12. See Table 12 in (2).
13. A systematic error of 41 km in the radii of all nine Uranian rings would be required in order for the resonance radius to match the δ ring radius. The estimated uncertainty in the δ ring semimajor axis is about 5 km (2).
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Model for the Intrusion of Batholiths Associated with the Eruption of Large-Volume Ash-Flow Tuffs

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Pyroclastic eruption and the intrusion of batholiths associated with large-volume ash-flow tuffs may be driven by a decrease in reservoir pressure caused by the low density of the magma column due to vesiculation. Batholithic intrusion would then be accomplished by the subsidence and settling of kilometer-sized crustal blocks through the magma chamber, resulting in eventual collapse to form large caldera structures at the surface. Such a model does not require the formation of a large, laterally extensive, shallow magma chamber before the onset of large-volume ash-flow eruptions. Eruption could commence directly from a deeper reservoir, with only a small channelway being opened to the surface before the onset of catastrophic ash-flow eruptions of the scale of Yellowstone or Long Valley. Such a model has wide-ranging implications, and explains many of the problems inherent in the simple collapse model involving shallow magma chambers as well as the process and timing of batholith intrusion in such cases.

IN THIS REPORT WE OUTLINE A MODEL for the emplacement of batholiths and the eruption mechanism for large-volume ash-flow tuffs. Because this model satisfies many of the physical data now available, and since it has dramatic implications for volcanic hazards and hydrothermal power projects, it should be considered as a possible alternative to the shallow magma chamber collapse model currently being proposed for all caldera-forming events.

The emplacement of batholiths has always been a problem in geology because of the need to explain where the original rocks went. Since the magma comes from below, some process must cause crustal subsidence to offset the upward movement of the magma. Large-volume ash flows such as the Bishop (*1*), Yellowstone (*2*), and San Juan Mountains (*3*) tuffs represent 500 to 3000 km^3 of material, implying magma systems of batholithic proportions (a batholith is a minimum of 100 km^2 in outcrop area; most are not more than 5 km thick). Since many batholiths are accompanied by cogenetic ash-flow tuffs immediately before their final emplacement, both share common origins and should be intimately related in intrusive and extrusive history.

A second problem is that a number of recent ash-flow eruptions do not seem to have left behind extensive and continuous shallow magma chambers, as would be suggested by the current model of caldera collapse into a preexisting shallow chamber (*4*). Therefore the question must be asked, where did the magma go?

As silicic magma coalesces in the lower crust, it will begin to rise diapirically when the body becomes large enough (*5*). This type of upward migration depends on the surrounding material being plastic because of high temperatures and confining pressures. As the magma moves upward, it must soften the surrounding rocks by heating before continued movement is possible (*5*). There is increasing geophysical evidence to suggest that this process leads to a ponding of magmas in the mid-crust, where the rocks become cooler and more brittle (*4, 6*). Geochemical evidence also suggests that the magma chemistry of batholiths represents crystal reequilibration at mid-crustal depths (*7*). The present model therefore starts with a large concentration of magma at depths of 10 to 20 km.

The density contrasts that lead to diapiric upwelling still exist, and cause distension of the overlying crust and propagation of outward fractures (Fig. 1A). The process is driven by the pressure differential between the upper part of the magma body and the surrounding rocks. The lower crustal rocks below the body are plastic, having been preheated. They push in on the lower margin with a pressure corresponding to the local geobaric conditions. Since the magma is a fluid, although a rather viscous one, pressure is transmitted to the upper surface where the magma pressure is equal to the lithostatic pressure at the base minus the magma column pressure. This pressure is higher than the local lithostatic pressure at the top by the term

$$P = 98.0655 \int_0^L \Delta\rho dl \qquad (1)$$

where L is the thickness of the body in kilometers and $\Delta\rho$ is the difference between the magma density and that of the country rock as a function of depth. If we assume that the densities do not vary much over the thickness of the body, then Eq. 1 simplifies to

$$P = 98.0655\Delta\rho L \qquad (2)$$

The resulting pressure differential is about 100 bars per 1 g/cm^3 of density differential for every 1 km in thickness. This force will aid crack propagation in the tensional environment, especially if preexisting zones of weakness are utilized, as is generally the case in the localization of plutons.

Crack propagation therefore progresses into the brittle environment, with fingers of magma expanding ring fractures toward the surface (Fig. 1A). Such a process is now occurring under the Mammoth Lakes area in California (*4, 8*).

Eventually, the magma will reach the surface. If the magma channelway is big enough, large volumes of magma may begin to be extruded as pyroclastic material, provided there are sufficient volatiles to drive the eruption. At this point, the fluid pressure in the magma chamber changes. If a significant amount of material is moving up the column, the pressure in the magma chamber will attempt to reach the pressure of the flowing magma column, which is given by the expression

$$P = 98.0655\int_0^H \rho_m dh \qquad (3)$$

where ρ_m is the density of the magma in grams per cubic centimeter and H is the

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depth to the magma chamber. The pressure differential between lithostatic pressure in the overlying rocks and the fluid pressure in the magma will be

$$\Delta P = 98.0655 \int_0^H \Delta\rho dh \qquad (4)$$

where $\Delta\rho$ is the differential in density between the magma and country rocks. If the magma body is 10 to 20 km deep, this pressure is 1 to 2 kbar for a density contrast of 1 g/cm^3.

The density of the magma column is not constant, however. During active pyroclastic eruption, the upper portions are vesiculated as gases are evolved to drive the eruption column. The density in the upper few kilometers may therefore be very low. Figure 2A shows a possible density profile in a vesiculating magma column (*9*). The pressure differentials calculated for various models (Fig. 2B) are in the range 1 to 4 kbar. In most cases the full pressure differential will not be realized since the viscosity is high and friction is significant. However, in a system under full pyroclastic eruption, the flow rates would be high enough that the differential may be approached. A more important limitation is that vesiculation in this dynamic system will not reach equilibrium. However, as Fig. 2 shows, even the effective evolution of 1 percent water by weight is sufficient to generate substantial depressurization of a deep magma chamber.

If nothing else occurred, eruption would cease once this pressure differential was established. However, as a consequence of the reduced magma pressure, the previously fractured roof rocks are pulled into the magma chamber (Fig. 1B). Kilometer-sized blocks may be engulfed, since regional joints and zones of weakness are often on this scale. Models by Marsh (*5*) suggest that the settling rates of such blocks exceed 10 m/sec. Such high settling rates of the blocks and their limited surface area means that they will not significantly cool the surrounding magma. The process will, however, cause upward movement of the magma body as it flows around the descending stoped blocks. Stoping may continue until the roof fails and a caldera forms (Fig. 1C).

The collapsing of the overlying crust allows eruption to continue with the extrusion of large volumes of magma. The remaining magma is now dispersed into smaller bodies around the stoped blocks of crust and a smaller, shallower magma chamber. These continue to cool, differentiate, and interact with crustal blocks, forming the complexly intruded lower portions of calderas.

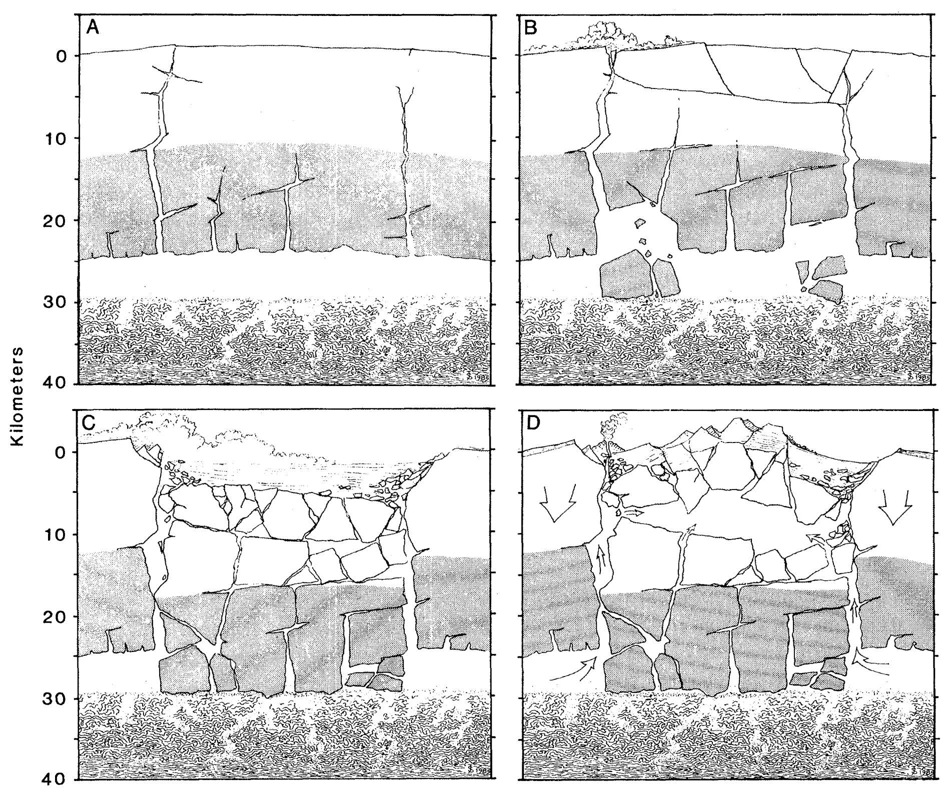

Fig. 1. (A to D) Balanced cross section of intrusion and extrusion model. There is no vertical exaggeration. (A) Initial upward migration of fractures and development of ring fractures. (B) Beginning of eruption, vesiculation, and fracturing of roof rocks. (C) Caldera subsidence. (D) Resurgence and intrusion along ring fractures.

As the blocks settle, the residual magma is squeezed upward into shallower magma bodies. In addition, the topographic depression of the caldera produces an isostatic readjustment. The surrounding terrain subsides into the residual magma chamber, with the magma being forced upward to cause resurgence of the caldera and establish isostatic equilibrium (Fig. 1D).

Since the magma is more dispersed, it will cool and crystallize much more rapidly than a single, large chamber. The heat also drives convection of ground water and alteration, most of which will be localized around shallow apophyses rather than in a regional pattern. The time it takes to establish a more regional convection and alteration system depends on the thickness of the overlying roof (*10*). The process may continue with satellite eruptions of smaller volume from remaining magma bodies, as in the case of the central San Juan Mountains, or with new injections from depth on the same scale.

The result is a crustal cross section consisting of (i) smaller bodies of magma in the catazone, intimately associated with plastically deformed underlying material and large pieces of crustal material showing various degrees of reaction with the cooling magma; (ii) zones of metamorphic screens surrounded by intrusive material; (iii) several periods of intrusion corresponding to the main batholith emplacement, latter batholith emplacement, and small-scale intrusions of residual melts; (iv) intensely injected and altered upper crustal or volcanic material; (v) blocks of the caldera floor composed of previous volcanic rocks and upper crustal material and overlain by the cogenetic ash-flow tuffs; and (vi) later caldera-filling materials surrounding zones of resurgence (Fig. 1D).

The emplacement model effectively duplicates that described for portions of the Andes by Pitcher (*11*). The space problem inherent in the intrusion of a batholith is solved by the subsidence of large crustal blocks through the magma, driven by the decreased reservoir pressure. This model also explains why the intrusion of batholiths often follows ash-flow eruptions, with the main phases of intrusion extending up into the ash-flow tuffs. The relation would be reversed if the shallow magma chamber were first intruded and then ash-flow eruption occurred.

This model explains why mineral data from several large ash flows appear to record high pressures of phenocryst equilibration (*7*). In the current model, the time between residence in the mid-crustal magma chamber and eruption would be too brief to allow phenocryst reequilibration.

The model successfully predicts that large

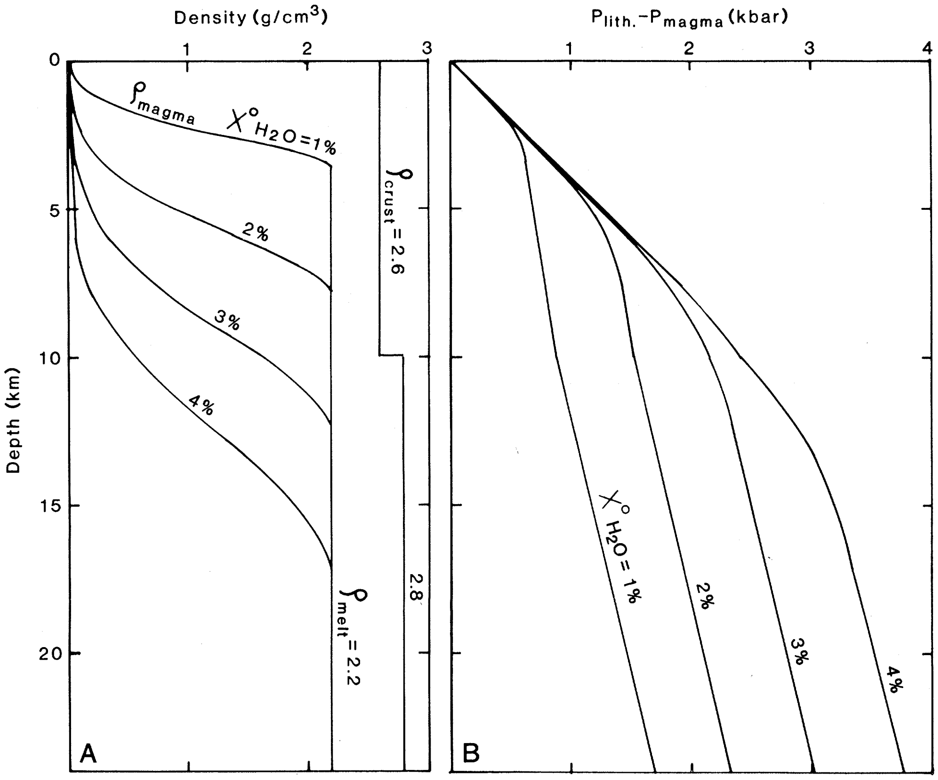

Fig. 2. Density (A) and pressure (B) model for an erupting magma body with various initial water contents ($X^{\circ}_{H_2O}$). The surface density in (A) is not zero, but on this scale is very close to the origin because of the low density of the vapor phase. Varying density model for crust and melt does not change the results significantly. Pressure differential in (B) is a maximum, as nucleation and vesiculation do not necessarily reach equilibrium conditions. See (*9*) for solution model.

magma chambers underlying areas such as the Long Valley area would be at depths of 15 to 20 km, with only isolated small apophyses extending to shallow depths (*4, 8*). The model also predicts that regional hydrothermal alteration driven by the deep magma chamber would occur substantially after the eruption process (*10*). This certainly seems to be the rule in most batholithic terrains. Furthermore, it predicts that the anomalously high heat flow which would follow an eruption such as the Bishop or Yellowstone tuffs would not be a regional, broad thermal anomaly, but would be concentrated above small, residual, shallow bodies. This pattern appears to be what has been reported for such caldera areas (*12*).

The overall resulting cross section (Fig. 1D) is substantially that reported for a number of batholiths where there is sufficient vertical relief to give a good crustal profile (*11, 13*). Therefore, the proposed model for intrusion and eruption being driven from a deep magma chamber appears to be a valid substitute for a shallow magma chamber model where the intrusion of a laterally extensive, shallow magma chamber precedes eruption of the ash-flow tuffs and associated caldera collapse.

If this model is substantially correct, there are many implications for the earth sciences:

1) Major ash-flow eruptions in environments such as Long Valley, Yellowstone, or the San Juan Mountains during the Oligocene may not be preceded by the emplacement of a shallow, extensive magma chamber, but may start directly from a deeper chamber. Therefore, the warning events may not be more significant than what is now occurring at Mammoth Lakes.

2) The intrusion of upper crustal batholiths in these areas may be initiated by the lowered reservoir pressure during ash-flow eruptions and not precede it. This initial intrusion may be accomplished by the stoping of kilometer-sized blocks of crust, as suggested for the Andes (*11*).

3) The collapse of large calderas is the surface manifestation of the stoping process, which would be the magmatic equivalent of block caving and the formation of a glory hole in mining.

4) The resulting shallow magma chambers that drive hydrothermal circulation for hydrothermal power projects may be discontinuous, dispersed, and much shorter in cooling history than would be suggested by the single shallow magma chamber model.

5) Regional circulation caused by the formation of an extensive batholith may occur significantly after the caldera-forming event, as the regional magma chamber may be deeper than supposed. If regional circulation is important in ore formation, the process may occur after volcanism stops.

Because these consequences are so far-reaching, future research should document the nature of the upper and lower margins of batholiths, the chronology of initial batholith injection and cogenetic ash-flow eruptions, the distribution and extent of shallow magma chambers under recent calderas, and the movement of magma in the mid-crust under active areas. A national drilling program, such as that outlined by the National Academy of Sciences (*14*), would be most beneficial for such research.

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