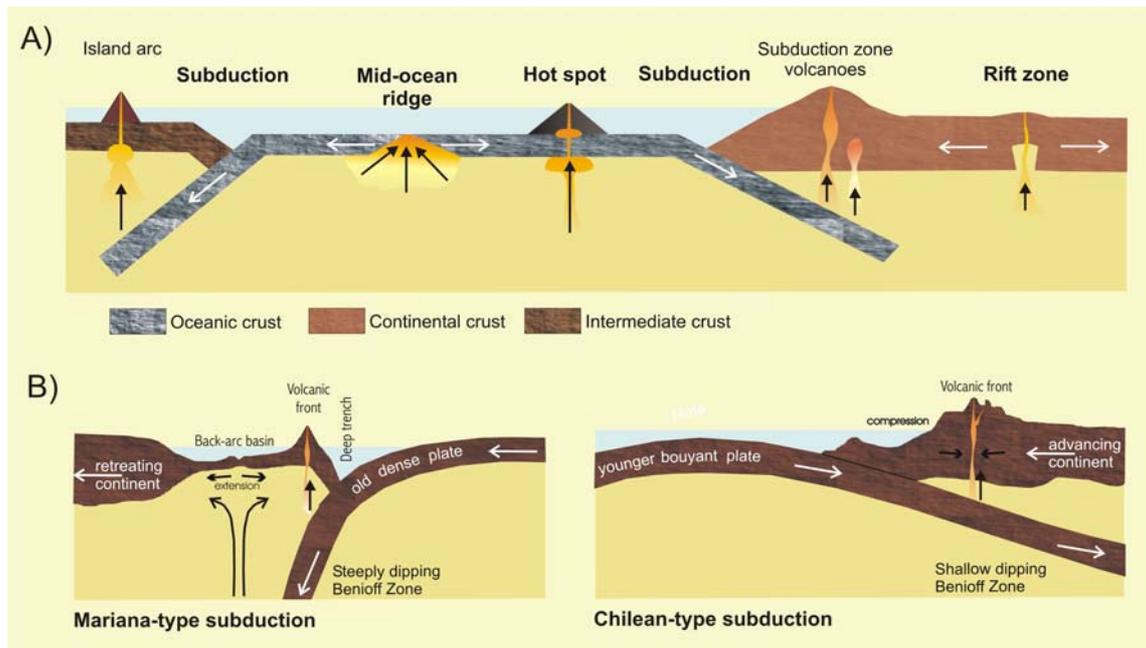


Fig. 1.3: (A) Summary of the different tectonic contexts where magma genesis and related volcanism takes place. (B) Summary of the differences between the Mariana-type and Chilean-type subduction modes based on the model of Uyeda (1982). Mariana-type subduction involves steeply descending plate, whereas in the Chilean-type the subduction angle is shallower. (Modified from Carey, 2005)

Besides the tectonic regime, the type crust (oceanic, continental or intermediate crust) involved in the melting process also affects the composition of the resulting magma and consequently, its physicochemical properties (Fig. 1.2). In short, more felsic magmas with high SiO_2 content are normally more viscous and may have higher water content. By contrast, magma density and temperature are lower. For his part, the melting process will determine the production and ascent rate, and the composition and physical properties of magma both parameters affecting the magma transport process (Fig. 1.1).

❖ MAGMA TRANSPORT

The second main process is magma transport, which occurs in a regional scale. Magma transport to shallow levels (incl. to the surface) may take place through two different mechanisms (Fig. 1.2). It may ascent through diapirs when the surrounding rocks are able to deform plastically (i.e. they do not get broken due to the magma upthrust) (e.g. Bateman, 1984; Weinberg and Podladchikov, 1994) or magma may flow up to the surface in narrow conduits, either as self-propagating dykes (Clemens and Mawer, 1992; Clemens et al., 1997; Petford et al., 2000), along pre-existing faults

(Petford et al., 1993; Petford et al., 2000) or as an interconnected network of active shear zones and dilatational structures (Lemos et al., 1993; Collins and Sawyer, 1996). In short, magma ascents in a diapiric way until the rocks located at depth are relatively plastic and through dikes when rocks become more brittle in shallower levels (Lister and Kerr, 1991; Rubin, 1995; Petford et al., 2000; Martí and Folch, 2005).

Both ascent mechanisms are strongly dependent on the magma rheology, which is determined by the geochemical composition and the physical properties of magma (e.g. temperature, proportion of crystals, amount of dissolved volatiles, and the abundance of gas bubbles) (Bateman, 1984). Principally, magma dynamic is controlled by the density and viscosity, properties determined by the composition, the temperature of magma and the water content (Fig. 1.1) (McBirney and Murase, 1984; Pitcher, 1993).

Further factors that control the transport of magma to upper crustal levels are the host rock rheology, the regional geology and the stress field (Rubin, 1993; Watanabe et al., 1999). The regional structural features (e.g. fault systems, mechanic discontinuities, etc.), the regional stress field (e.g. compressive, extensive) and the regional stratigraphy (i.e. composition and physical properties of host rocks and stratigraphic discontinuities) play an important role in the magma ascent and storage.

Additionally, the magma production and ascent rate, as well as the ascent mechanism, are factors controlling the magma transport in a regional scale but are partially controlled by plate scale variables (melting process).

❖ **MAGMA STORAGE**

The next step, occurring in a transition between the regional and the local scale, is the storage of magma in shallow levels and its posterior differentiation inside the magmatic reservoir (Fig. 1.1). Although, this is the most common case, direct eruptions of magma from deep reservoirs may also take place (Fig. 1.2).

The creation of these shallow magma chambers is controlled by a combination of mechanical interactions (either pre-existing or emplacement generated wall rock structures), and density effects between the spreading flow and its surroundings, i.e. the principle of neutral buoyancy: a diapir will be allow to rise, as long as the magma density is less than the surrounding rocks (Lister and Kerr, 1991; McLeod, 1999). As magma rises into the upper levels of a volcano and attains neutral buoyancy, it will

stagnate and begin to spread out laterally within a neutral buoyancy horizon (Fig. 1.2). This horizon is expected to have limited vertical extent but a more extensive lateral expanse. The horizon of neutral buoyancy is not a static level during the growth of a volcano. As eruptions occur and volcanic centres grow in elevation the density profiles will adjust to the new overlying load (Pinel and Jaupart, 2004). The neutral buoyancy level will rise in an attempt to keep pace with the density adjustments of the edifice.

Furthermore, the magma ascent through fractures is controlled by pressure difference between the magmatic and the lithostatic pressure. Magma ascent continues until the magma pressure equals the lithostatic pressure (Fig. 1.2). If a vertical dyke does not reach the Earth's surface, magma either gets or propagates horizontally.

Magma chambers play a fundamental role in the way in which volcanoes erupt because they act as the source reservoirs for eruptions at the surface. In fact, the size of the eruption is controlled by the eruptible magma inside the chamber (Carey and Sigurdsson, 1989).

❖ **MAGMA CHAMBER DIFFERENTIATION AND ERUPTION**

Once the magmatic reservoir has formed, factors controlling the eruption are in a local scale. The type of magmatic reservoir and its evolution, as well as the effects of its emplacement on the local state of equilibrium play a fundamental role in the way in which volcanoes will erupt (Carey and Sigurdsson, 1989). Simplifying, there exists to magmatic reservoir end-members: large plutonic accumulations and small rechargeable magma chambers.

Large plutonic accumulations generally produce local (magma chamber size) doming (Smith and Bailey, 1968). In most cases, local tumescence leads to the appearance of radial fractures at surface and, in more evolved stages, of incipient ring fractures (Komuro et al., 1984; Komuro, 1987; Pollard and Segall, 1987).

The evolution of small rechargeable magmatic reservoirs is completely different. Small magma chambers live episodes of pressure increase (inflation) and pressure decrease (deflation) inside the magma chamber over extended periods of time. Inflation of the chamber has principally two origins (Fig. 1.1): the injection of new magma into the reservoir (i.e. the magmatic system is open) (Sparks et al., 1977; Blake, 1981; Tait et al., 1989; Folch and Martí, 1998; Troll et al., 2000) or by an increase of the volatile

content (volatile oversaturation) due to the differentiation process (Smith and Bailey, 1968; Blake, 1984; Tait et al., 1989; Martí and Folch, 2005). In several cases, a combination of both factors is also possible. Occasionally, other mechanisms such as seismic excitation of entrapped bubbles (Sturtevant et al., 1996), tectonic triggers (Linde and Sacks, 1998) and the interaction of water with magma (e.g. Inyo Craters – *U.S.A.*, Mastin, 1991; Ambrym – *Vanuatu*, Robin et al., 1993) have also been recognized. Whatever the case, the magma chamber responds to restore the mechanical equilibrium with its surroundings by injecting dikes or deforming the wall rocks (Phillips, 1974). If these magma-filled fractures reach the surface, they trigger an eruption. The overpressure at which fracture occurs is determined by the tensile strength of the surrounding rocks (necessary force to break the magma chamber roof by tension) (Tait et al., 1989). The withdrawal of a few percent of the stored magma leads to a pressure decrease (deflation) inside the magma chamber. Once the mechanical equilibrium has been restored, the eruption ceases, the system is reequilibrated and prepared for the next eruptive cycle (Martí and Folch, 2005). The remaining magma can continue to cool and evolve, and a new eruptive event can occur if conditions for overpressure are again reached (Tait et al., 1989). This eruption can be the initial phase in the evolution of a stratovolcano, which is built up over extended periods of time by a large number of eruptive events.

The evolution of the volcano throughout repeated eruptions and tectonic adjustments can produce variations in the bulk characteristics that combine to determine the tensile strength of the host rock, i.e. its resistance (Tait et al., 1989). If the volcano is weakened, in the next eruptive cycle the required over or underpressure required to break the roof is smaller. Furthermore, the withdrawal of magma from the reservoir causes inside the chamber compositional changes (Leat et al., 1984) (e.g. variations in the volatile content) and morphological changes (i.e. volume and geometry) (Martí and Folch, 2005). These modifications in the system may affect the eruptive conditions and possibly style of the next eruptive cycle (Martí and Folch, 2005).

The period between consecutive eruptions is known as the repose period and is typically of the order of a few years to a few thousands of years.

Eruptions may be effusive (magmas with a low volatile content) or explosive (magma with a high gas content). The latter eruptions are the most violent ones and are principally associated with evolved magmas (e.g. Teide – *Tenerife*, Ablay, 1997; Aira – *Japan*, Aramaki, 1984).

I.1.3 Collapse calderas

Collapse calderas are defined as the volcanic depression that result from the disruption of the geometry of the magma chamber roof due to down faulting during the course of an eruption (Fig. 1.4). The diameter of these collapse depressions, more or less circular in form, are many times greater than the diameter of the including vents (Williams and McBirney, 1979; Lipman, 2000). In the definition, we do not include apical craters also named calderas, which result from the collapse and subsequent blocking of the volcanic conduit without affecting the roof of the magma chamber.



Fig. 1.4: Some examples of calderas worldwide. **(A)** Crater Lake caldera, United States (Photograph: <http://www.geol.lsu.edu>). **(B)** Silali caldera, Kenya (Photograph: <http://www.volcano.si.edu>). **(C)** Katmai caldera, Alaska (Photograph: <http://www.skimountaineer.com>)

I.1.4 The importance of calderas

I.1.4.1 General aspects

The relationship between people and volcanoes is as old as human race. Since volcanic eruptions are the most awesome and powerful display of nature's force, some volcanic processes constitute a major natural hazard (a single eruption can claim thousands of lives in an instant) (e.g. Blong and McKee, 1995; Sigurdsson, 2000; Blong, 2000, 2003; Witham, 2005). By contrast, other processes associated with volcanic

activity are highly beneficial to society (e.g. fertile soils, ore deposits, etc.) (e.g. Guillou-Frottier et al., 2000; Ping, 2000).

Caldera-forming events are rare (Mason et al., 2004), good examples are Jemez Mountains, where major ignimbrite eruption and caldera formation happened twice at an interval of nearly 10^6 years (Smith and Bailey, 1968). By contrast, small calderas may be produced by vents of a magnitude that recur relatively frequently. A good example is Fogo caldera on Sao Miguel (Booth et al., 1978). Caldera collapse there may not be a rare event but many recur at intervals of about 10^4 years (Walker, 1984). In fact, calderas have formed at the rate of about three per century since 1783 (Walker, 1984).

Although their low frequency of occurrence, collapse caldera are the most destructive volcanic events. In fact, large pyroclastic eruptions and associated caldera collapse structures represent one of the most catastrophic geologic events that have occurred on the earth's surface in the Phanerozoic time and more important in the historical time (e.g. Tambora – *Indonesia*, Self et al., 1984; Krakatau – *Indonesia*, Self and Rampino, 1981; Pinatubo – *Philippines*, Hattori, 1993).

Collapse calderas have received considerable attention due to their link to Earth's ore deposits and geothermal energy resources, but also because their impact on the environment (e.g. climate) and the human society (Lipman, 2000). Moreover, since volcanoes and their eruptions are merely the surface manifestation of the magmatic processes operating in the Earth interior, calderas also provide key insights into the generation and evolution of large –volume silicic magma bodies (Lipman, 2000).

Thus the study of the volcanism is highly interdisciplinary, most closely linked to geophysics (e.g. Davy and Caldwell, 1998; Milner et al., 2002; Troise et al., 2003), petrology (e.g. Francis et al., 1989; Edgar et al., 2002; Troll et al., 2000) and geochemistry (e.g. Leat et al., 1984; Francis et al., 1989; Edgar et al., 2002).

1.1.4.2 Calderas and volcanic hazard

1.1.4.2.1 General aspects

Similar to other volcanic activities, caldera-forming eruptions can have both positive and negative consequences on the environment and human society. Obviously, eruptions involving caldera collapse represent an enormous risk to people and property,

due to their tremendous destructive potential and their effects on global and local climate. By contrast, the volcanic material erupted generates some of the world's most fertile soils and agricultural areas (Lipman, 2000).

The various volcano hazards differ markedly, in the spatial and temporal scale, in their impact on the physical environment and on people. Short-term and long-term consequences of volcanic eruptions and especially of caldera-forming events have killed more than 300.000 people since A.D. 1000 (Self, 2005; Witham, 2005).

❖ **SHORT-TERM EFFECTS (DIRECT VOLCANO HAZARDS)**

Hazardous events that are produced during or shortly after the eruption (minutes-days) are considered to be “direct” volcano hazards (Tilling, 2005). These include: fall processes of ash (e.g. Blong and McKee, 1995) and ballistic projectiles (e.g. Anderson and Flett, 1903), flowage processes (e.g. pyroclastic flows and surges, etc.) (e.g. Fisher et al., 1980) and other processes like phreatic explosions (e.g. Feuillard et al., 1983) or volcanic gases and acid rain (e.g. Kling et al., 1987; Farrar et al., 1995).

❖ **LONG-TERM EFFECTS (INDIRECT VOLCANO HAZARDS)**

Those destructive processes that are incidental to the eruptive phenomena themselves are considered “indirect” volcano hazards (Tilling, 2005). These include: earthquakes and ground movements (e.g. Shimosuru, 1972), tsunamis (e.g. Simkin and Fiske, 1983), secondary debris (e.g. Rodolfo et al., 1996) and pyroclastic flows (e.g. Torres et al., 1996), atmospheric effects (COMVOL, 1975), climate changes (e.g. Stommel and Stommel, 1983; Self et al., 1996), post-eruption famine and disease (e.g. Thorarinsson, 1979).

Possibly one of the most interesting volcanic hazard related to caldera-forming eruptions are the effects of these on global climate and further atmospheric effects (e.g. Self et al., 1981; 1996). Historical caldera-forming eruptions such as Tambora in 1815 (Self et al., 1984; Newhall and Dzurisin, 1988), Krakatau in 1883 (Self and Rampino, 1981; Simkin and Fiske, 1983; Newhall and Dzurisin, 1988), and Pinatubo in 1991 (Hattori, 1993; Lipman, 2000; Darteville et al., 2002), although several orders of

magnitude smaller than the largest prehistoric eruptions appear to have modified global climate, including visible atmospheric effects such as strange colours and halos on the sun and moon, vivid sunsets and sunrises, and anomalously cold weather. During these large pyroclastic caldera-forming eruptions were generated fine-grained ash and sulphur aerosols on vast scales and injected into the upper atmosphere. Besides from the volcanic ash effects, the dominant influence is believed to be the injection of sulphur aerosols into the stratosphere, a process whose efficiency varies in relation to magmatic sulphur content and height of the eruption column (Self et al., 1981; Kazahaya et al., 2004). The global climate impact of large eruptions is a major reason that an understanding of volcanic phenomena is important for society today. A Tambora-size eruption in the 21st century will have much more profound effects on human living on an overcrowded Earth, where natural resources are strained to the limit (Sigurdsson, 2000).

❖ **EFFECTS OF HISTORICAL CALDERA-FORMING ERUPTIONS AND THEIR CONSEQUENCES**

The powerful impact of caldera-forming eruptions on the environment and society has been demonstrated by several historical eruptions. Two important well-studied examples are the eruption of Mt. Pinatubo in 1991 and of Tambora in 1815.

➤ **Pinatubo 1991**

A well-studied historical example is the Mt. Pinatubo eruption in 1991. It was a plinian style eruption with a column over 35 km in height that produced 5-10 km³ of dacitic pumice and ignimbrite and formed a 2.5 km diameter caldera (Fig. 1.5) (Self, 2005).

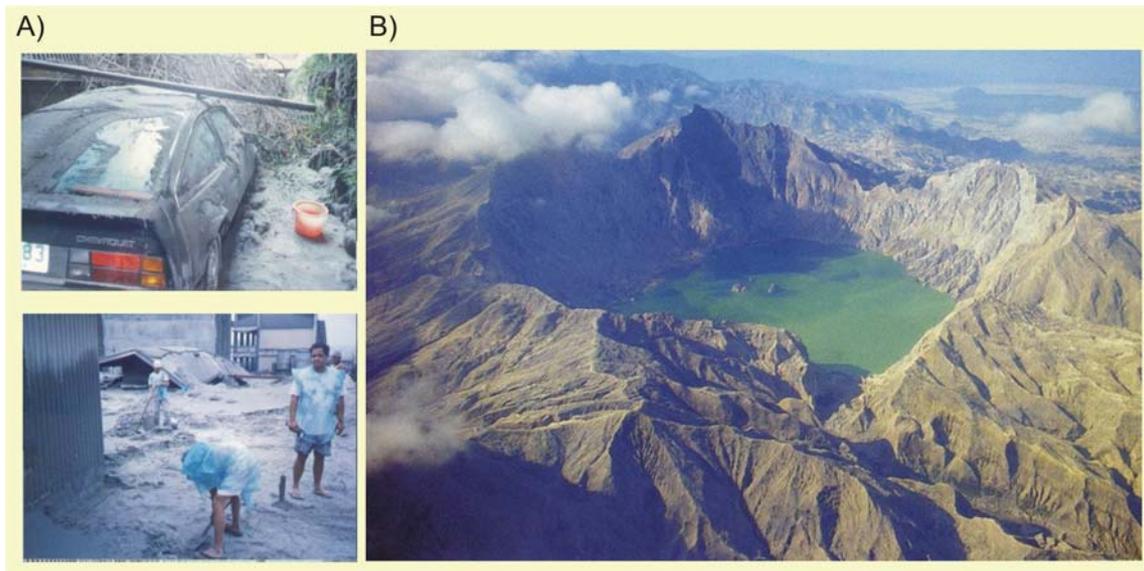


Fig. 1.5: (A) Effects of the fined-grained ash generated during the eruption of Mt. Pinatubo in 1991 (Photograph: <http://faculty.weber.edu>) (B) The collapse structure at the summit part of the volcano generated during the climactic eruption (Photograph: <http://www.volcano.si.edu>).

During the eruption, large amounts of water, gasses (e.g. CO_2 , SO_2 , etc.) and ash were ejected into the atmosphere (Fig. 1.6 and 1.7) (Westrich and Gerlach, 1992; Self, 2005). In fact, the volcano emitted the largest SO_2 cloud ever observed by modern measurements (Bluth et al., 1992). These ash and gasses were removed into the stratosphere where sulphur dioxide converted into sulphuric acid that condensed into sulphate aerosols. The generated aerosol veil hung over the Earth to more than 18 months causing spectacular sunsets and sunrises worldwide (Self et al., 1996) (Fig. 1.6). The presence of these kind of aerosols enhances the reflectance of sunlight back into space, i.e. the Earth surface and the lower atmosphere cool (Fig. 1.7). The cooling process is enhanced by the presence of large amounts of volcanic ash in the atmosphere blocking sunlight, but since ash settles out relatively quickly, in days to months, it has only a short term cooling effect. Furthermore, the sulphates then absorb the heat-radiated from the Earth and warm the stratosphere and also accelerate chemical reactions, which destroy the ozone, allowing dangerous levels of ultraviolet radiation to reach the Earth's surface. Furthermore, the injection of greenhouse gasses into the atmosphere enhances the greenhouse effect (Fig. 1.7) (Darteville et al., 2002).

Long-term effects of the Pinatubo eruption in 1991 were first a process of cooling of the Earth during a period of 1 to 3 years (Fig. 1.6) and afterwards the additional carbon dioxide remains to potentially increase global warming.

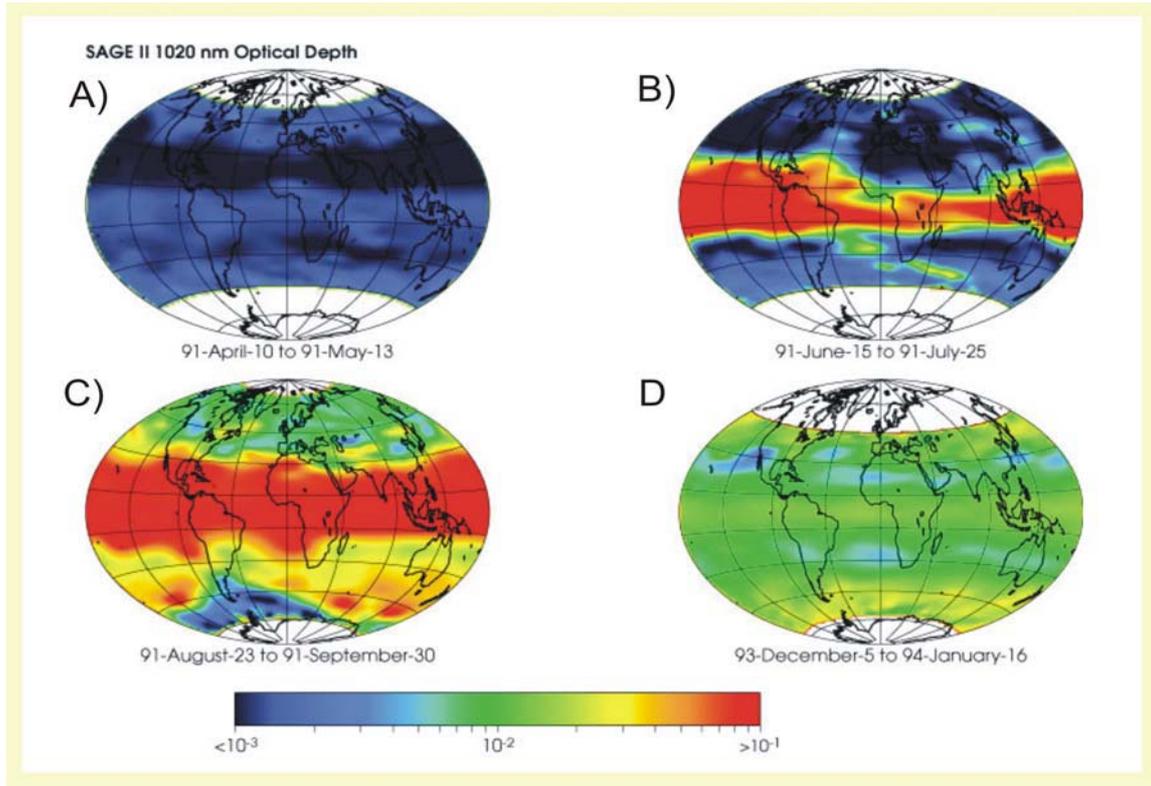


Fig. 1.6: The early summer date of the eruption (June 12-15) and the tropical location of the volcano (15° 07' N 120° 20' E), resulted in a rapid spread of the aerosol cloud around the Earth in about 3 months. (Images: <http://weimages.gsfc.nasa.gov>) **(A)** Aerosols content in the stratosphere before the eruption of Mt. Pinatubo on June 12-15,1991. **(B)** Aerosols content just after the eruption and during the following month. The aerosol cloud has spread worldwide from the equator to both Tropics. **(C)** The aerosol cloud during the subsequent months. It has covered almost all the world. **(D)** Two years later a certain quantity of aerosols remain on the stratosphere.

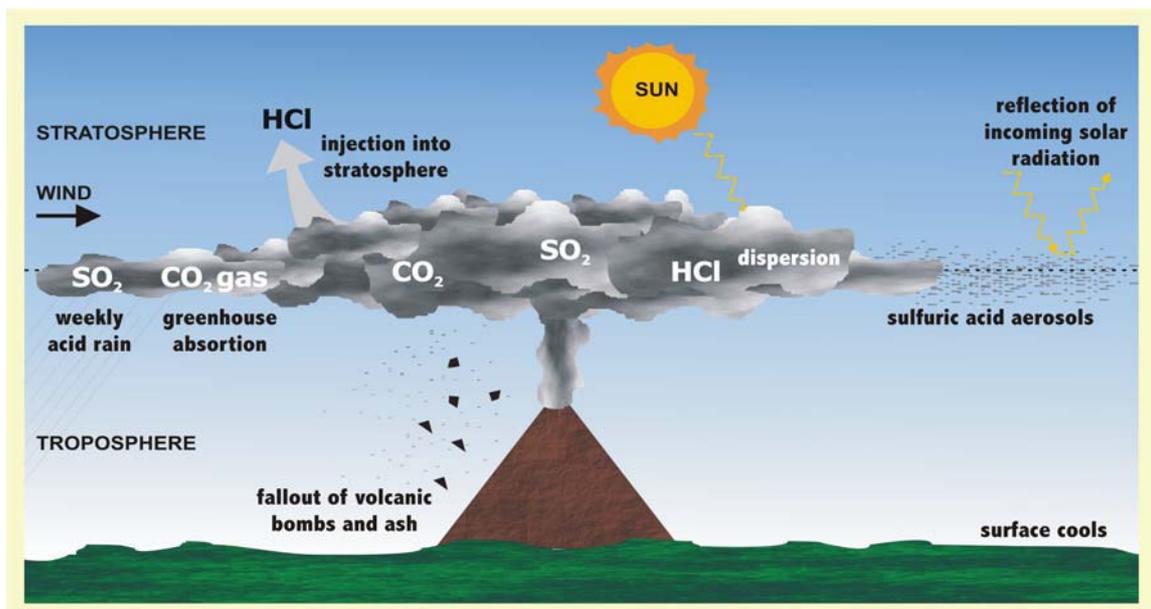


Fig. 1.7: Schematic diagram of the various volcanic inputs to the atmosphere and their interaction, fates, and radiative impact, including the formation of sulphuric acid aerosols. (Modified from Self, 2005 after McCormick et al., 1995 and Robock, 2000)

➤ **Tambora 1815**

A further example is the caldera-forming eruption of Tambora, which took place on the island of Sumbara in Indonesia in April 1815. Approximately, 30 km³ of phonolitic magma was extruded and at least 100Mt of sulphuric acid was generated. This eruption is the largest recognized of the past millennia (Self et al., 1984; Sigurdsson and Carey, 1988a, 1988b).

The disaster struck both on the islands of Sumbawa, where Tambora is located, and on the neighbouring islands (e.g. Lombok, Flores, etc.). Only in Tambora 10,000 people were killed by the pyroclastic flows and the blanket of ash fall out on Sumbawa also destroyed all the crops, resulting in a famine immediately after the eruption (Sigurdsson, 2000). It was accompanied by a variety of diseases and clean water was not available anywhere. For some reason, the island climate changed, becoming so hot and dry, that crop would not grow, probably because of the destruction of the vegetation and higher runoff. An additional 38,000 people died of starvation and disease, having lost 75% of their livestock, and 36,000 fled the island. It took the island at least one century to recover from this ecological and human disaster.

The effects of the Tambora eruption were not only restricted to Sumbawa. Other islands like Lombok and Bali the ash was so thick that many people were killed immediately as the roofs of their homes collapsed under the weight of huge quantities of ash. A conservative number of dead in the East Indies from the Tambora eruption is at least 117,000. The effects of the Tambora eruption produced unusual atmospheric phenomena and climate deterioration set in during 1816, called “the year without summer”, probably the best-known example of volcanically induced climate cooling event (Stothers, 1984; Harington, 1992). The explosive volcanic eruption emitted large quantities of SO₂ into the stratosphere also converted into a sulphuric acid aerosol dust veil that encircled the Earth. The aerosol reduced the solar heat reaching the surface of the planet leading to global cooling. Strange atmospheric phenomena take place in Europe during 1816. Additionally, the generated aerosol veil may have caused cooling up to 1°C regionally and more locally. In fact, the mean temperature in Europe was about 3°C below average across the entire continent. The effects lasted until the end of 1816 and extended to both hemispheres.

The abnormal climate conditions in 1816 have frequently been proposed as an important factor in the first worldwide cholera epidemic.

1.1.4.3 Calderas and natural resources

In addition to potential problems of volcanic hazards, field observations indicate that calderas and their associated volcanic activity are important controlling factors on many ore deposits and geothermal energy resources (Smith and Bailey, 1968; Lipman et al., 1984; Self et al., 1984; Redwood, 1987; Guillou-Frottier et al., 2000; Lipman, 2000).

The hydrothermal activity and the mineralization may occur during precaldera volcanism, or caldera-related postcollapse igneous activity. Furthermore, in some cases, it can be associated with much later activity localized by caldera structures (Steven et al., 1974; McKee, 1979). Examples of calderas with associated ore deposits are worldwide, some of them are summarized in Table 1.1.

Location	Caldera	Age (Ma)	Minerals	Age Min. (Ma)	Eruptive stage	Ref.
<i>Low sulfidation epithermal ore deposits</i>						
Bolivia	Soledad/La Joya	5.4	Au-Ag-Cu-Pb-Zn	15	Pre	[274]
USA	Creede	26.6	Pb-Ag-Zn-Au	26-24	Syn-Post	[30, 197]
USA	Bursum/Mogollon	29-28	Pb-Zn-Ag	17-18	Post	[273]
USA	Valles/Cochiti	1.1	Au-Ag-Mo	6;0.6	Pre and Post	[127]
<i>High sulfidation epithermal ore deposits</i>						
USA	Summitville	29.5	Au-Ag-Cu	22.4	Post	[130]
<i>Porphyry-type ore deposits</i>						
USA	Silverbell	73-69	Cu		Syn-Post?	[193, 199]
USA	Questa	26-18	Mo		Post?	[193, 195]

Table 1.1: Examples of mineralized ash-flow calderas (Modified from Guillou-Frottier et al., 2000).

Furthermore, specific caldera processes may be key factors in the formation of economic mineral deposits like volcanogenic massive sulphide (VMS) deposits (Stix et al., 2003) (Table 1.1). These deposits, rich in zinc, lead and copper, related to silicic volcanic rocks in extensional and transtensional environments (Sillitoe, 1982). Occasionally, show a close spatial and temporal relationship with silicic submarine calderas (Ohmoto, 1978). Particularly favourable structural environments for the

formation of certain VMS deposits are asymmetric calderas. These have a principal fault system on one side of the caldera that facilitates fluid flow and concentration of VMS deposits in a narrow zone. The asymmetric profile of these calderas provides a structural basin suitable to accumulate and preserve VMS deposits (Fig. 1.8). A good example is the Hunter Mine caldera (Mueller and Mortensen, 2002).

During caldera collapse, large-scale structural pathways are created instantaneously opening the magmatic system to heat and mass transfer. The orientations of the caldera faults will determine, in part the openness of the system: inward-dipping faults tend to close as the caldera block subsides, by contrast, outward-dipping structures tend to open the system progressively (Stix et al., 2003). This structural opening can lead to an enhance degasification of the magmatic reservoir, then the released magmatic fluids are able to contribute large quantities of ore metals (e.g. Cu, AU) to the hydrothermal fluid (Yang and Scout, 1996). Once caldera has formed large hydrothermal systems also can develop during resurgence when magma is injected at shallow levels, providing a renewed heat source magmatic input of volatiles and metals and new structural pathways.

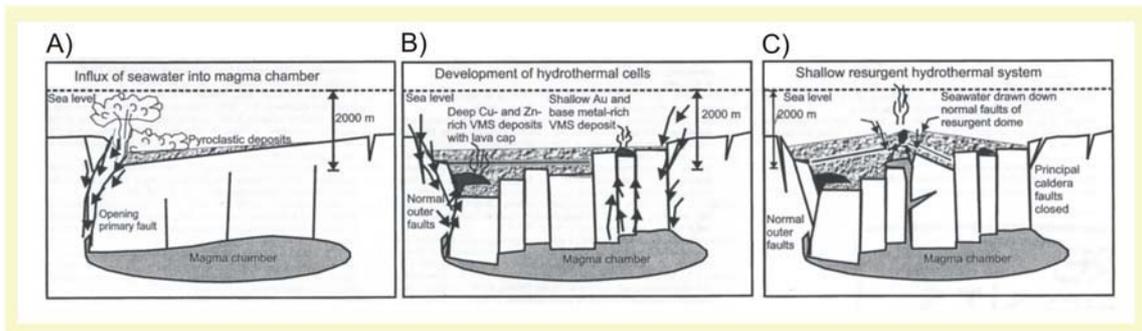


Fig. 1.8: conceptual view in space and time of caldera-related submarine magmatic-hydrothermal system. (Modified from Guillou-Frottier et al., 2000) **(A)** Asymmetric collapse of caldera exposes the underlying magmatic-hydrothermal system. Magma degasses while seawater flows into the system. **(B)** Cold seawater flows downwards through the faults while mineralising fluids heated by magma move up. Upflow and downflow generate hydrothermal convection cell controlled by the caldera structure. **(C)** Renewed magmatism causes resurgence and intrusion into roof rocks above magmatic reservoir uplifting the block rocks.

I.1.5 Historical review

The relation of calderas to pyroclastic activity became evident after the historical eruption of Krakatau in 1883 (Verbeek, 1886). Verbeek noted that the volume of older rock in the deposits was much less than that of the missing structure. Consequently, he rejected the idea that the mountain had been blown away and concluded that most of the mass had foundered into the space evacuated by a rapid discharge of huge volumes of magma. Some years before, Fouqué (1879) inferred from similar reasoning, that the bay enclosed by the islands of Thera and Therasia in Santorini had resulted from collapse associated with a strong pyroclastic eruption. Fouqué concluded that whatever the mechanism, it was the collapse of the roof that had triggered the explosive eruptions. The main explosive discharge then was a consequence, not a cause of the collapse. Furthermore, the formation of calderas by post-eruptive collapse was also recognized by Dutton (1884) during the study of Kilauea Volcano in Hawaii.

At the beginning of the 20th century, some examples of eroded calderas (cauldrons) were first interpreted as subsided blocks associated with igneous activity, e.g. Glen Coe in Scotland (Clough et al., 1909) and the eroded calderas in the Tertiary San Juan volcanic field, Colorado (Burbank, 1933). Subsequent investigations recognized a collapse origin for Crater Lake, Oregon (Diller and Patton, 1902) and identified the Toba caldera in Indonesia and the post-collapse resurgent uplift of its floor (Van Bemmelen, 1939). Few years later, were recognized large Pleistocene calderas associated with voluminous explosive volcanism in southwestern Japan (Matumoto, 1943).

In 1941, Williams presented an insightful analysis of calderas and their origin. He assumed the reverse relation as Fouqué (1879) that most major calderas form by collapse after the eruption of magma during pyroclastic volcanism. In the sixties and seventies, several studies (Smith, 1960, 1979; Smith and Bailey, 1968) provided a conceptual framework for stratigraphic studies of ash flow sheets, a model of collapse resurgence within ash flow calderas, and a petrologic framework for ash flow magmatism. Also important were the studies of pyroclastic deposits (e.g. Sparks and Walker, 1977; Walker, 1980; Wilson and Walker, 1982) that showed the way for

correlating the physical properties of pyroclastic deposits with their emplacement mechanism.

The tremendous progress made since 1960 in the related fields of pyroclastic flows and caldera geology are difficult to summarize. After these pioneering works, collapse calderas have been the subject of studies of diverse disciplines. In fact, at the beginning of the eighties, analogue modelling became a useful tool in the study of caldera collapse processes. The first experiments were performed by Ramberg (1967), who carried out centrifuge experiments using putty (magma chamber) under a roof of clay (host rock). This and posterior works (e.g. Komuro et al., 1984; Komuro, 1987; Martí et al., 1994; Roche et al., 2000; etc.) were focused on understanding caldera collapse mechanism and structure through analogue modelling. The main difference between them was the applied materials to simulate magma and host rock and experimental set-ups to reproduce the magma chamber, the magma ascent or withdrawal and the host rock. These analogue experiments reproduce successfully permanent deformation structures such as fractures and faults during a caldera collapse. The experiments simulate the rigid crust using cohesive, dry, powder mixtures (sand, fused alumina, flour, etc.), while silicone is used to simulated magma or a mantle plastic layer. The obtained results are similar and allow one to reproduce the formation of tectonic structures commonly associated with the formation of collapse calderas. Still now, analogue models are on the increase. Recent works are more focused on studying the influence of regional tectonics (e.g. Acocella et al., 2004; Holohan et al., 2005; etc.) and the effect of pre-existing topography (volcanic edifice).

Simultaneously to these works on analogue modelling, theoretical models based on thermodynamics, solid and fluid mechanics, etc, have progressively become and indispensable tool to study caldera-forming processes. There exists a large list of works focused on caldera collapse processes. These are oriented to different aspects of the collapse processes and may be divided in those models that look for conditions on magma to achieve a certain chamber underpressure (Druitt and Sparks, 1984; Bower and Woods, 1997; 1998; Martí et al., 2000; Roche and Druitt, 2001) and models that determine stress conditions for normal-fault caldera initiation (Komuro et al., 1984; Gudmundsson et al., 1997; Gudmundsson, 1998; Folch and Martí, 2004).

I.1.6 State of the art

After more than a century studying collapse calderas, what do we really know about them?

Field studies have provided a large amount of information concerning caldera-forming eruptions and related structures and deposits. Broadly speaking, field data indicate that volcanic calderas structures occurring in the Earth since ancient times, then there are examples from the Archean (e.g. Ramat Yotam – *Israel*, Eyal and Peltz, 1994) to the present (e.g. Krakatau – *Indonesia*, Simkin and Fiske, 1983). Moreover, these structures may have diverse dimensions, from few (e.g. On-Take – *Japan*, Newhall and Dzurisin, 1988 and references therein) to thousands square kilometres (e.g. Kilgore – *U.S.A.*, Morgan et al., 1984). Furthermore, field studies have shown that calderas have a wide range morphologies and compositions. Despite the difficultness of classifying them into well-defined types, some authors (e.g. Williams, 1941; Williams and McBirney, 1979; Lipman; 1997; Cole et al., 2004) presented possible proposals of classification based on the caldera morphology, composition of the extruded deposits and style of the caldera-forming eruption (see section II.3.4). Moreover, calderas may be found in almost all tectonic regimes, from subduction zones (e.g. Cerro Galán – *Argentina*, Francis et al., 1989) to areas of rifting (e.g. Suswa – *Africa*, Newhall and Dzurisin, 1988 and references therein) (see section II.5.6). Additionally, field studies have provided information about the main structural parts of a caldera, summarised and defined by Lipman (1997). In some cases, analysed calderas indicate episodes of tumescence, prior and after the collapse event. Some authors considered this a general feature of a whole caldera cycle (Smith and Bailey, 1968), although a great number of calderas do not present these types of deformations. Furthermore, volcanic activity occurring after a caldera collapse event is diverse, although there exist general trends (Walker, 1984). The distribution and type of post-caldera activity provide information about structural aspects and about the state of the magma chamber (e.g. Acocella et al., 2002, 2004) (see section II.3.6).

Additionally to field studies results, analogue models have reproduced structural features observed on the field (Branney, 1995). Furthermore, numerical results have helped to understand some aspects of the origin and structure of caldera collapses (e.g. Burov and Guillou-Frottier, 1999; Folch and Martí, 2004, Gray and Monaghan, 2004).

However, some important aspects on caldera dynamics and structure remain poorly understood yet. One of the most significant future tasks for volcanologist is to distinguish precursors for potentially devastating caldera-forming eruptions. Any cluster of active volcanoes should be evaluated as a possible site of a growing upper crustal magma chamber that could lead a future caldera eruption (Lipman, 2000). Since calderas are uncommon structures, the current understanding about the variety of potential precursors is scarce. No enough information exists about the proportion of precursor events that lead to major eruptions and particularly the characteristic time intervals between precursors and eruption. The main problem is that the most important factors controlling if a caldera collapse will occur during a certain eruption are still unknown (at least partially).

I.2 OBJECTIVES

The objectives of this work are diverse. The first aim is to compile existing information about collapse calderas from to field studies, analogue models and mathematical models. It is important to define what has been done so far and to clarify at the same time, the main uncertainties around caldera collapse processes. These two points are crucial to establish the next objectives in the study of these processes in order to advance in the research and not to repeat existing information and works.

Once reached this point, the second aim of this work is to advance in each discipline, both reinterpreting and complementing existing data (fieldwork) or performing new experiments (analogue modelling) or models (mathematical modelling).

Finally, we are going to compare the different results obtained by the three distinct disciplines, in order to establish the restrictions of the applied methods and to draw innovative conclusion about the dynamics and structural evolution of collapse calderas. Finally, we propose a genetic classification of collapse calderas based on all the information compiled and obtained in this work. Additionally, thanks to the information coming from field studies, we are going to create a database with well-studied collapse calderas. This database will be available for the scientific community in order to improve communication and data exchange between scientific groups.

Occasionally, the lack of communication between scientists affects the advance of knowledge, therefore, time to time, it is necessary to reorganize and recapitulate existing information in order to redefine the point of departure for new investigations and studies. Therefore, comparisons and analyses in this work try to be objective and critical taking into account all possibilities and existing interpretations without ruling out any theory or statement.

I.3 METHODOLOGY

In order to achieve the objectives of this work was necessary a multidisciplinary approach, considering:

- **Field studies**
- **Analogue modelling**
- **Theoretical/mathematical modelling**

❖ FIELD STUDIES

A detailed revision of existing fieldwork studies on collapse calderas which provides information about, the caldera structure, the main characteristics of the deposits associated with the caldera-forming event and general trends of pre-, syn- and postcaldera activity. In some cases, it is also possible to determine pre-eruptive conditions of magma and the most probable causes that triggered the eruption. Moreover, the reconstruction of past caldera collapses is a fundamental tool to understand the collapse mechanism of calderas occurring nowadays.

Furthermore, field studies are useful to adjust analogue and mathematical models, to compare the results obtained with the different models and to check their reliability and veracity.

❖ **ANALOGUE MODELS**

Experimental research is essential to understand caldera collapse processes, then without models we are limited to qualitative interpretations of field observations and remote sensing (Martí and Folch, 2005). In short, analogue and scale experiments are a tool to identify, investigate and visualize processes that cannot be directly observed in nature, and in some cases, they help us to establish the guidelines to be followed when applying computational models.

In general, analogue models indicate us how the process of caldera collapse takes place but are not able to quantify it and indicate (from the view of rock mechanics and fluid dynamics) when the collapse will take place and why.

❖ **MATHEMATICAL MODELS**

Mathematical models contrary to analogue models are useful to quantify variables. They are also an important tool for predicting semi-quantitative general conditions for fracture and fault formation. Furthermore, in contrast to analogue models, numerical approximations are able to reproduce and take into account the physical properties of host rock and magma.

Mathematical models are adequate to perform parametric studies but an inappropriate knowledge of the system (i.e. rheology, geometry, boundary conditions) may lead erroneous conclusions. Furthermore, whereas analogue models are able to reproduce permanent deformations (fractures and faults), in the case of mathematical models to simulate fracture/fault propagation is a serious numerical challenge and is still in development.

In general, theoretical models indicate when a caldera collapse will take place but not how.

I.5 STRUCTURE

This work is structured in seven different chapters or parts:

1. **INTRODUCTION:** The final goal of this part is to introduce volcanoes and collapse calderas and the importance of their study. Objectives, methodology and structure of the presented work are also included.
2. **FIELD STUDIES:** This chapter is focused on compiling existing field studies on collapse calderas, elaboration of a calderas database and the reinterpretation of field data. Definition of the existing caldera collapse end-members according to the analysed field observations and evidences.
3. **ANALOGUE MODELS:** Objectives of this chapter are the corresponding state of the art on analogue modelling applied to collapse caldera, the performance of new experiments with innovative results and the comparison of different experimental set-ups.
4. **MATHEMATICAL MODELS:** The main objectives of this chapter are the corresponding state of the art on mathematical modelling applied to collapse caldera, the application of new models with innovative results and the comparison of different mathematical models.
5. **DISCUSSION:** The principal aim of this part is to draw information of caldera collapse mechanism from the comparison of the results obtained in the chapters: “FIELD STUDIES”, “ANALOGUE MODELS” and “MATHEMATICAL MODELS”. Furthermore, the final aim is a proposal of genetic classification of collapse calderas.
6. **SUMMARY AND CONCLUSIONS:** Summary of the obtained results and conclusions
7. **REFERENCES:** List of those references cited in the text including those consulted for the preparation of the collapse caldera database.