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# Igneous Rock Associations of Canada 2. Stages in the Temporal Evolution of Calderas

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Article abstract

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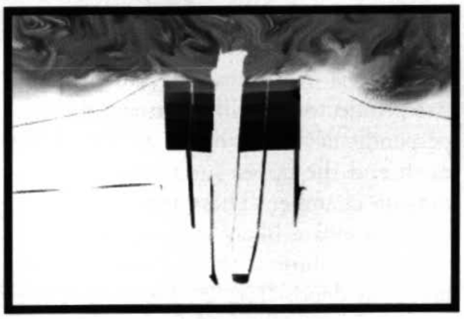
# SERIES



# Igneous Rock Associations of Canada 2. Stages in the Temporal Evolution of Calderas

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## SUMMARY

This paper combines the temporal model of caldera formation presented by Robert Smith and Roy Bailey in 1968 with recent volcanological concepts. Field examples, experimental models and theoretical studies are synthesized to illustrate the process of caldera collapse conceptually as a series of stages of eruption and deformation. During each stage, physical changes occur at the surface, within the underlying magma chamber, and within the subsiding block or blocks that lie between the surface and the top of the magma chamber. The stages are as follows: 1) magma chamber intrusion, 2) initial eruption, downsagging and the onset of subsidence, 3) main subsidence and eruption phase, 4) peripheral extension and eruption quiescence, 5) continued eruption, subsidence and change of eruptive style, and 6) resurgence and extrusion of lava domes and flows. These stages may then be repeated as a subsequent caldera cycle. Every caldera has an individual history and may deform in a different manner at each stage. The paper outlines how these stages can give rise to different caldera types.

## SOMMAIRE

Le présent article fait état des concepts et des publications les plus importantes en matière d'effondrement des caldeiras. On y présente également une vue d'ensemble des différents types de caldeira, leurs caractéristiques illustrées d'exemples concrets, ainsi qu'un glossaire de la nomenclature afférente. Nous traiterons des concepts de piston, d'affaissement, d'effondrement concentrique par paliers, ainsi que de caldeiras de style chaotique, de fossé tectonique et fragmentaire. Il semblent que certaines caldeiras soient le résultat d'une combinaison de style. Nous considérons les interactions complexes de variables qui déterminent la structure et la morphologie des caldeiras et en conditionnent le style.

## INTRODUCTION

Calderas are subsidence structures formed by the evacuation of magma from a subsurface chamber. In a companion paper, we have illustrated the different morphological types of calderas and outlined some of the most influential papers on calderas (Kennedy and Stix, 2003). This second paper adopts a largely conceptual approach to the temporal development of calderas and focuses upon large (8–80 km diameter) silicic systems. The explosive eruptions associated with these calderas are among the largest in the history of the earth and the magmatic systems associated with them may remain active for up to several million years. Calderas are interesting to a wide range of geoscientists. Petrological studies of the eruptive products from calderas illustrate chemical changes involved in the evolution of large, silicic, magmatic systems (Wolff et al., 1990; Hawkesworth et al., 2000). The long-lived magmatic system and the faults associated with caldera formation provide an ideal environment for epithermal and mesothermal mineralization. Calderas also pose an important challenge to volcanologists in terms of hazard assessment, as the threat of further eruption remains long after the caldera has formed.

During their lifetimes, calderas evolve through a series of stages. A variety of caldera morphologies can result from the same basic stages of caldera evolution. The deformation that occurs at each of these stages may be different as a result of different initial conditions such as chamber size and depth, volatile concentrations, and pre-existing structures. These deformational differences become exaggerated throughout the caldera cycle, since the deformation at each stage is partly dependent on the previous stage. We outline the different deformational situations that may exist at each stage, and we present supporting evidence for these stages from experimental and field data. We also attempt to explain these different stages using the distribution and orientation of the principal stress trajectories in the crust above the magma chamber. Our approach builds upon previous models by incorporating results from recent experimental and mathematical studies in which the temporal evolution of calderas has been simulated (Druitt and Sparks, 1984; Gudmundsson, 1988, 1998; Roche et al., 2000; Marti et al., 2000). This study also benefits from new field interpretations and the discovery of different caldera types (Ventura, 1994; Branney, 1995; Hallinan and Brown, 1995; Moore and Kokelaar, 1998). Such studies have revealed that the internal structure of calderas is rarely the simple piston envisaged by Smith and Bailey (1968).

Calderas form from a series of eruptive and deformational events. The collapse of a caldera may take only hours or days (Self and Rampino, 1981; Wolfe and Hoblitt, 1996; Wilson and Hildreth, 1997), yet the deformation preceding and following collapse may last hundreds of thousands of years. A caldera may even undergo several caldera-forming cycles (Smith and Bailey, 1968). We envisage the formation of large silicic calderas to be the result of the following six stages: 1) Magma intrudes and ponds to form a chamber at a shallow level (<10 km) causing some deformation. This chamber begins to increase its pressure by crystallizing and exsolving gas (Tait et al., 1989). 2) The pressure from gas bubbles in the crystallizing magma and the force supplied by the buoyancy of the magma is eventually sufficient to fracture the chamber roof and initiate a plinian style eruption (Druitt and Sparks, 1984; McCleod, 1999). As the eruption proceeds, pressure in the magma chamber drops sufficiently to cause downsagging and subsidence of the chamber roof. 3) The caldera block or blocks subsides along these faults into the magma chamber, increasing its pressure and driving further eruptions. 4) This subsidence causes peripheral extension to occur in the area surrounding the subsiding block(s). 5) The eruption may pause if the conduits close or if the magma becomes depleted in volatiles. If this occurs, the chamber pressure may build up again to a level sufficient to drive further eruption and collapse events. Partly degassed magma then may be erupted as lava domes or lava flows. 6) A period of little surface activity may occur, which can represent the end of the life of the caldera. However, the system may be rejuvenated if magma reenters the shallow chamber to cause resurgence and shallow-level intrusion, uplift and lava dome or flow extrusion. The entire cycle then may repeat itself.

In this paper we discuss the series of evolutionary stages that can be recognized at most calderas, and explain the variability of the different caldera types in terms of the physical processes occurring at and beneath the surface.

## STAGES OF CALDERA DEVELOPMENT

### Stage 1. Magma Chamber Intrusion

Little is known about the early stages of caldera formation, and it is often neglected in the literature. This is because evidence for pre-collapse deformation is rarely well preserved, and structural details are overprinted by later events. The emplacement of large, shallow (2–10 km depth), tabular magma chambers precedes caldera formation (Gudmundsson, 1989; 1998; Gudmundsson et al., 1997; Lipman, 1997; 2000). The intrusion of such chambers requires space. Space can be created in the upper crust by lateral tectonic movements or by vertical movements above or below the chamber (Hutton, 1988). If lateral extension or subsidence beneath the chamber is not sufficient, structural doming of the overlying crust may occur (Fig. 1a-d). Precursory structural domes are found at several calderas and vary from tens to hundreds of kilometres in diameter. The domes can be much larger than the subsequent caldera.

Before eruption, the magma chamber is overpressured, and principal stress trajectories will be oriented perpendicular both to the surface of the earth and the upper surface of the magma chamber. These principal stress trajectories are likely to promote tensional failure at the surface and shear failure at depth (Fig. 2). Bending at the surface forms tensional fractures by extrados-type extension, which develops from the length increase that occurs on the outside of a fold (Fig. 2). At depth, the elevated lithostatic pressure prevents tensional fractures from penetrating beyond a kilometre or so below the

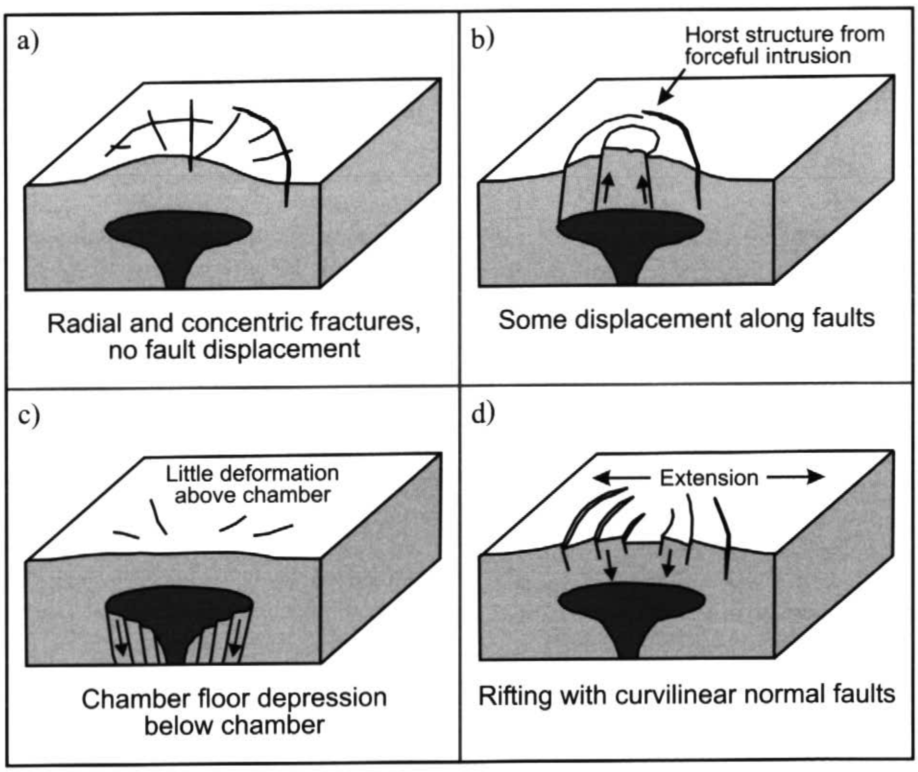


**Figure 1** Possible scenarios after magma chamber intrusion but prior to eruption. a) Doming with radial and concentric fractures. b) Block faulting uplift. c) Very little surface expression as a result of subsidence beneath the chamber (Cruden, 1998). d) Rifting.

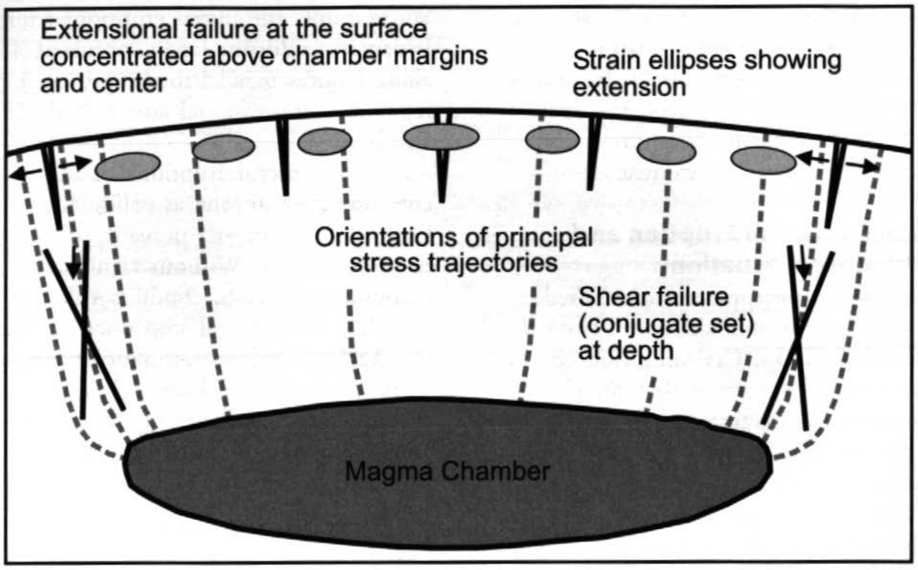


**Figure 2** Principal stress trajectories at a shallow, tabular magma chamber that has been recently intruded and overpressured. These trajectories run orthogonal to the free surfaces, which are represented by the margin of the magma chamber and the surface. Faults will form where the principal stress trajectories are closest together above the margins of the magma chamber (Gudmundsson et al., 1997). These trajectories cause tension at the surface and form a conjugate shear set that can produce either inward- or outward-dipping faults at depth. The strain ellipses illustrate extrados-type extension, as the bent crust is stretched.

surface, instead promoting shear failure and normal fault formation (Gudmundsson 1998).

Doming is caused by thermal expansion, upward magmatic pressure, magmatic buoyancy and gas pressure. The crust overlying the chamber deforms by 1) expansion, 2) elastic plate bending (Withjack and Scheiner, 1982), or 3) fault-controlled uplift. Both upper crustal thermal expansion and elastic plate bending may cause the formation of radial and concentric tension fractures (Fig. 1a). Fault-controlled uplift may show both normal and reverse faults related to the same deformation event (Acocella et al., 2001) (Fig. 1b). Thermal expansion associated with a single chamber may be low as a result of the size of the chamber and the low heat conductivity of the crust. However, the magma chamber may represent only a small thermal portion of a larger magmatic system at depth, which may result in a broad regional dome due to thermal expansion. Doming by elastic plate bending is limited by the small, recoverable, elastic response of the upper crust. Fault-controlled deformation can produce large amounts of uplift. A combination of these driving forces and deformation styles will be present during the development and growth of a large silicic magma chamber.

Experimental studies can simulate doming by increasing the volume of the experimental chamber. Such chamber growth produces both radial tensional fractures and concentric faults (Walter and Troll, 2001) that may be either normal or reverse in nature (Marti et al., 1994; Acocella et al., 2001; Walter and Troll, 2001). The greater the uplift, the more likely normal faults will form. Experiments have shown that doming can produce some subsidence at this stage as a result of the development of polygonal, concentric normal faults (Komuro et al., 1984; Komuro, 1987; Walter and Troll, 2001), but examples of subsidence caused by precursory doming are rare.

Evidence for doming can be found in the sedimentary record; unconformities may reveal details of uplift and erosion, and sedimentation rates can be used to constrain the amount of uplift. The migration of surface drainage patterns also can be useful to identify paleodomes (Pierce and Morgan, 1992).

The "Green Tuff beds" along the inner Japan arc reveal the presence of a 80 km-diameter structural dome which preceded a 5 km-diameter collapse (Komuro et al., 1984). At Yellowstone caldera, Wyoming, there is a crescent of high terrain 350 km across which is raised 500 m above the surrounding landscape (Pierce and Morgan, 1992). This type of regional doming may be due largely to thermal effects.

At Grizzly Peak caldera, Colorado, emplacement of cone sheets and radial dikes, as well as faulting, preceded collapse, the faults being later reactivated (Fridrich et al., 1991). At Kakeya caldera, southwest Japan, collapse was preceded by more than 350 m of uplift by block faulting (Sawada, 1984). Precursory uplift of this nature may be related to forceful intrusion of magma. Many examples of radial and concentric dikes can be observed at calderas; however, the dikes exposed at the surface are usually post-collapse features, and older pre-collapse structures are buried. However, radial dikes are seen at Las Cañadas caldera on Tenerife, and radial fissures also are observed on the flanks of calderas on Fernandina (Fig. 3) and Isabella in the

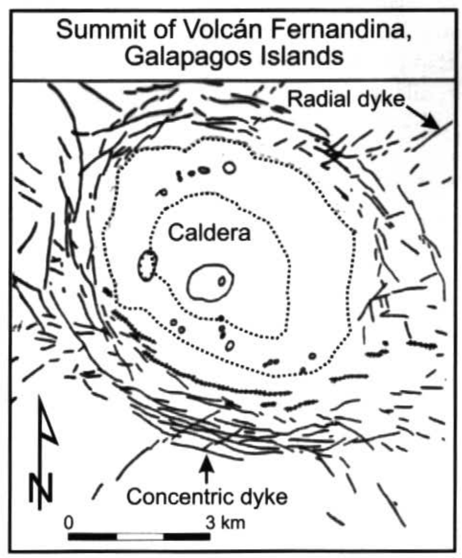


**Figure 3** Radial and concentric dikes around the caldera of Volcán Fernandina, Galapagos Islands (from Chadwick and Howard, 1991, Fig. 4; copyright Springer-Verlag, reprinted by permission of Springer-Verlag).

Galápagos Islands (Chadwick and Dieterich, 1995). Cone sheets are beautifully preserved in Gran Canaria, Canary Islands (Schirnick et al., 1999). Cone sheets also are common features at eroded igneous ring complexes, as observed in many of the Tertiary intrusions of northwest Scotland. The radial fissures and the cone sheets of Galápagos volcanoes result from elevated pressure within a flat-topped magma chamber but do not necessarily require large amounts of doming (Anderson, 1936; Chadwick and Dieterich, 1995). Interestingly, structural evidence for precursory doming is frequently lacking at calderas (Lipman, 1997), implying either it only sometimes occurs or that evidence is lost because of later deformation. By contrast with uplift, space also can be created by depression of the magma chamber floor, which allows intrusion without much surface deformation (Fig. 1c). Subsidence of the floor of the magma chamber occurs as magma intrudes. This subsidence may be either a brittle or ductile process but will produce little surface deformation (Cruden, 1998). Chamber floor depression is one explanation for the lack of doming at many calderas.

If the regional stress field is extensional, some space for intrusion may be provided by rifting (Fig. 1d). Rapid extension during intrusion may result in the formation of a graben. Toba caldera in Indonesia and Snowdon caldera in Wales both show rift formation prior to caldera collapse (Chesner and Rose, 1991; Kokelaar, 1992). Calderas commonly form in transtensional environments, with pull-apart basins forming at an angle oblique to master strike-slip faults. The morphology of the main caldera on Vulcano Island, Italy, resembles a pull-apart basin oriented in this fashion relative to strike slip faults (Ventura, 1994).

The structures produced before caldera collapse may be critical to later stages. Smith and Bailey (1968) proposed that the ring fault used for collapse was created during a stage of precursory doming. There is little evidence that ring faults are created during doming, but some structures created during the intrusion stage may be re-activated during collapse or resurgence. The stress field that exists during chamber emplacement also affects the subsequent eruption and collapse history of a caldera.

## Stage 2. Initial Eruption and Chamber Evacuation

Maximum overpressure is reached within the magma chamber before the initial eruption. Crystallization causes gas vesicles to form within the surrounding magma and to exert a gas pressure (Tait et al., 1989). The buoyancy of a vesicular magma also exerts pressure on the chamber roof (McCleod, 1999). Intrusion of magma into the chamber will contribute additional pressure and may serve to trigger the initial eruption. If fractures or faults produced during stage 1 penetrate from the surface down to the magma chamber, they can be used for the initial eruption (Fig. 4). If overpressure exceeds the tensile strength of the rock, new fractures can be created which will propagate upward from the magma chamber (Fig. 4a). Arching of the chamber roof and thermal expansion of the roof rocks are likely to modify the stress field around the chamber and aid the tensile failure of the roof rocks (Fig. 2).

To propagate a dike from the chamber roof to the surface requires additional overpressure (McLeod and Tait, 1999). The nature of the initial fractures and dikes is controlled by the orientation and concentration of the principal stress trajectories (Fig. 2). A single vent is formed when the magmatic pressure overcomes the strength of the rock. Caldera-forming eruptions often begin with a single vent phase (Bacon, 1983; Wohletz et al., 1995; Druitt et al., 1999).

The overpressure within the chamber is likely to exploit preferentially oriented weaknesses or fractures in the overlying rock which represent the area of minimum tensile strength. This allows a dike-like conduit to be established from the magma chamber roof to the surface (Fig. 4b). If rifting occurs during intrusion of the magma chamber, these faults can be used for the initial eruption (Fig. 4c). For example, the initial eruptions from Ishizuchi cauldron, Japan, occurred along fissures parallel to the principal regional compressional stress (Yoshida, 1984).

The initial eruption is likely to continue only so long as conduits are kept open by magma pressure or conduit erosion. Without significant amounts of erosion, conduits will close, and the eruption will stop once pressure is reduced to below lithostatic at the top of the chamber (McLeod, 1999). Plinian eruptions are thought to occur when the chamber is overpressured.

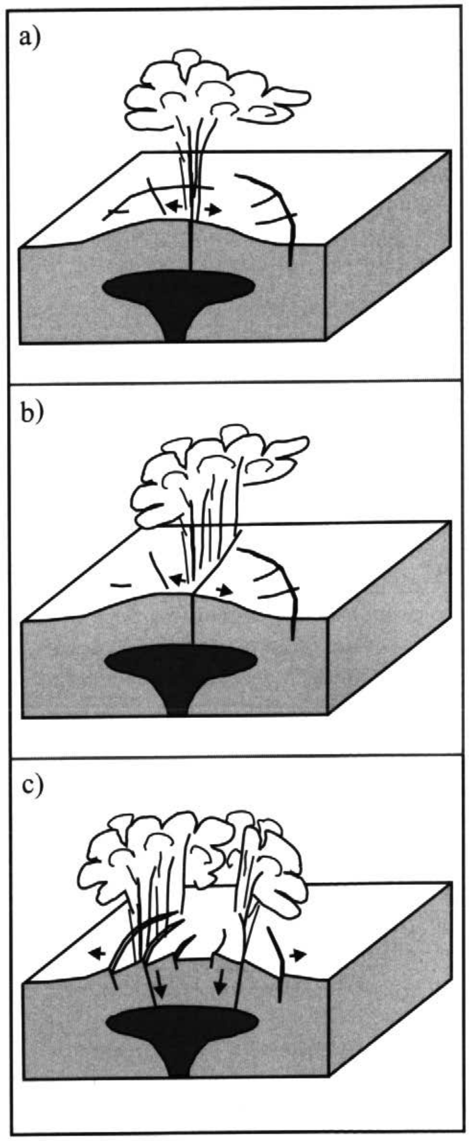


**Figure 4** Possible scenarios for the initial plinian stage of a caldera-forming eruption. a) Central vent eruption; b) regional fissure eruption; and c) rift-related eruption.

These volumes are relatively small compared to the volumes of pyroclastic flows erupted during caldera collapse (Druitt and Sparks, 1984; Gudmundsson, 1998). For example, the plinian fallout unit from the Lower Bandelier Tuff has a volume of 20 $km^3$ dense rock equivalent (DRE) compared to 400 $km^3$ DRE of pyroclastic flows. The Upper Bandelier Tuff consists of 15 $km^3$ DRE of plinian fall and 250 $km^3$ DRE of pyroclastic flows (Self and Lipman, 1989).

This initial plinian stage may not occur if magma viscosities and gas contents are low, as in the case of most basaltic calderas, or if vent geometries are not appropriate. If the magma chamber is an open system and allowed to degas, high overpressures may never develop, and the initial plinian stage may not occur.

The change in eruption style from plinian column to pyroclastic flow may occur during this phase. The bleeding of overpressure by the plinian eruption results in a drop in chamber pressure, causing the plinian column to collapse to form pyroclastic flows (Druitt and Sparks, 1984). For example, the Climactic Pumice and the Wineglass Welded Tuff were erupted from a single vent at Mt. Mazama, Oregon, indicating collapse of the plinian column to form pyroclastic flows before Crater Lake caldera formed (Bacon, 1983). If chamber pressure falls below lithostatic pressure, eruptions may even cease prior to subsidence.

During the initial chamber evacuation stage, the gradual depressurization of the chamber also may cause a period of downsagging, unless early-formed faults allow instantaneous failure of the crustal block above the magma chamber. The bending caused by downsagging helps to propagate fractures upward from the chamber margins. This early stage of surface downsagging is clearly seen in experimental models (Marti et al., 1994; Roche et al., 2000; Kennedy, 2000), and an initial period of downsagging-dominated subsidence has been documented at Grizzly Peak caldera before ring faults became active (Fridrich et al., 1991). Many calderas show inward-tilted beds, especially during early stages, which are a direct result of downsagging (Walker, 1984). If subsidence-controlling faults never fully develop, a downsag-style caldera is formed. Taupo caldera in New Zealand has been interpreted as such a caldera (Walker, 1984). Experimental models show that a ring fault will develop on one side first, producing initially asymmetrical subsidence (Burov and Guillou-Frottier, 1999; Roche et al., 2000; Kennedy, 2000). This initial asymmetry will be enhanced by heterogeneities in the magma chamber geometry or crustal strength. If the initial conduit is near vertical and located close to the chamber margin, it may be used for subsidence at this stage. This is observed at Ishizuchi cauldron, where sections of the initial eruptive fissures were interpreted to control early subsidence (Yoshida, 1984). By contrast, the presence of a large mass of pre-existing topography may promote collapse with less downsagging (Lavallée et al., in press). Ring structures from a previous collapse or from large amounts of tumescence may allow early piston-like collapse. Pre-existing normal faults related to extension also can facilitate early collapse. If well-developed regional faults suitable for subsidence exist, they also can be used at this stage without the need for downsagging. Use of such faults can be seen during the early stages of Glencoe caldera, Scotland, which is a tectonically controlled piecemeal caldera (Moore and Kokelaar, 1998).

If downsagging does occur, it may result in the rotation of pre-existing regional faults and those formed by tumescence. The sagging at the surface and at the chamber roof also will affect the orientation of the principal stress trajectories (Fig. 5a). Rotation will cause inward-dipping fractures to be rotated towards the vertical, which is a better orientation for subsidence (Fig. 5a). Depending on the degree of rotation caused by sagging and the original orientation of the fractures, certain structures produced by tumescence theoretically could be re-used during the various stages of subsidence. The orientation of the initial subsidence-controlling fault will depend on the presence or absence of a superimposed regional stress regime. In the absence of such a stress regime, the first subsidence-related fault is likely to be outward dipping (Branney, 1995; Roche et al., 2000). This fault will allow continued eruption, since the geometry of the subsiding block will keep this fault or conduit open, even when the pressure in the magma chamber drops below lithostatic (Fig. 5b).

### Stage 3. Main Subsidence and Eruption Phase

Faults develop during the main subsidence phase, allowing hundreds to thousands of metres of displacement. Subsidence upon a fault on one side of the caldera will significantly modify the stress field associated with the chamber. Movement on an outward-dipping fault will increase the bending on the opposite side of the caldera, promoting tensional fractures and normal faulting there (Branney, 1995; Odonne et al., 1999; Roche et al., 2000) (Fig. 6). Initial subsidence upon an inward-dipping fault, without sufficient extension, may result in an outward-dipping fault forming on the opposite side of the caldera, as seen at Ishizuchi cauldron (Yoshida, 1984).

As subsidence continues, the initial fault will propagate laterally from the side of maximum subsidence (Roche et al., 2000; Kennedy, 2000) (Fig. 6a). This fault may never propagate fully to the other side, forming a trapdoor-style caldera, as seen at Silverton and Kumano calderas (Miura, 1999; Lipman, 2000).

Arcuate faults may eventually join up or propagate to produce a ring fault, such as at Grizzly Peak and Ishizuchi (Yoshida, 1984; Lipman, 2000). The orientation of the faults and the way in which they propagate and join will depend mainly on the following aspects: 1) the external stress field can control the dip of the subsidence-controlling faults, as observed at Vulcano Island (Ventura, 1994); 2) the diameter of the subsiding block and its thickness from the surface to the top of the magma chamber (the aspect ratio) are critical in determining the type of caldera that forms (Roche et al, 2000; Kennedy, 2000); and 3) the three-dimensional symmetry of the magma

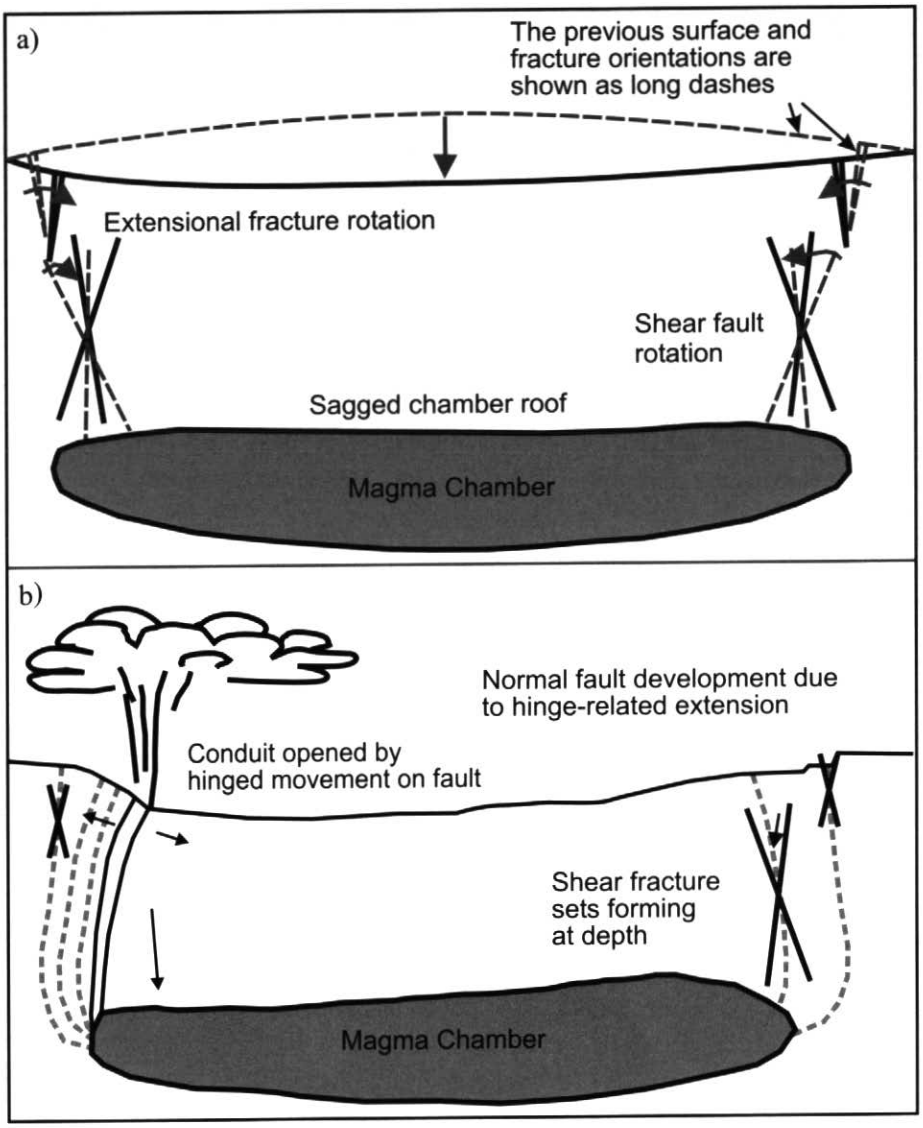


**Figure 5** a) A schematic view of downsagging. New shear faults begin to form at depth, and sagging at the surface causes extension at the hinges of the depression. The sagging also rotates inward-dipping faults towards a vertical position. b) Initially asymmetrical subsidence along an outward-dipping fault, maintaining an open conduit.

chamber may control the symmetry of collapse (Burov and Guillou-Frottier, 1999; Lipman, 2000; Roche et al., 2000; Kennedy, 2000). For example, a magma chamber with an irregular upper surface could result in a complex piecemeal-style caldera. Other factors that are likely to affect the development of caldera faults include pre-existing faults, folds, metamorphic grain and topography; lateral and vertical variations in the tensile strength of the rocks that comprise the subsiding block, such as hydrothermally weakened and fractured areas, which are common in volcanic terrains; and the position of the erupted ignimbrite sheets within the caldera, which add load to the subsiding block. These elements will reduce the likelihood of simple piston collapse, and their interactions will result in a complex mix of caldera types. Indeed, caldera faults may develop as a series of linear faults that connect to form distinct corners, forming a polygonal subsiding block (Kennedy, 2000) rather than a continuous ring as in the piston example.

Collapse occurs when the upward pressure on the chamber roof is unable to support the lithostatic load. At this stage, lithostatic pressure exceeds the sum of the shear strength of the rock, the pressure from magma buoyancy, and the gas pressure. The subsiding block collapses into the magma chamber, in the process potentially repressurizing the magma. If this occurs, the repressurization will increase the gas pressure of the magma, decrease the buoyancy and slow collapse. However, since the lithostatic pressure from the subsiding block continues to exceed the sum of the gas pressure and the pressure from buoyancy, subsidence along vertical or outward dipping faults will continue, limited only by the eruption rate. Gas pressure gradually will be reduced as vesicles redissolve and equilibrate with magmastatic pressure. Collapse will stop or pause when eruption conduits close or become choked, or when fault geometries are no longer suitable for collapse.

### Stage 4. Peripheral Extension and Eruption Quiescence

As collapse proceeds along the subsidence-controlling faults, peripheral extension develops, and outer inward-dipping faults are formed (Walker, 1984; Branney, 1995; Roche et al., 2000; Kennedy, 2000). These faults are thus a response to the subsidence-controlling faults found in more central regions of the caldera. Fresh fault scarps are unsupported and hence under tension; as a result, normal faults form, accompanied by slumping and sliding of caldera walls, which can form megabreccias and mesobreccias within the caldera (Branney, 1995; Lipman, 1997; Roche et al., 2000). The same forces that are responsible for these slide blocks also will produce peripheral normal faulting in the region outside the main subsidence area (Fig. 7). Arcuate crevasses related to slide blocks are commonly observed (Branney, 1995). Peripheral graben and normal faulting also can form in areas where the angle of downsagging is high, without major displacement upon a subsidence-controlling fault. The faults may form preferentially along tensional fractures that were created by bending during either the tumescence or downsagging stages. Peripheral extension structures are well exposed at Scafell caldera,

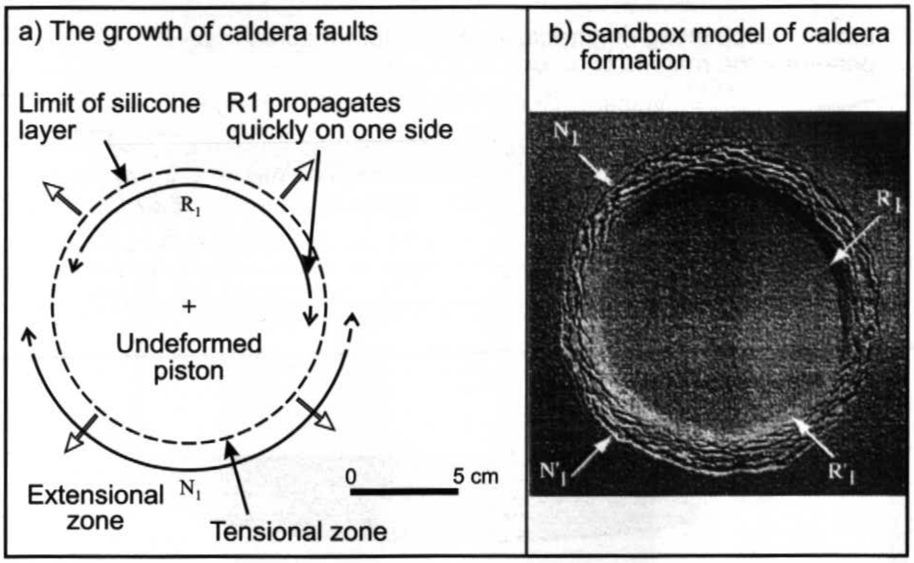


**Figure 6** a) Fault development and propagation according to stress trajectories. b) Early fault development as illustrated by scaled experimental studies (from Roche, et al., 2000).

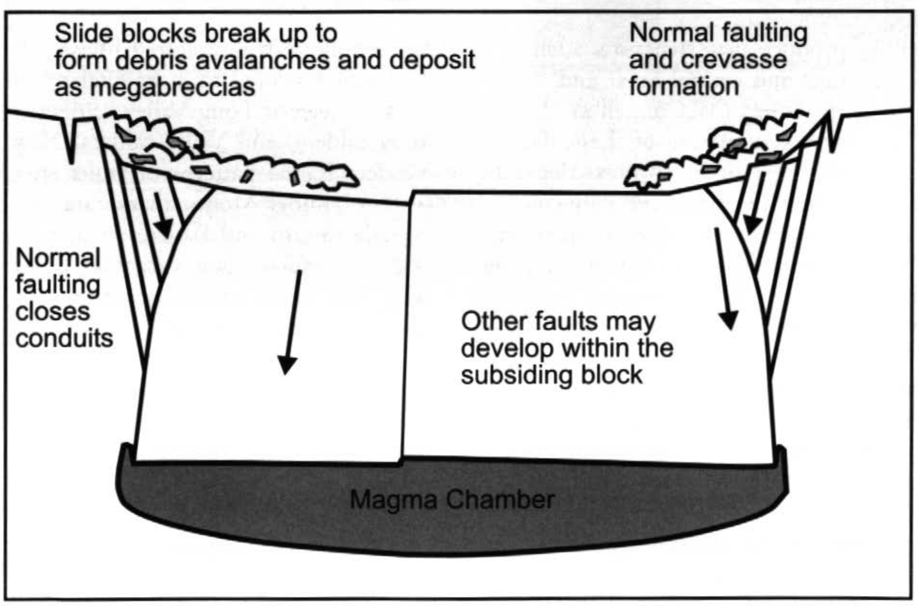


**Figure 7** Late-stage peripheral downsagging, continued normal faulting, slide block, debris avalanche, megabreccia and crevasse formation.

England, and Snowdon (Branney, 1995; Kokelaar and Branney, 1999; Kokelaar, 1992). At Suswa volcano, Gregory Rift, Kenya, both inner and outer ring fractures are seen (Skilling, 1993), with collapse occurring on the inner ring fault and eruptions on the outer ring fractures. Interestingly, both the outer and inner ring faults at Suswa are inward dipping (Skilling, 1993).

A coupled relationship may exist between the inner, subsidence-controlling faults and the outer extensional faults. Movement upon the subsidence-controlling fault will significantly alter the stress trajectories near the surface, while movement on the outer faults and slumps from these fault scarps may close or block any conduits that are still open on the inner faults (Fig. 7); this process may result in eruptions migrating to the outer faults. Alternatively, a pressure buildup and lull in eruption may occur. If the chamber becomes sufficiently repressurized to cause conduit reopening, eruptions and subsidence along the main caldera faults can proceed.

### Stage 5. Continued Eruption, Subsidence and Change of Eruptive Style

After collapse, the magma chamber is likely to be near lithostatic pressure. Yet, many calderas show multiple collapse and eruption episodes within a single caldera-forming cycle, as observed for Scafell, Glencoe and Snowdon. This could be due to a post-collapse increase in chamber pressure that drives further eruptions and continued collapse. Possible mechanisms for increasing the chamber pressure to above lithostatic are further crystallization and exsolution of volatiles, renewed input of magma into the magma chamber, magma interaction with water (e.g., a caldera lake), or further collapse.

A suitably oversaturated magma that continues to crystallize will produce overpressured vesicles, reopen conduits and drive further eruptions. Renewed influxes of magma into the chamber and magma mixing also may promote continued eruption. For example, the partial emptying of the magma chamber during the main collapse phase may allow new magma to enter the chamber and interact with the pre-existing magma, driving continued eruptions. Eruptions also may be driven by magma interactions with seawater, groundwater or water in a caldera lake. At Ilopango caldera, El Salvador, many post-collapse eruptions show evidence for interaction with water from the caldera lake (J. Vallance, pers. comm., 2000). Continued removal of magma from the chamber also will promote subsidence of the caldera.

In these cases of renewed activity, it is likely that the subsidence-controlling faults will also be reactivated, and peripheral extension features will continue to form. Eruptions again will cease when the conduits close, and the pressure at the top of the chamber returns to lithostatic. This eruption-collapse scenario may repeat itself several times if gas concentrations in the magma remain sufficiently high after a second pulse of eruption and collapse. We hypothesize that the length of time between these eruptive and subsidence events will become increasingly longer as the magma becomes progressively degassed. As dissolved gas concentrations decrease in the magma, it will become progressively more difficult to diffuse gas from magma into bubbles, since diffusion times increase significantly as magma is dewatered (Watson, 1994).

Eruptive events also may become progressively less explosive as a result of magma degassing, and the eruptive style can shift from explosive to effusive, as observed at many calderas where lava domes and lava flows are extruded soon after the main pyroclastic eruptive phase. At Long Valley caldera, California, large volumes of rhyolite lava flows were erupted within 100,000 years of collapse (Bailey et al., 1976). These vents are very often the same as those used for the pyroclastic eruptions. A similar shift is observed after the most recent caldera collapse at Yellowstone caldera, where large volumes of rhyolite lavas were erupted after ignimbrite eruptions (Christiansen, 1984; 2001). The accumulation of these lava flows may produce a topographic high within the caldera that should not be confused with the resurgent dome (Fig. 8).

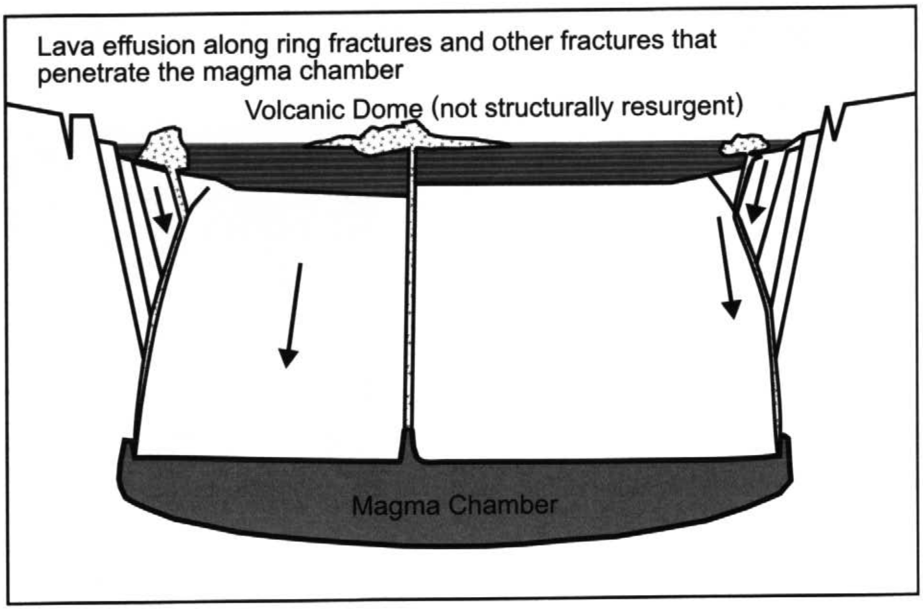


**Figure 8** Pre-resurgence lava effusion along ring fractures and other conduits. The central extrusion of lava does not represent a structural resurgent dome.

## Stage 6. Resurgence and Extrusion of Lava Flows and Domes

Smith and Bailey (1968) used the term resurgence to explain the formation of structural domes after caldera collapse. They proposed that it is an essential component of the caldera-forming process and is caused by a renewed rise of magma beneath the subsiding block. They also proposed that regional detumescence, which is the sinking of the region around the caldera, could also explain these structural domes. Marsh (1984) calculated that the timescale of regional detumescence was relevant to resurgence and could be a significant mechanism to produce such structures.

Resurgent domes have different characteristics; they may form a broad structural dome (Smith and Bailey, 1968), re-activate subsidence-related or regional faults (Acocella and Funiciello, 1999), produce new structures such as radial faults and central horst and graben structures (McConnell et al., 1995) or a combination of these. The nature of the resurgent process depends, in part, upon the structural coherence of the roof above the magma chamber. The collapse style of the caldera controls this coherence. We hypothesize that the style of collapse will influence the style of resurgence. For example, a faulted roof may permit smaller, shallow intrusions to form above the main magma chamber; associated with these intrusions, faulting, structural uplift and extension will occur. Frequently resurgence is not a simple re-inflation of the magma chamber, but instead a complex upward-stoping and intrusion process into and above the magma chamber. At the Okueyama volcano-plutonic center, southwest Japan, stoping is interpreted as an important process during resurgence, allowing extrusive rocks to be intruded by the resurgent magma, with no evidence for lateral pushing (Takahashi, 1986). At Long Valley caldera, the intrusion of 300 m of sills above the main magma chamber in early post-caldera time has been theorized to account for the resurgent uplift (McConnell et al., 1995).

New sets of extensional faults appear to form at the surface, commonly with a preferred linear orientation related to a regional stress field, as seen at Long Valley caldera, Toba caldera, and Valles caldera, New Mexico. Radial patterns of faults are seen at Timber Mountain caldera, Nevada (Smith and Bailey, 1968) (Fig. 9a), and to some extent at Valles (Fig. 9b). These patterns resemble doming experiments without and with an external extensional stress field (Withjack and Scheiner 1982) (Fig. 9c, d). At Toba caldera, pre-caldera rift structures appear to have been reactivated. Resurgent doming on the island of Ischia, Italy, is clearly related to the regional tectonics of the area where pre-existing normal faults were reactivated during resurgence (Acocella and Funiciello, 1999). Perhaps one reason that subsidence structures are not often reactivated by resurgence is because they have become "sealed" or "stitched" by intrusions such as ring dikes and cone sheets at the end or after the main subsidence phase.

Effusive events also occur after the main period of resurgence and after a period of eruptive quiescence. These effusive events occur in the moat area of the caldera between the resurgent dome and the topographic boundary, as seen at Long Valley and Valles calderas. The distribution of these moat domes has been interpreted to mark the location of a ring fracture and subsidence-controlling faults (Smith and Bailey, 1968). The area of peripheral extension outside and concentric to the subsidence-controlling faults may be a suitable location for dome emplacement if the faults are connected to the chamber at depth (Fig. 10). By contrast, the main subsidence-controlling faults may be in an area of compression. Therefore, the interpretation that post-collapse domes and vents mark the main subsidence-controlling ring fault (Smith and Bailey, 1968) may not always be correct. Instead, these extrusive centers may mark the area peripheral to the main subsidence-controlling fault (Fig. 10). Another preferred location for extrusions is along regional extensional and transtensional fissures (Walker, 1984). Where regional structures intersect the resurgent dome, effusive eruptive vents also may form, as clearly seen at Long Valley caldera. It is perhaps surprising that lavas are rarely extruded along faults formed by the resurgent dome itself; perhaps these do not penetrate sufficiently deep to reach the main chamber.

After the caldera-forming eruption and subsidence events, the roof of the magma chamber lies at a deeper level. The pressure at the base of the chamber is the sum of the pressures due to the magmatic head and the lithostatic head. After the eruption, a certain mass of magma expelled from the chamber will pond within the caldera. However, significant volumes of magma also can be deposited outside the caldera as pyroclastic flows and falls. As a result, the pressure above the base of the chamber may be substantially reduced because of the net loss of mass from the area above the base of the magma chamber (Fig. 11). This reduced pressure at the base of the chamber provides a mechanism by which magma replenishment can occur into the chamber. Therefore, any buoyant magma at deeper structural levels (e.g., mid-crustal regions) can flow into the chamber causing resurgence, dome extrusions or explosive eruptions.

Resurgence occurs more frequently in large calderas than in small calderas (Smith and Bailey, 1968). This

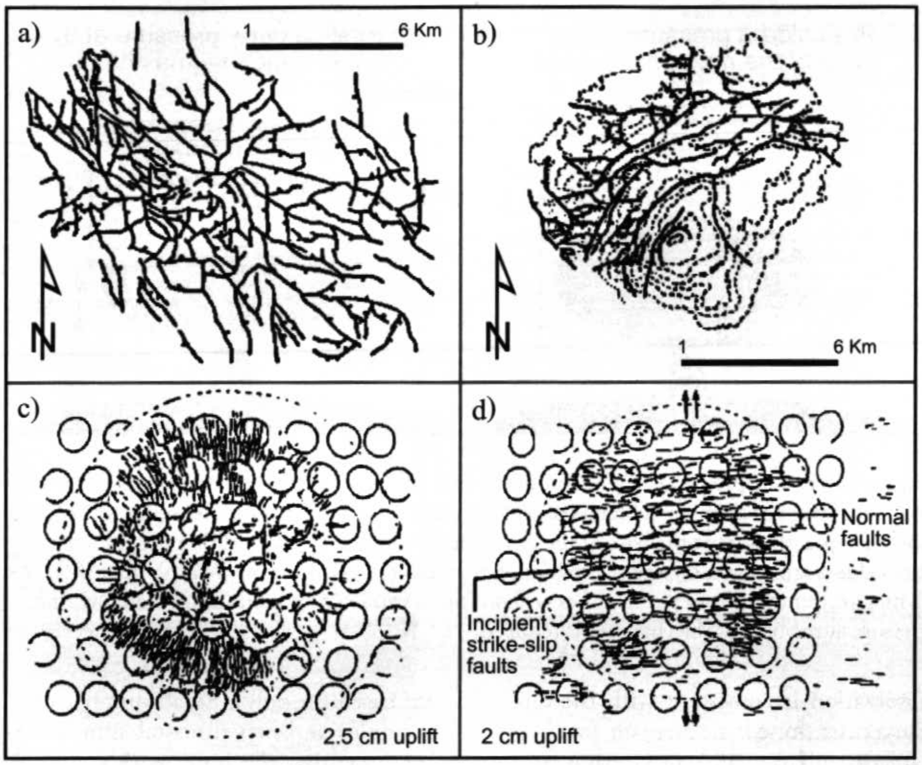


**Figure 9** Comparison between experimental doming and resurgence. a) The Timber Mountain resurgent dome (from Smith and Bailey, 1968; reprinted by permission of the Geological Society of America). b) The Valles resurgent dome (from Smith and Bailey, 1968; reprinted by permission of the Geological Society of America). c) Experimental doming without extension (from Withjack and Scheiner, 1982; reprinted by permission of the AAPG whose permission is required for further use). d) Experimental doming with extension (from Withjack and Scheiner, 1982; reprinted by permission of the AAPG whose permission is required for further use).

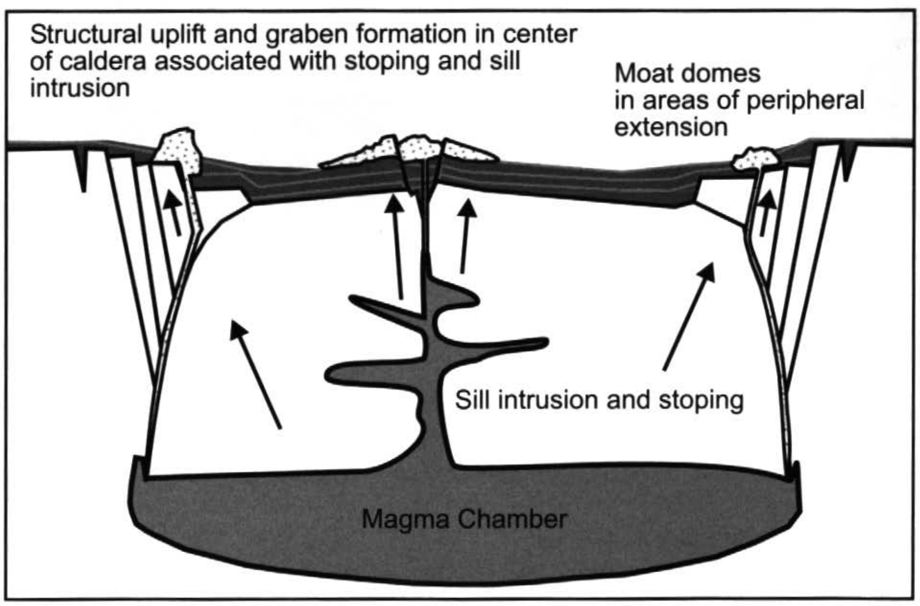


**Figure 10** Structural resurgence of the caldera, forming a broad, central dome. Magma chamber stoping and shallow intrusion may produce localized uplift and graben formation. The ring faults also may be reactivated during resurgence, which may cause peripheral uplift.

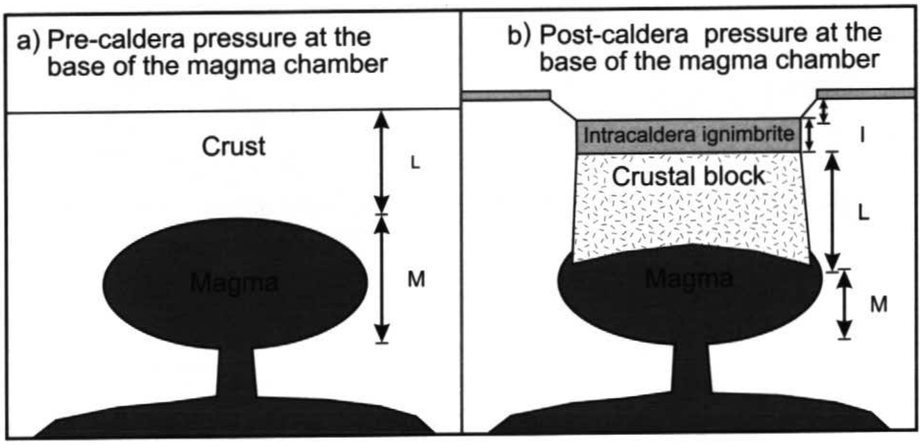


**Figure 11** a) The pre-collapse pressure at the base of the main magma chamber is a function of the thickness of the crustal block (L) and the thickness of the magma chamber (M). b) The post-collapse pressure at the base of the chamber is modified by the addition of ponded intracaldera ignimbrite (I) and the reduction in the thickness of the magma chamber (M). Not all magma removed from the chamber is deposited as intracaldera ignimbrite; therefore, the pressure at the base of the chamber is reduced.

observation is consistent with the above-mentioned mechanism for caldera collapse, since large calderas are associated with large-volume eruptions, producing a correspondingly big reduction in magmastatic pressure at the base of the chamber after the eruption. Such a mechanism even may be responsible for initiating a new caldera cycle.

## CONCLUSION

A caldera is commonly the result of many eruptive and deformation events that occur over a long time interval. For example, Cerro Galan in Argentina was active for 3 Ma (Francis, 1993). Although these events occur as a series of stages, not all calderas experience each of the stages that we have illustrated here. For example, a small magmatic system with a limited supply of gas-rich magma may not exhibit all the stages. However, the longer the lifespan and the size of the caldera, the more likely that all stages will occur. Long-lived caldera systems may cycle several times through the various stages. The different stages should be considered when interpreting caldera structure and eruptive stratigraphy, and caldera formation should not be considered as a single discrete event.

As the scientific community has never directly observed a large caldera-forming eruption, some of these stages are based on only a small number of examples and experimental simulations. We hope that the framework we provide will encourage research into the stages that are poorly understood, such as chamber intrusion, peripheral extension, and resurgence.

Understanding the interplay and sequencing of these stages will improve our insight regarding currently restless calderas. At least 43 separate post-caldera eruptive events have occurred from calderas in the last 100 years alone (Newhall and Dzurisin, 1988). In many cases, it is difficult to know if a caldera is at the end of a cycle and undergoing resurgence and late-stage magmatism, or whether it is at the beginning of a new cycle and preparing for a large eruption. Perhaps the best example of this dilemma is Yellowstone caldera, which last erupted to form a caldera at 0.6 Ma. Based on previous caldera-forming eruptions at 2.0 Ma and 1.6 Ma, the recurrence interval for such eruptions may be between 0.4 and 1 Ma (Christiansen, 2001). Is Yellowstone going to experience another caldera-forming eruption in the geologically near future? Clearly, this is a critical question. A caldera undergoing resurgence represents a comparatively small hazard, with the possibility of dome growth and relatively minor explosions. By contrast, a large, caldera-forming eruption represents one of the greatest natural hazards to mankind, with the possibility of devastating vast areas and affecting Earth's climate for many years.

The subsurface structure of calderas also is important for certain types of economic ore deposits. These structures play a primary role in controlling hydrothermal pathways associated with post-caldera activity. These pathways in turn control the deposition of epithermal and mesothermal mineralization associated with calderas and ring complexes (Lipman and Sawyer, 1985). Volcanogenic massive sulphide deposits are commonly associated with submarine calderas, and the structure of the caldera is likely to control the location of this type of mineralization as well (Stix et al., 2003). Understanding these hydrothermal pathways is also vital to the geothermal power industry, as the circulating fluids control the nature and distribution of the heat flux within a caldera block.

In conclusion, it is our hope that the conceptual framework outlined in this paper provides a basis for better understanding the physical mechanisms and processes that are at work at the surface and in the subsurface during caldera development. This framework represents our particular view of how calderas develop with time, and we hope that this attempt will stimulate other researchers to modify and improve this model.

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