



The volcanic–plutonic connection as a stage for understanding crustal magmatism

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Abstract

The Earth's magmatism produces both volcanic and plutonic rocks. These two rock types share many similarities, but also display significant differences that have led to a tendency to view (and study) them as separate realms. This review tries to bridge the gap to provide more incentive to integrate data from all *magmatic* rocks in a hope to better understand the processes that led to the differentiation of the Earth and generation of a silicic continental crust. We strongly reinforce recent statements made in the literature suggesting that most disparities between volcanic and plutonic rocks can be resolved if volcanic rocks are seen as erupted melt-rich regions (magma “chambers”) expelled from crystal-rich (mushy) reservoirs, which later crystallize to form plutons.

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1. Introduction

Despite many attempts during the last half century to bring volcanic and plutonic rocks into a common framework (Buddington, 1959; Smith, 1960; Lipman, 1984; Wyborn and Chappell, 1986; Miller and Miller, 2002; Clemens, 2003; Metcalf, 2004; Lipman, 2007), the volcanic and plutonic environments are still regarded by some as two different realms. Until the mid-20th

century, although most petrologists had accepted the idea that granite was a magmatic rock, influential researchers claimed that the mechanisms responsible for granite generation could not directly apply to the generation of silicic volcanic rocks (e.g., Read, 1957). In response to this skepticism, many igneous petrologists since, starting with Buddington (1959) and Smith (1960), pointed out the strong kinships between shallow granitoids and silicic volcanism. Fundamental questions persist in how these two rock types relate to each other. For example,

- a. At the most basic level, how do plutons and volcanic rocks relate petrogenetically? Are they two faces of the same coin, or do they arise from different processes (see Marsh, 1988; Marsh, 1990; Sparks, 1990 for a discussion)?

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- b. Accepting that there is some connection between volcanism and plutonism: Did all plutons lose magma at some point by eruption, or did some (or most) reach their solidi before any eruption could occur? And for those that did erupt, were the erupted products systematically different from the retained material that solidified as plutonic rock?
- c. Do plutons only provide a near-solidus view of magmatic systems (e.g., Glazner et al., 2004) or do they preserve a useful record of magma chamber processes (e.g., Wiebe, 1994; Robinson and Miller, 1999; Barnes et al., 2001; Miller and Miller, 2002; Wiebe et al., 2002; Metcalf, 2004)?
- d. How are large magmatic systems built spatio-temporally? Are they assembled in pulses, during intense, flare-up episodes (Ducea, 2001; de Silva et al., 2006a; de Silva and Gosnold, this volume) or more steadily over time?

This paper focuses on intermediate and evolved rock compositions (SiO_2 contents >60–65 wt. %) of calc-alkaline affinities. The less abundant alkaline magmas, which do not evolve towards such silica-rich end-members, behave generally in a similar way despite being less viscous; they also produce both volcanic and plutonic rocks, and can lead to highly explosive volcanic eruptions (e.g., the Naples area with Vesuvius volcano and the Campi Flegrei caldera: Sigurdsson et al., 1982; Civetta et al., 1997). We will review ideas and models that we believe are important for crustal magmatism and call for more integration of plutonic and volcanic data sets as a step toward better understanding of the processes that lead to the observed petrologic diversity on our planet. We stress that the purpose of this review is to state the questions as clearly as possible within the framework of the Earth's magmatism, leaving the great task of answering some of them to papers of this volume.

2. Plutons vs. volcanoes: observations

2.1. Similarities

We consider the following to be the most important well-documented similarities between volcanic and plutonic rocks (Table 1).

2.1.1. Spatial kinship

As recognized long ago (at least since Buddington, 1959; Smith, 1960), volcanic and plutonic rocks occur in the same tectonic settings. In addition, spatial kinship is directly observable in many locations, where erosion has exposed both the remnants of the volcanic sequence and

the upper parts of apparently co-genetic intrusions (cf. Hamilton and Myers, 1967). Examples include the Andes (Myers, 1975a,b), the Latir and Organ Mountains areas in New Mexico (Johnson et al., 1990), the San Juan Volcanic field in Colorado (Lipman, 1967; Hon and Lipman, 1989; Lipman, 2000; Lipman, 2007), the Atesina–Cima d'Asta volcano–plutonic complex in northern Italy (Barth et al., 1993), the Colorado River region of southern Nevada and adjacent Arizona (Miller and Miller, 2002; Metcalf, 2004), and the Sifton range in northern Canadian Cordillera (Miskovic and Francis, 2006).

2.1.2. Geochemical and petrological kinships

Volcanic and plutonic rocks broadly span similar ranges of whole-rock composition (SiO_2 contents from <45 wt.% to >75 wt.%), although the compositional range extends to somewhat lower SiO_2 in plutonic rocks (presumably because uneruptable mafic cumulates occur in plutons). Volcanic and plutonic rocks also have similar mineral phases (but slightly different modal abundances; see next section); typical mineral assemblages (not taking into account alkaline rocks) in mafic rocks include olivine, pyroxene (both clino- and orthopyroxene), and plagioclase (+minor phases), whereas hornblende, biotite, pyroxene, plagioclase, alkali feldspar and quartz can be found in the more silicic endmembers. Finally, similar ranges in isotopic ratios are typically found in spatially and temporally related volcanic and plutonic rocks (Fig. 1).

2.1.3. Geophysical kinship

When erosion levels preclude direct observation of the roots of a volcanic province, geophysical methods come into play. The presence of silicic batholiths has been inferred from the common presence of negative gravity anomalies centered on the source calderas of ignimbrites in volcanic fields (Plouff and Pakiser, 1972; Heiken et al., 1990; Masturyono et al., 2001; de Silva et al., 2006a,b, Fig. 2). Moreover, seismic experiments in areas such as

Table 1

Summary table of first-order similarities and differences between plutonic and volcanic rocks (see text for details)

Volcanoes and plutons	
Kinships	Differences
Spatial proximity	Compositional range
Petrological, mineralogical, and geochemical affinities	Hydrous mineral abundance
Evidence for erupted plutons	Crystal fractionation signature
Geophysical evidence of "plutons" or crystal mushes beneath large calderas	Preserved record (Peak (volcanic) vs. late (plutonic) magmatic stages)

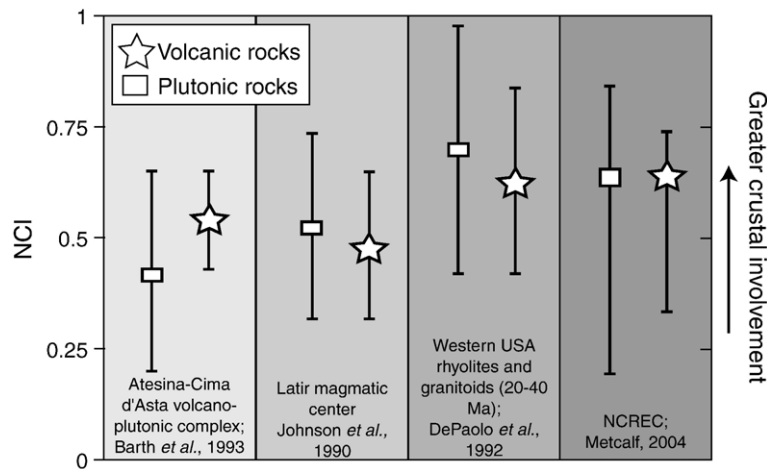


Fig. 1. Variation of the NCI value (Neodymium Crustal Index, DePaolo et al., 1992) for co-genetic volcanic and plutonic rocks from the Atesina–Cima d’Asta volcano–plutonic complex (Barth et al., 1993), Latir magmatic center (Johnson et al., 1990), northern Colorado River extensional corridor (NCREC, Metcalf, 2004) as well as some 20–40 Ma rhyolites and granitoids from the western USA (DePaolo et al., 1992). Modified from Bachmann and Bergantz (2004).

the Central Andes, Yellowstone, Valles, Vesuvius–Campi Flegrei, and Toba calderas (Table 2) have all revealed crustal-scale magmatic systems beneath large, active calderas (Dawson et al., 1990; Lutter et al., 1995; Weiland et al., 1995; Zollo et al., 1996; Steck et al., 1998; Miller and Smith,

1999; Auger et al., 2001; Masturyono et al., 2001; Wilson et al., 2003). The largest identified low velocity zones is present in the Central Andes (at depths of 17–19 km), beneath the Altiplano–Puna Volcanic Complex (Zandt et al., 2003a; de Silva et al., 2006a,b). Velocity models of

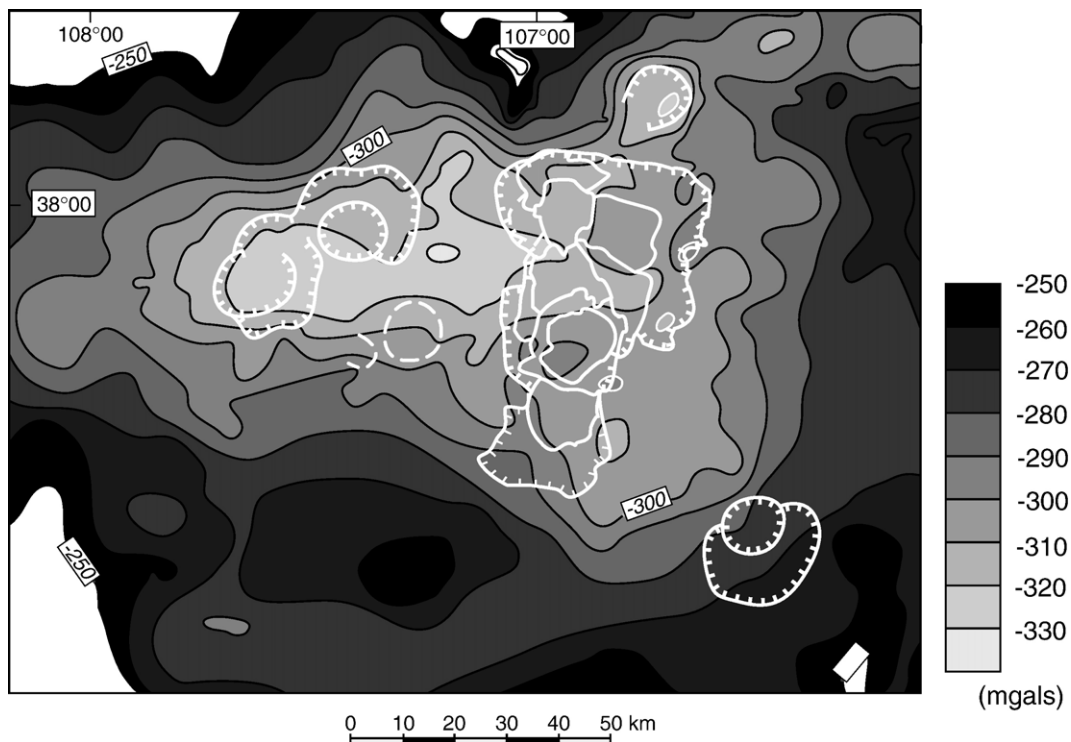


Fig. 2. Outlines of the topographic rims of calderas from the San Juan volcanic field superimposed on the Bouguer gravity anomaly (modified from Plouff and Pakiser, 1972; Drenth and Keller, 2004).

Table 2

Summary characteristics of seismically imaged large silicic volcanic systems

Volcanic System	Observation	Interpretation	Source
Long Valley, CA, USA	• 15% low velocity zone at 6 km depth	• The residual Bishop Tuff magma chamber with 7 to 100% partial melt beneath the caldera. • Basaltic magmas ponded in the mid crust	Weiland et al. (1995)
Age (Volume)	• 25–30% low-velocity zone centered at 11.5 km		
0.73 Ma (700 km ³ — Bishop Tuff)	• 15% low velocity zone at a depth of 24.5 km		
Areal extent — 15 × 30 km			
Valles, NM, USA	• 24% low velocity zone located between 10 and 15 km depth	• Residual Valles magma chamber	Lutter et al. (1995)
Ages (Volumes)	• Broader 10% low velocity zone between depths of 30 and 35 km	• Basaltic magma ponded at the base of the crust	Steck et al., 1998
1.6 Ma (300 km ³)			
1.3 Ma (300 km ³)			
Areal extent — 15 × 30 km			
Coso, CA, USA	• Shallow layer thickness ~ 5 km	• Single shallow	Wilson et al. (2003)
Age (Volume)	• Shear velocity decreasing by 30% across the interface	• magma reservoir (5 km below the surface)	
0.4 Ma to recent (1.5 km ³)		• No lower crustal magma reservoir is inferred	
Areal extent — 10 × 15 km			
Toba, Indonesia	• P-wave velocities 37% below normal	• Upper crustal system at about 10 km	Masturyono et al., 2001
Ages (Volumes)	• Low frequency earthquakes in 20 to 40 km depth range	• Lower system — melt filled basaltic feeder system extending to a mantle source	
840 ka (500 km ³)			
700 ka (60 km ³)			
75 ka (2800 km ³)			
Areal extent — 100 × 30 km			
Yellowstone, WY USA	• 15% decrease from regional P-velocity from 6 to 12 km beneath caldera	• Hot subsolidus granite batholith	Miller and Smith, 1999
Ages (Volumes)	• 30% reduction in P-wave velocity beneath resurgent domes	• Partial melts and residual magma system	
2.0 Ma (>2,500 km ³)			
1.2 Ma (300 km ³)			
0.6 Ma (1000 km ³)			
Areal extent — 7500 km ²			
Altiplano–Puna Volcanic Complex, Central Andes	• Very-low-velocity layer (SV 1.0 km/s) at 17–19 km depth with a thickness of ~2 km	• Large percentage (>> 10%) of partial melt	
Age range (Volumes)		• Mush zone associated with the APVC	de Silva, 1989
10 – 1 Ma (>20,000 km ³)		• Estimated volume of 60,000 km³	Chmielowski et al. (1999)
Areal extent — 70,000 km ²		• No geophysical evidence of a shallower level system	Zandt et al., 2003a,b; de Silva et al., 2006a

magmatic provinces are interpreted as revealing zones of partial melt within the crust (see Lees, this volume). Notwithstanding the fact that physical properties of partially molten rock are poorly understood (and hence it

is difficult to uniquely interpret velocity models as percent melt fraction), these studies have led to the recognition of a common link between shallow and deeper levels of melt-bearing regions that suggests vertical integration of

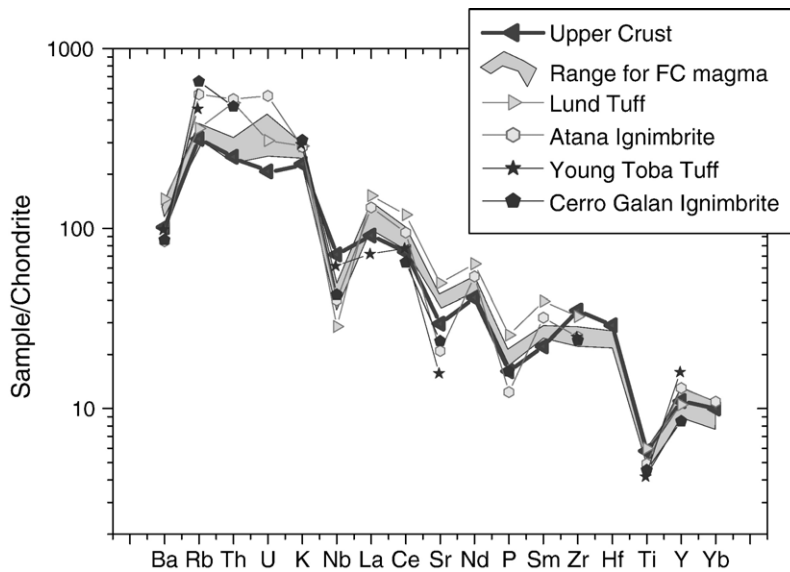


Fig. 3. Trace element diagram showing the compositional similarity between different Monotonous Intermediates and the upper continental crust (trace element concentrations are normalized to chondritic abundances given by Thompson, 1982). FC magma refers to the three units that erupted from the Fish Canyon magma chamber (Pagosa Peak Dacite, Fish Canyon Tuff, and Nutras Creek Dacite, Bachmann et al., 2000).

the magmatic systems feeding calderas. While not definitive, the seismic studies combined with gravity and magnetotelluric surveys (Brasse et al., 2002) lend support to the view that large silicic volcanic systems are underlain by partially molten regions at different depths (ultimately driven by basaltic flux from the mantle), and support the view that volcanic provinces are underlain by shallow and deep plutons in the crustal record.

2.1.4. Crystal-rich volcanic units: remobilized plutons?

In long-lived calc-alkaline volcanic provinces, large, crystal-rich, ash-flow sheets with probable kinship to granodioritic batholiths (and upper crust, which has a granodioritic composition; Taylor and McLennan, 1985; Fig. 3) are commonly erupted (e.g., Steven and Lipman, 1976; de Silva, 1991; Lindsay et al., 2001; Bachmann et al., 2002; Maughan et al., 2002). These large volcanic units, referred to as Monotonous Intermediates (MIs, Hildreth, 1981), provide unambiguous evidence for the presence of large magmatic *mushes* (magma bodies of several thousands of km³ with ~45–50 vol.% crystals) that are eruptable near their rheological lock-up point (a rigid crystalline framework is thought to form at ~50% crystals in these magmas, Vigneresse et al., 1996; Petford, 2003). These Monotonous Intermediates display textural and mineralogical characteristics that are reminiscent of plutons: (1) glomerocrysts and multi-mineral crystal clots are common in these rocks (de Silva et al., 1994; Lindsay et al., 2001; Bachmann et al., 2002, Fig. 4) and (2) they contain a high modal abundance of hornblende + titanite

(Lindsay et al., 2001; Bachmann et al., 2002; Maughan et al., 2002), a rare assemblage in crystal-poor volcanic rocks (Nakada, 1991; Deer et al., 1992) but common in granodiorite batholiths (e.g., Bateman and Chappell, 1979). One noticeable difference is the higher abundance of quartz and K-feldspar in plutonic rocks compared to Monotonous Intermediates, but this observation can be accounted for by crystallization of the highly silicic interstitial melt that these Monotonous Intermediates and their derivatives contain (high-SiO₂ rhyolitic melt near the haplogranite eutectic; Lindsay et al., 2001; Bachmann et al., 2002; Maughan et al., 2002).

In addition to these petrological similarities, large granodioritic plutons and Monotonous Intermediates display similar dimensions. When compared to the largest granodioritic plutons, the calderas collapsing during these Monotonous Intermediate eruptions mimic the shapes displayed by the plutonic bodies. The La Garita caldera, which collapsed during the Fish Canyon Tuff eruption, is roughly the same length (~80 km) and the same width (15–30 km) as the Mount Givens pluton, and also shows (at least) three bulbous segments (e.g., McNulty et al., 2000). Similar dimensions are displayed by other large caldera systems like La Pacana (Lindsay et al., 2001) and Toba (Chesner, 1998). The main difference may reside in the thickness of the magma body. On the basis of volume estimates, the Fish Canyon Tuff is thought to have a draw-down thickness of two kilometers (total volume divided by caldera surface area: 5000 km³/2500 km²=2 km, e.g., Bachmann et al., 2002), whereas the Mount Givens pluton

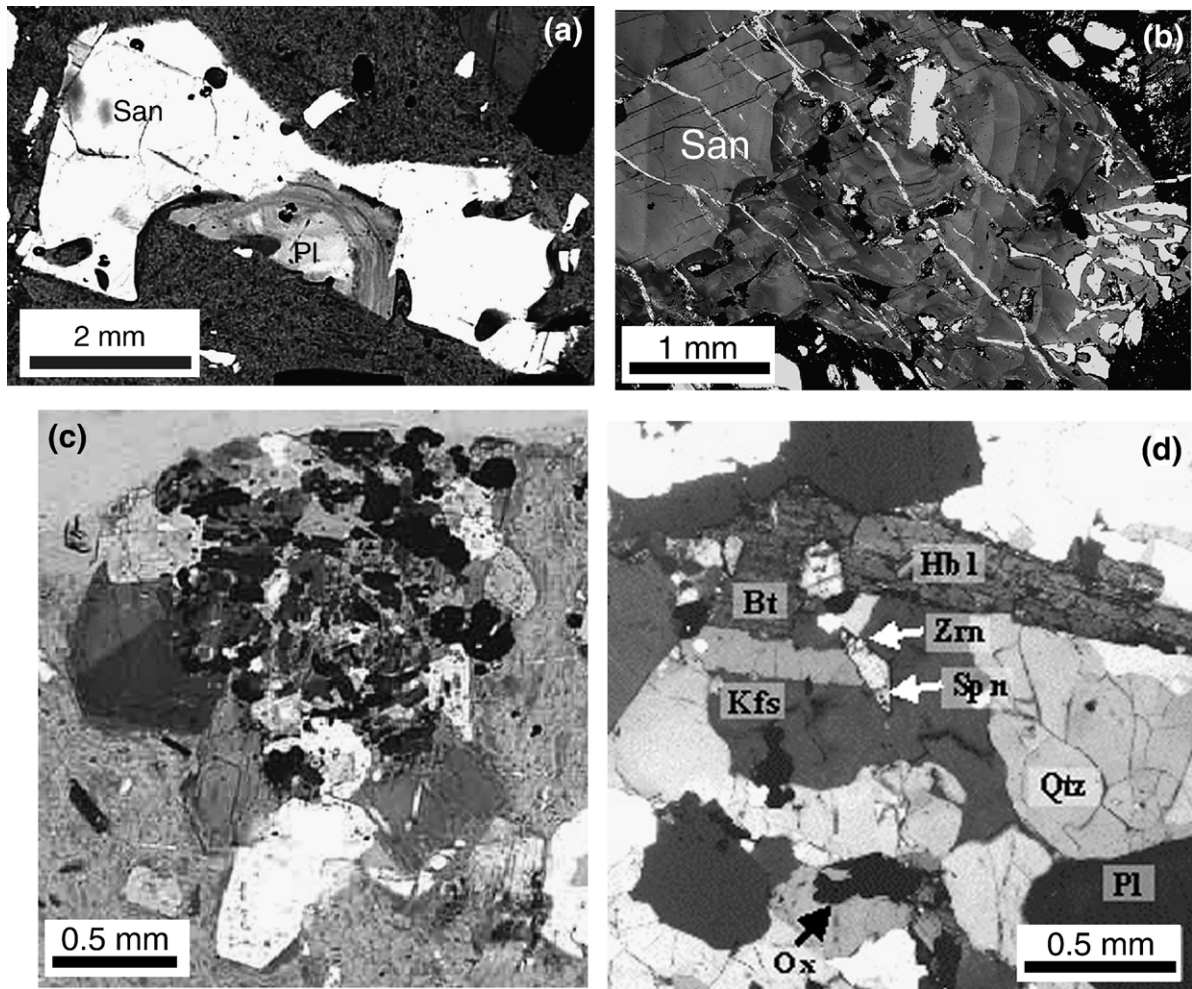


Fig. 4. Photomicrographs of plutonic textures in an erupted rock (Fish Canyon Tuff). (a) plagioclase–sanidine aggregate showing partially melted grain boundaries. (b) large sanidine with granophyric overgrowth. (c) crystal clot consisting of plagioclase, amphibole, biotite and Fe–Ti oxides. (d) co-magmatic granodioritic xenolith found in late-erupted Fish Canyon Tuff (modified from Bachmann et al., 2002).

is approximately 5 km thick (McNulty et al., 2000). The larger volume stored in the plutonic record (which could be even larger if the Mount Givens pluton has lost any magma by eruption, which is likely for such a large, shallow pluton) suggests that even an eruption of the size of the Fish Canyon Tuff may not entirely drain its magma chamber (see further discussion in Section 3.3.2). Such an inference is also supported by (1) the fact that large calderas typically resurge after collapse, marking the presence of large, non-erupted magma in the shallow crust (Smith, 1979; Lipman, 1984); (2) the ubiquitous crystal fractionation signature that most volcanic rocks convey (Fig. 5; see Section 2.2), requiring that some co-magmatic crystalline residue was left in the crust, and (3) geophysical evidence for “plutons” or batholith-scale crystal mushes beneath large calderas and silicic volcanic fields (Dawson et al., 1990; Weiland et al., 1995; Steck

et al., 1998; Miller and Smith, 1999; Masturyono et al., 2001; Zandt et al., 2003b).

Less voluminous, crystal-rich, silicic volcanic units that have been interpreted as remobilized proto-plutons also occur in several locations in modern arcs. Such units include the Chao dacite and the Chascon–Runtu Jarita complex in the Central Andes (de Silva et al., 1994; Watts et al., 1999), Montserrat Andesite (Murphy et al., 2000; Couch et al., 2001), the Mt. Unzen Dacite (Nakamura, 1995), and Kos Plateau Tuff (Keller, 1969; Bachmann et al., 2007a).

2.2. Differences

In addition to the differences in (1) total abundance in the crust (a typical intrusive:extrusive ratio for most magmatic system is ~5:1, when all uncertainties are

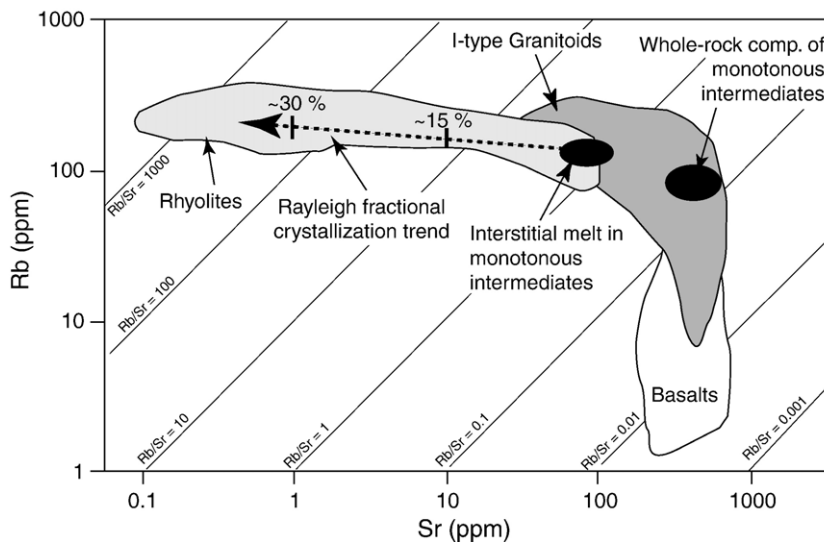


Fig. 5. Rb vs. Sr content for magmatic rocks from several magmatic provinces (western North America, Andes, Great Britain, Australia. Modified from Halliday et al., 1991; Bachmann and Bergantz, 2004). Dashed line is representing a crystal fractionation trend, using the concentration of Rb and Sr in the interstitial melt of the Fish Canyon Tuff (FCT; typical monotonous intermediate) as the initial value and typical bulk partition coefficients for these elements in FCT ($D_{Sr}=12$, and $D_{Rb}=0.5$). Tick marks are 15 and 30% of Rayleigh fractional crystallization. Due to the different behaviour of Sr and Rb in silicic magmas (in the presence of abundant plagioclase, Sr is compatible, and Rb incompatible), higher Rb/Sr ratios highlight a stronger crystal fractionation signature.

considered, White et al., 2006) and (2) abundance and size of the crystalline material (volcanic rocks are defined by the presence of quenched liquid and tend to have small crystals with a relatively low abundance of near-solidus minerals, such as alkali feldspars and quartz, in comparison to plutonic rocks), these two rock types display significant disparities, listed below.

2.2.1. Compositional range

The compositional range in plutonic rocks is greater than that observed in volcanic rocks, extending towards less silicic compositions (although the most abundant plutonic rock types by far are granitoids, forming ~77 vol.% of the upper crust, Table 1.3 of Taylor and McLennan, 1985); the most mafic whole-rock compositions in the magmatic realm (aside from mantle rocks) occur in plutonic rocks (ultramafic cumulates, such as those found in large mafic intrusions).

2.2.2. Crystal fractionation signature

Volcanic rocks are, on average, depleted in compatible elements (and enriched in incompatible elements) compared to their plutonic equivalents. This signature is a well-known fact from large mafic intrusions (see, for example, chapter 6 of McBirney, 1993), but is also present in silicic rocks (Fig. 5), although small-volume granitic rocks (aplites, leucogranites) can also show elemental signatures

of extreme fractionation (e.g. Miller and Mittlefehldt, 1984; Walker et al., 2007-this volume).

2.2.3. Hydrous minerals abundances

Hydrous mafic phases (amphiboles, micas) are more common in plutonic than in geochemically equivalent volcanic rocks. This distinction can be explained by the fact that volcanic rocks are quenched above their solidi, whereas plutonic rocks continue to crystallize to their solidi, in the interval wherein falling T and rising H_2O in residual melt stabilize the hydrous phases (e.g., Naney, 1983; Dall'Agnol et al., 1999). Early crystallizing anhydrous mafic minerals, mostly pyroxenes, are thus commonly preserved in volcanic rocks but largely or entirely reacted out in plutonic rocks.

2.2.4. Preserved rock record

Perhaps the most fundamental difference between volcanic and plutonic rocks lies in the record they preserve. Plutonic rocks mostly document the terminal stages of multiple intrusive events (and varying depths) and can be strongly influenced by late, near- and subsolidus processes (e.g., second boiling effects, textural coarsening, metamorphic reactions, deformation). These late processes may partly or completely overprint the early history of the rock (e.g., Zieg and Marsh, 2005), although some plutonic exposures do preserve information about early stages of the

intrusive history (e.g., Miller and Miller, 2002). In contrast, volcanic rocks provide an instantaneous snapshot of the state of the magmatic system at the time of eruption (although in both cases a complex history may be preserved in minerals by chemical zoning and various populations of melt inclusions). Moreover, they may record the peak of magmatic activity in a given province, rather than late, waning stages as plutons do.

3. Discussion

Reconciling the terminal record of plutons with the instantaneous image of a volcanic unit is potentially the most difficult task that igneous petrologists face, and most likely led to the controversies still alive today. General questions that we try to address in this review are:

1. Can plutons be used as arguably complex but representative images of magma chambers and processes therein (Wiebe et al., 2002; Harper et al., 2004; Walker et al., 2007-this volume) or did near-solidus processes largely eliminate clues to their earlier magmatic histories (Glazner et al., 2004; Zieg and Marsh, 2005)?
2. How do we petrogenetically link plutons and volcanic rocks, and how does this relationship fit into the global framework of the Earth's magmatism?

Other important questions that we pose include: Do large volcanic systems, like plutons, grow incrementally, or are they formed by a single episode of magma intrusion? Do all plutons erupt, or do some stall and remain sealed? And, conversely, are plutons formed beneath all sizable volcanic systems? We begin with a short review of differentiation mechanisms in crustal magmatism to set the stage as precisely as possible, and continue with a discussion on characteristics of the different rock types, sources and processes that are thought to be important in crustal magmatism.

3.1. Mechanisms of magmatic differentiation

Magmatic differentiation is thought to operate either in the liquid state (liquid immiscibility and thermogravitational diffusion) or by the physical separation of crystals from liquid (see review of Wilson, 1995). Although liquid state differentiation (in particular liquid immiscibility) may be important for magmas such as carbonatite (e.g., Freestone and Hamilton, 1980; Pinkerton et al., 1995), crystal-liquid fractionation (in its various forms) is considered to be the main drive for magmatic differentiation (at least since Harker, 1894). As pointed out by many since Bowen (1928), chemical

fractionation induced by crystal-liquid separation can be theoretically achieved by several physical mechanisms, depending primarily on the crystallinity of the magma:

1. At low crystallinity (<40 vol.% crystals), the solid particles are predominantly removed from the melt by particle sedimentation. Crystal agglomeration is likely to occur (forming crystal-rich plumes and leading to dripping instabilities, Bergantz and Ni, 1999). Stokes settling is not very likely to occur (e.g., Sparks et al., 1984), but can provide some minimum sedimentation rate (using a hindering function when particle concentration $> \sim 5\%$, e.g., Davis and Acrivos, 1985).
2. At high crystallinity (>60 vol.% crystals), liquid can be extracted from solid residuum by compaction (Bowen, 1928; McKenzie, 1984; McKenzie, 1985; Tait and Jaupart, 1992; Bachmann and Bergantz, 2004; Hildreth, 2004) and/or filter-pressing (including by mush fracturing, Sisson and Bacon, 1999; Bachmann and Bergantz, 2003; Petford and Koenders, 2003; Bachmann and Bergantz, 2006).
3. In magmas of intermediate crystallinity ($\sim >40$ and <60 vol.%), crystal-liquid separation can occur by both crystal settling and compaction, depending on local conditions. In addition, a process described by (Niemi and Courtney, 1983) and referred to as microsettling (Miller et al., 1988) can occur at these crystal fractions; following the establishment of a self-supporting skeleton of crystals, micro-rearrangements of the crystalline framework takes place, leading to the expulsion of some of the interstitial melt towards the top.

For a specific magmatic system, the feasibility of each of these processes will obviously depend on the local conditions (gradients in crystallinity are ubiquitous in magma chambers) and on parameters such as (a) the density and viscosity of the melt and (b) the presence of convective movements (and the convective velocities if applicable).

Besides these closed-system differentiation mechanisms, open-system processes such as magma mixing and wall-rock assimilation clearly also contribute to petrological diversity. The blending of two dominantly liquid pods of magma has been long recognized as an important magmatic process (at least since Bunsen, 1851) and the assimilation of crustal melts by hot, mantle-derived magmas (often accompanied by crystallization) is very commonly called upon to explain the variations in magmatic series (AFC, e.g., Allègre and Minster, 1978; Taylor, 1980; DePaolo, 1981; DePaolo et al., 1992). However, since it requires the existence of compositional

endmembers, it only produces intermediate products, and cannot be the ultimate cause of the magmatic differentiation. For example, if one looks far enough back in time, there was no or very little (although Hadean granitoids may have existed, e.g., [Watson and Harrison, 2005](#)) pre-existing evolved crust (i.e., granitoids) that could act as a silicic endmember in the early Earth. The voluminous silicic magma bodies formed at that time (the Archean Tonalite–Trondhjemite–Granodiorite suites) must have been dominantly derived by igneous distillation induced by crystal-liquid separation. Potential models include either remelting of subducted oceanic crust ([Martin, 1986](#); [Drummond and Defant, 1990](#); [Prouteau et al., 2001](#); [Foley et al., 2002](#); [Rapp et al., 2003](#)), crystallization of hydrous basalts in the upper mantle ([Prouteau and Scaillet, 2003](#); [Macpherson et al., 2006](#)) or remelting of the base of a thickened lithosphere (e.g., [Smithies, 2000](#)).

3.2. Mafic compositions

The relationship between basalts *s.l.* and mafic plutonic rocks is relatively clear. Mantle melting generates a basaltic composition, should it be by decompression ([Kushiro, 2001](#)) or addition of fluids ([Ulmer, 2001](#); [Parman and Grove, 2004](#)). Therefore, by definition, the most primitive basalts do not need to change much from source to eruption, and do not require any associated rock other than the residual peridotite. However, most basalts are not primary (they left some residual cumulates in the crust or upper mantle). Many authors have claimed that the coarse-grained plutonic rocks have formed by crystal accumulation from cooling basalts (e.g., [Daly, 1933](#); [Wager et al., 1960](#); [Wager and Brown, 1968](#); [McBimey and Noyes, 1979](#); [Irvine, 1982](#)), should it be by crystal settling or *in-situ* crystallization. Therefore, the relation between plutonic and volcanic rocks is relatively straightforward; non-primary basalts are the liquids that escaped crustal, mafic magma chambers, particularly well preserved today in many places as layered mafic intrusions.

The more controversial question with mafic rocks is why so few plutonic rocks are present in the crust when several considerations suggest that more mafic cumulates than are actually seen should be found in the crust. Most important is the fact that the continental crust is too evolved (felsic) to be a melt in equilibrium with the mantle (e.g., [Rudnick, 1995](#)); there must be a “missing” mafic end-member, required to balance the continental crust composition. Therefore, several workers have suggested that mafic components must return to the mantle ([Arndt and Goldstein, 1989](#); [Kay and Mahlborg Kay, 1993](#); [Rudnick, 1995](#)). Indeed, mafic lithologies in the deep crust that are deficient in plagioclase and rich in garnet (the consequence

of high pressure crystallization and loss of melt enriched in plagioclase component) can be denser than the typical underlying mantle lithologies. This results in a density inversion that permits formation of dripping instabilities on time scales of $<10^7$ yr at temperatures greater than 700 °C ([Jull and Kelemen, 2001](#); [Dufek and Bergantz, 2005](#)).

3.3. Silicic compositions

The plutonic–volcanic connection among silicic rocks (>65 wt.% SiO_2) remains controversial for several reasons. First, most granitoid plutons display complex internal contacts, evidence for numerous reintrusion events ([Wiebe and Collins, 1998](#); [McNulty et al., 2000](#); [Mahan et al., 2003](#); [Zak and Paterson, 2005](#)). These field observations, complemented by geochronological evidence for emplacements over millions of years, have led some authors (see, in particular, [Glazner et al., 2004](#), but also [Annen et al., 2006b](#)) to suggest that large plutons are the result of amalgamation of multiple smaller magma bodies, and were never (or very rarely) large, integrated magma chambers ([Coleman et al., 2004](#)). This hypothesis creates a gap between volcanic and plutonic rocks; if voluminous plutons never formed a large magma chamber, they could not be the source of the caldera-forming eruptions with volumes $>10^2$ – 10^3 km³. In turn, the large ignimbrites, such as the Monotonous Intermediates, should not leave large batholithic bodies in the crust ([Glazner et al., 2004](#)). However, although it seems indisputable that (1) plutons are built incrementally, commonly over time spans of hundreds of thousands to millions of years (e.g., [McNulty et al., 2000](#); [Zak and Paterson, 2005](#); [Walker et al., 2007-this volume](#)); and (2) although Monotonous Intermediates haven’t been observed in direct contact with their *source* granitoid plutons, the spatial, compositional, and geophysical kinships between Monotonous Intermediates and granitoid batholiths (see Section 2.1) and the obvious whole-rock homogeneity of both Monotonous Intermediates and granitoid rocks strongly suggest that plutons can form integrated magma chambers undergoing some chamber-wide stirring. These chambers are most likely never entirely crystal-poor, but form large crystal mushes in the upper to mid crust ([Bachmann and Bergantz, 2004](#); [Hildreth, 2004](#); [Zak and Paterson, 2005](#)) that have been imaged seismically (e.g., [Dawson et al., 1990](#); [Weiland et al., 1995](#); [Masturyono et al., 2001](#); [Zandt et al., 2003a](#)).

Incremental growth over long periods of shallow magma chambers, which is used as a major argument setting plutonic bodies apart from giant eruptions ([Glazner et al., 2004](#)), is not easily ruled out in erupted rocks. On the contrary, some evidence suggests that most

silicic volcanic rocks come from long-lived magma chambers that did grow episodically. Obviously, the eruptive process erases all textural evidence for sporadic magma emplacement at depth, but the following arguments all appear to be incompatible with a single intrusion event leading to a major volcanic burst: (1) the complex thermal evolution of silicic magmas revealed by chemical zoning in minerals of erupted rocks (Pallister et al., 1992; Bachmann and Dungan, 2002; Devine et al., 2003; Rutherford and Devine, 2003); (2) the very large volumes of magma involved in eruptions of some silicic magmas ($>1000 \text{ km}^3$); (3) the protracted time scales of magma assembly as indicated by zircon U/Th/Pb geochronology (up to 600 ka, Reid et al., 1997; Brown and Fletcher, 1999; Miller and Wooden, 2004; Vazquez and Reid, 2004; Bacon and Lowenstern, 2005; Charlier et al., 2005; Bachmann et al., 2007a,b); and (4) progressions of volcanism suggesting sequential construction of composite plutons on time scales of several million years (Grunder et al., 2007). Of particular note are nested multi-cyclic caldera complexes. Systems such as Valles, New Mexico (Smith and Bailey, 1968), the Central San Juan caldera cluster (Lipman, 2000), Toba, Indonesia (Chesner et al., 1991), Cerro Galan, Argentina (Sparks et al., 1985), Yellowstone (Lowenstern et al., 2006) and La Pacana, Chile (Lindsay et al., 2001) are composite collapse features produced by repeated major ignimbrite producing eruptions with sequential and overlapping subsidence. Each individual eruption and collapse would represent direct evidence for a shallow magma body at the time of eruption with the integrated result being a composite pluton of “nested” intrusions built in time scales from $<1 \text{ Ma}$ to several Ma.

A second fundamental difference that seems to distinguish granitoids from silicic volcanic rocks is the fact that petrogenetic models favored by the community diverge for the two rock types. The two most widely accepted processes for the generation of silicic magma (not mutually exclusive, as we will see below) are:

1. Partial melting of pre-existing rocks (in particular of meta-basalts; see Sisson et al., 2005 or Dufek and Bergantz, 2005, for comprehensive recent summaries of numerous papers).
2. Differentiation by crystallization of mantle-derived magmas (Bowen, 1928), either deep in the crust (e.g., Müntener et al., 2001; Mortazavi and Sparks, 2004), or at shallow depths (Grove et al., 1997; Pichavant et al., 2002), often associated with some crustal melting to account for the evidence of an open-system (see, among many, Allègre and Minster, 1978; Taylor, 1980; DePaolo, 1981; Davidson and Tepley, 1997; Bohrsen and Spera, 2001).

In the vast number of studies that have looked at silicic plutonic and volcanic rocks in the last decades, nearly every possible model has been suggested for both types of rocks. However, it is interesting to note that *most* models proposed for the generation of granitoids have been inclined towards crustal melting of fertile lithologies (e.g., Clemens and Vielzeuf, 1987; Miller et al., 1988; Beard and Lofgren, 1989; Bergantz, 1989; Brown, 1994; Patino-Douce and Beard, 1995; Petford et al., 1996; Petford et al., 2000; Altherr and Siebel, 2002; Clemens, 2003; Miller et al., 2003; Chappell, 2005; Sisson et al., 2005; Vigneresse, 2005), whereas fractional crystallization of more mafic parents (often coupled with some crustal assimilation) has been favoured for many silicic volcanic rocks (Hildreth, 1979; Mahood, 1981; Michael, 1983; Bacon and Druitt, 1988; Halliday et al., 1991; Metz and Mahood, 1991; Duffield et al., 1995; Francalanci et al., 1995; Druitt et al., 1999; Wolff et al., 1999; Anderson et al., 2000; Hildreth and Fierstein, 2000; Lindsay et al., 2001; Ginibre et al., 2004). How can we reconcile the fact that plutonic and volcanic rocks appear genetically related, but are thought to arise from different petrogenetic processes? Does the undeniable crystal fractionation signature found in silicic volcanic rocks (Fig. 5) necessarily require that fractional crystallization of more mafic parents is the dominant differentiation process in volcanic rocks? If the answer to this last question is yes, are therefore most plutonic rocks *not* formed dominantly by crustal melting? Alternatively, is there a way to obtain this crystal fractionation signature in silicic volcanic rocks from a magma that may have ultimately formed by melting of crustal lithologies?

To circumvent this plutonic–volcanic petrogenetic dichotomy, Hildreth (2004) and Bachmann and Bergantz (2004) have proposed a model that allows for a common petrologic evolution of volcanic and plutonic rocks until it diverges during shallow storage in the crust. As pointed out by Thompson (1972) and Brophy (1991), large silicic magma bodies stored in the relatively cold upper crustal environment are likely to crystallize until they reach “viscous death” (approximately 50 vol.% crystals; e.g., Vigneresse et al., 1996) without undergoing much crystal-liquid separation (due to the slow settling of crystals in the viscous silicic melt and the occurrence of, admittedly slow, convective currents). Once the rheological lock-up has been attained, significant changes take place in the magma chamber. First, in addition to being buffered by latent heat release, cooling can only occur by

(slow) conduction and is periodically counteracted by influx of hot magmas coming from below. Therefore, it is likely that magmas remain as mushes for a fairly long period (zircon crystallization suggests ~ 100 ka to >1 Ma, Reid et al., 1997; Brown and Fletcher, 1999; Vazquez and Reid, 2002; Miller and Wooden, 2004; Bacon and Lowenstern, 2005; Charlier et al., 2005; Bachmann et al., 2007a,b; Walker et al., 2007-this volume). Second, the absence of chamber-wide stirring by convection allows crystal-liquid separation to occur; some of the interstitial melt can be expelled upward as it is less dense than the surrounding crystal framework. Melt expulsion from crystalline mushes is slow, but if this mush state is preserved for a long enough period ($>10^4$ – 10^5 yr), it will produce a sill-like cap of crystal-poor magma (Bachmann and Bergantz, 2004), which will be kept from rapidly and thoroughly crystallizing due to the presence of water and heat stored in the mush (Hildreth, 2004). Being expelled from a crystalline mush, the crystal-poor rhyolite cap will have a more obvious crystal fractionation signature than its residue (the granitoid), which will show a complementary cumulate record (Miller and Miller, 2002). However, the cumulate signature in the plutonic “residue” will remain subtle as only a small proportion of the interstitial liquid will escape (~ 10 – 20% , Bachmann and Bergantz, 2004).

In summary, it seems that the remaining controversy regarding the relationship between *silicic* volcanic and plutonic rocks may be due to a fainter geochemical signal than in the case of mafic rocks; whereas the “cumulate–residual melt” relationship is obvious between non-primary basalts and mafic plutonic rocks, it is generally dimmer in the silicic realm (although it can be very strong in the case of very felsic plutonic rocks, e.g., Miller and Mittlefehldt, 1984). The likely cause for this difference is the viscosity of the melt, which is much higher in the silicic case, inducing a less-efficient crystal-liquid separation. Therefore, only limited separation can occur before silicic systems reach their solidi or erupt (although it is sufficient to form $500 + \text{km}^3$ crystal-poor silicic bodies due to the sill-like geometries of silicic crystal mushes, which allow crystal-melt separation to occur on a large surface area). Nonetheless, we believe that many observations of silicic systems support the assumption that most granitoids are, like the layered mafic intrusions, frozen magma chambers that have lost some melt by crystal-liquid separation followed by eruption (see also Metcalf, 2004).

3.3.1. Do all plutons erupt?

Shallow magmatic intrusions unequivocally connect with the volcanic realm. Examples include exposed resurgent plutons (e.g., Johnson et al., 1990), crypto-

domes and dikes within volcanic edifices, and in some cases very shallow laccoliths. At slightly deeper levels, “blind” dikes that terminated beneath the surface and small pods, plugs, and sills also are readily envisioned as parts of volcanic systems. Furthermore, the presence of co-magmatic holocrystalline fragments in many felsic volcanic rocks provides good evidence that eruptable magma bodies are commonly surrounded by shallow intrusive roots (Ewart and Cole, 1967; Bacon, 1992; Lipman et al., 1997; Brown et al., 1998; Burt et al., 1998; Bachmann et al., 2002; Charlier et al., 2003; Bacon and Lowenstern, 2005).

It is sizeable plutons that comprise 10 's of km^3 or more, almost invariably with roofs at depths >4 km, that have engendered controversy concerning the volcano–plutonic connection and the question of eruptive history of intrusive bodies. Relatively large plutons must either mark stalled magmas that never erupted at all or the complementary residue that was left beneath the surface during eruptions. Despite the blurring effect of near- and sub-solidus processes, they may preserve valuable records of their integrated histories that shed some light on possible relationships to erupted products.

3.3.1.1. Shallow-level plutons. Not surprisingly, shallower level plutons (roofs $<\sim 8$ km) appear to show the closest kinship with volcanic counterparts. They commonly preserve remnants of quenched porphyry border zones that may be almost indistinguishable from dikes in the country rock and even from overlying volcanics (e.g., Seager and McCurry, 1988; Seaman et al., 1995; Verplanck et al., 1995; Verplanck et al., 1999; Seaman, 2000; Dodge et al., 2005; Hodge et al., 2006), demonstrating initial emplacement into shallow, cold country rocks. Relict porphyry borders, and in some cases thicker texturally and compositionally gradational zones, may record the preservation of solidification fronts (e.g., Marsh, 1996; Bachl et al., 2001). Vesicle-rich zones at roofs reflect the shallow depths and mark ascent of either fluid-rich melt or bubbles through the magma (Gualda et al., 2004). Where vertical crustal sections are exposed, both smaller hypabyssal intrusions and coeval volcanic sections may be present, and it appears inescapable that many such intrusions erupted (Dilles, 1987; Faulds et al., 1995; Metcalf et al., 1995; Seaman et al., 1995; Bonin et al., 2004; Wada et al., 2004; Dodge et al., 2005; Hodge et al., 2006).

Field relations and geochronology reveal that emplacement, at least in shallow to mid-crustal plutons, is incremental, in some cases over periods of millions of years (Wiebe, 1996; Coleman et al., 2004; Glazner et al., 2004; Annen et al., 2006b; Walker et al., 2007-this volume). Increments may be compositionally mono-

tonous and only subtly distinguishable in the field (Glazner et al., 2004; Walker et al., 2007-this volume), or they may contrast dramatically (Wiebe, 1993; Falkner et al., 1995; Metcalf et al., 1995). Evidence for mafic input into felsic chambers is widespread, but the style and volume of intrusion appears to vary greatly. Mafic input can be marked only by relatively small, ellipsoidal, fine-grained dioritic enclaves; by local concentrations of quenched mafic pillows; by disrupted, synplutonic mafic dikes; or by lava flow-like sheets, interpreted to have been emplaced at the extant base of a magma chamber (Didier and Barbarin, 1991; Wiebe et al., 1997; Wiebe and Collins, 1998; Miller and Miller, 2002).

Many shallow intrusions preserve stratification in one form or another (Bergantz, 1991; Miller and Miller, 2002). In some cases, this is seen as depositional features that have clear surface analogs (sedimentary and lava flow-like structures, Wiebe and Collins, 1998; Miller et al., 2004; Miller et al., 2005), attesting to intrachamber gravity-driven magma dynamics (Bergantz, 2000). Many granites also preserve subtler evidence of crystal accumulation (Wiebe, 1996; Harper et al., 2004; Walker et al., 2007-this volume) and of subhorizontal sheeting that appears to reflect melt segregation (Coleman et al., 2005; Walker et al., 2007-this volume).

Although individual increments may be relatively small, several lines of evidence suggest that, at least periodically, large volumes of melt-bearing mush, and in some cases crystal-poor magma, are present within the plutonic system. We argue that, although entire plutons were probably never melt-rich at any one time, portions of plutons periodically were substantial magma chambers (e.g., Zak and Paterson, 2005; but see Glazner et al., 2004, for a cautionary view). Absence of well-defined internal contacts where U–Pb evidence indicates $\sim 10^5$ – 10^6 yr of zircon growth episodes suggest that some large plutons and batholiths formed from thick mush zones that were constructed by many magmatic increments. Initial intrusive boundaries were blurred or destroyed by continuing crystallization and movement within the mush. These relationships are consistent with the view that the plutons comprise a fluctuating patchwork of melt-richer and melt-poorer zones that respond to repeated replenishment (Vazquez and Reid, 2002; Miller and Wooden, 2004; Vazquez and Reid, 2004; Bacon and Lowenstern, 2005; Walker et al., 2007-this volume). Large blocks within plutons that were clearly derived from roofs and walls are rarely abundant enough to indicate that stoping was a major emplacement mechanism (Yoshinobu et al., 2003; Glazner et al., 2004; Glazner and Bartley, 2006), but they do imply downward transport through a significant melt-rich

column and emplacement in a crystal-rich substrate (e.g., Hawkins and Wiebe, 2004). This in turn suggests the presence of a sizeable magma chamber.

Highly silicic granites ($> \sim 75$ wt.% SiO_2), equivalent in composition to high-silica rhyolite, are very common but rarely voluminous in shallow felsic intrusions. Most are exposed as small aplite dikes, pods, and sheets and roof zones of a few km^2 . Trace and minor element chemistry demonstrates that they are products of fractional crystallization. For example, very low Ba and Sr (10's of ppm or less) seem to require fractionation of feldspars, and very low light and middle REE and Zr/Hf reflect fractionation of accessory minerals (allanite, chevkinite, sphene, zircon; e.g. (Claiborne et al., 2006)). These granites reflect effective extraction of fractionated melt from crystal mushes.

3.3.1.2. Deep-seated plutons. Although shallow felsic magma chambers may not necessarily erupt, all of the features described above from plutons are consistent with processes that can be inferred for subvolcanic chambers from the volcanic record. In contrast, the record of a volcanic connection in deeper-seated intrusions is, as expected, not obvious: (1) Porphyries and vesicular granites are rare or absent, (2) evidence for mafic–felsic interaction, though widespread in the form of mafic enclaves, is generally much less spectacular, and (3) evidence for stratigraphic accumulation and depositional processes is subtler or nonexistent. This fainter volcanic connection to deeper-seated bodies may simply be a consequence of the thermal environment: shallow intrusions look more like volcanic rocks than deeper ones due to the fact that they cool more rapidly and have less time for modification of textures and structure. It is interesting to note that the largest shallow intrusions are known for their long histories and also lack well-defined internal contacts, strengthening the case that the thermal environment dictates the preservation record of intrusive bodies (e.g., Davidson et al., 1994).

3.3.2. The erupted fraction of a magma chamber

A critical follow-up to the question, “Do all plutons erupt?,” would appear to be, “What fraction of a magma chamber erupts during a single eruptive episode?” A rough estimate of $\leq 10\%$ by Smith (1979) has been widely cited and fairly widely accepted for almost 30 years, but new views of magma systems call into question the assumptions on which the estimate was based and require reassessment of this issue. Smith (1979) equated pluton size with magma chamber size. The nature and extent, and even the existence, of magma chambers is now debated, but as discussed above, many plutons are probably

constructed incrementally; at any given time, only a portion of the reservoir is melt-rich and eruptable. This implies that Smith underestimated the erupted fraction. We suggest that large eruptions may indeed evacuate most of the eruptable material (<50% crystals, Marsh, 1981; Petford, 2003) from the “chamber” (e.g. Glazner et al., 2004; Charlier et al., 2005), and even that eruption is inevitable whenever a large melt-rich volume accumulates (assuming that magmatic overpressure is sufficient to generate rhyolite dikes that can propagate to the surface and cause an eruption; Jellinek and DePaolo, 2003).

3.3.3. *Is there a rheological or mechanical control on whether a silicic magma erupts or remains sealed in the crust?*

It is plausible that some deep plutons are remnants of ascending systems doomed to stall due to their low temperature and/or high crystal content. As suggested by Chappell et al. (1998) and Miller et al. (2003), granites may be distinguished as either hot (>800 °C) or “cold” (<800 °C). In contrast to the more abundant hot (and somewhat drier) granitoids, the low-T (and generally wetter) magmas may be initially relatively crystal-rich, reach their solidi well below the surface, and typically crystallize in the crust without allowing much liquid to escape. Low-T granites may, therefore, represent plutons with no (or little) volcanic equivalent.

More generally, however, Scaillet et al. (1998) have shown that average viscosities of most silicic magmas seem to converge around $\sim 10^{4.5}$ Pa s, irrespective of the level of emplacement and temperature — i.e., that there is no significant difference between volcanic and plutonic magmas. This observation suggests that the dominant parameter controlling the ultimate fate of a magma may not be viscosity, but rather external factors such as age and rheology of wall rocks, regional tectonic stresses, or overpressurization of the magma chamber by new influx of magma and heat. Whether a silicic magma stalls in the crust or rises to the surface therefore commonly depends not only on its magmatic characteristics, but also on these external factors. One could thus hypothesize that in the early and late stages of a magmatic province, when magma productivity is low and/or surrounding crust cold, the plutonic fate will be favored, while volcanic output will be larger during the paroxysmal phases of magmatism. This scenario (at least the late part of it) can be illustrated by the evolution of long-lived silicic magmatic fields (e.g., de Silva et al., 2006a; Grunder et al., 2007; de Silva and Gosnold, 2007-this volume); emplacement of shallow granitoids during the dying stages of a magmatic province (warm crust but low(er) magma productivity) is clearly seen when exposures allow it (e.g., Johnson et al.,

1989) or inferred by caldera resurgence (e.g., Lipman et al., 1978; Smith, 1979; Lipman, 1984).

3.3.4. *Do all volcanoes leave plutons behind?*

If it is likely that not all plutons erupt, one could also ask the question as to whether all volcanoes leave plutons behind. It appears likely that most mafic magmas transit rapidly through the crust (e.g., Reagan et al., 2003; Turner et al., 2003) and that some erupt without leaving much crystalline residue in the upper crust (however, see Dungan and Davidson, 2004 for evidence of crystalline residue in mafic systems). Such a scenario (volcanic rocks without plutonic counterparts) appears more doubtful for silicic magmas, but has been suggested for some. This situation could arise when silicic magmas (1) are nearly completely evacuated from their source during eruption (Glazner et al., 2004), (2) formed in the deep crust and ascend rapidly to the surface (e.g., Annen et al., 2006a), or (3) are the result of punctuated crustal melting events in the upper crust (e.g., Lewis-Kenedi et al., 2005). The absence of magma residence and differentiation in the upper crust would imply that shallow plutonic rocks would not be associated with the volcanic episodes. This case of volcanic activity without plutonic left-overs in the upper crust can certainly occur for small systems, but we would argue that most long-lived and productive magmatic provinces generate co-genetic volcanic and shallow plutonic rocks.

3.3.5. *The ultimate source of silicic magmas: deep crystallization and/or crustal melting?*

The rhyolite expulsion model from partly crystallized mushes (Bachmann and Bergantz, 2004) permits a genetic link between granitoids and rhyolites, but it does not resolve the problem of how silicic magmas (andesite, dacite) are formed in the first place. As far as the authors can tell, most igneous petrologists today seem to favor crystal-liquid differentiation as the fundamental process in magmatic differentiation, but despite a considerable amount of geochemical data on numerous systems, the main physical mechanism by which it occurs remains controversial. Does the dominant contribution to silicic magmas come from remelting pre-existing crust or from interstitial melts of basalts that never reached complete solidification?

An attractive way to explore this question further is to numerically simulate how basalts and surrounding rocks would behave (thermally and dynamically) when basalt is injected at a geologically reasonable rate at different levels in the crust. Although such models have been designed for nearly 30 yr (Yunker and Vogel, 1976; Wells, 1980), a recent collection of papers has

added a new level understanding (Petford and Gallagher, 2001; Annen and Sparks, 2002; Dufek and Bergantz, 2005). These studies have shown that both residual melt in the injected basalt and new melt from partial melting of host rocks can be present in the crust under certain conditions. The primary controls on silicic magma generation are (1) basalt flux, (2) crustal thickness, (3) composition (including water content) of surrounding crust, and (4) temperature and water content of intruding basalts. For nearly all reasonable basalt fluxes, a thin crust will preferentially lead to the generation of silicic magmas by interstitial melt escape from crystallizing basalts, as there is not enough heat to significantly melt the surrounding crust. In contrast, low basalt flux and thick crust will produce the largest ratio of crustal melt to basalt interstitial melt, and this ratio will decrease with time (see Fig. 12 of Dufek and Bergantz, 2005). Similarly, cool, wet basalts intruding infertile lower crust will produce a large proportion of interstitial melt with minimal crustal component, whereas hot, dry basalts will promote crustal melting and magmas rich in crustal component (in particular for fertile crust composition such as pelite, Annen and Sparks, 2002).

As demonstrated by these thermal models of silicic melt generation, both of the mechanisms discussed above ((1) remelting pre-existing crust and (2) crystallization of intruding basalts, providing heat and silicic interstitial melt) seem possible depending on the tectonic setting, structure (thermal and physical) of the crust and vigor of the local magma production rate in the mantle. An obvious first-order observation that suggests agreement between models and Nature is that silicic magmas are in minor quantities in areas lacking continental crust; although Iceland and island arcs such as Izu-Bonin and Kermadec can generate some dacitic–rhyolitic magmas (Gunnarsson et al., 1998; Tamura and Tatsumi, 2002; Smith et al., 2006), mid-ocean ridges produce only miniscule amounts of extremely fractionated silicic liquids and areas such as Hawaii will have no evolved magmas at all (Marsh, 1988). In areas where basalts intersect continental crust, two different trends arise:

1. Areas where magmatism is driven by mantle upwellings (hot spots). These magmatic provinces are typically bimodal and can produce fairly large amounts of silicic magmas, generally with hybrid characteristics between mantle and crust (e.g., Yellowstone, Hildreth et al., 1991; the Yemen Trap Series, Chazot and Bertrand, 1995). As basalts in these areas are typically dry, they will reach their solidi at fairly high temperature, but they are also relatively hot (≥ 1200 °C; e.g., Sobolev and Chaussidon, 1996; Putirka, 2005). In these conditions, some interstitial melt can be preserved in crystallizing basalts, and crustal melting is to be expected, especially when fertile lithologies are encountered by the mafic magmas. Because the melted pre-existing crust is highly variable (mantle upwellings can impinge the crust anywhere), resulting silicic melts will be highly variable.
2. Areas where magmatism is driven by subduction processes. Although some arc basalts are dry (e.g., Sisson and Bronto, 1998; Elkins-Tanton et al., 2001; Leeman et al., 2005), most of them are thought to be relatively wet and cooler than typical Mid Ocean Ridge and Ocean Island basalts (e.g., Sisson and Grove, 1993; Roggensack et al., 1997). Therefore, significant amounts of residual melts from crystallizing basalts should be preserved, especially if the crust is thin (i.e., not favoring crustal melting; Dufek and Bergantz, 2005). This situation is exemplified by the Aegean arc, where the conjunction of arc (wet, cool) basalts, thin crust and low mantle productivity leads to silicic melts dominated by differentiation of mafic parents (Di Paola, 1974; Briquieu et al., 1986; Francalanci et al., 1995; Druitt et al., 1999), even for rhyolitic magma bodies >60 km³, like the Kos Plateau Tuff (Bachmann et al., 2007a). In contrast, a thick crust and/or a high mantle power input (e.g., Central Andes, San Juan volcanic field) will lead to a much higher proportion of crustal melt in the newly-produced silicic magmas. As demonstrated by the dynamic 2-D simulations of Dufek and Bergantz (2005), both melts have the tendency to mix in these conditions, leading to the MASH process (Melting, Assimilation, Storage, Homogenisation, as proposed by Hildreth and Moorbath, 1988). The Taupo Volcanic Zone appears to lie in an interesting middle ground; magmatic evolution seems to be dominated by AFC (Graham et al., 1995) despite the fairly thin crust (highly extensional environment). However, as a consequence of high basalt flux the crustal heat transfer in the Taupo area is the highest known for any arc environment (Hochstein, 1995), and this intense mantle power input is probably what allows some crustal melting (inducing AFC) to occur.

3.3.6. “Flare-ups” as the drive for batholith construction?

Large silicic volcanic fields (10,000's of km²; 1000s–10,000's of km³) are constructed during ignimbrite “flare-ups”: intense periods of volcanism dominated by eruptions of ignimbrites with volumes in excess of 1000 km³ (e.g., Coney, 1978). Following workers such as Hildreth (1981), Lipman (1984), and our discussion here, these ignimbrite flare-ups must be accompanied by significant (and more voluminous) plutonic activity at depth. Coining the term

“magmatic flare-up” for the integrated plutonic and volcanic episode, Ducea (2001) suggests that cordilleran batholiths are constructed during these events. The eruptive histories and spatiotemporal development of ignimbrite flare-ups may therefore provide valuable insight into the development of cordilleran batholiths.

De Silva et al. (2006a) and de Silva and Gosnold (2007-this volume) draw attention to the broad similarity of volume-time patterns of several ignimbrite flare-ups. These are typically episodic and show a three stage evolution:

1. An early “waxing” stage characterized by eruptions of relatively small volume and low magma production rates with wide spatial distribution.
2. A catastrophic stage with successively larger eruptions and higher magma production rates that may be more focused.
3. A late “waning” stage of small eruptions.

This pattern is interpreted as reflecting the progressive thermal and mechanical maturation of the crustal column due to advection of heat from an elevated (above background) thermal input from the mantle (high mantle power) into the crust. This is most likely in the form of a mantle upwelling (delamination is invoked as the trigger for this by Ducea (2001), but other mechanisms (such as slab rupture, Kay and Mahlborg Kay, 1993) have been invoked). The transient nature and intensity of flare-ups suggest that the mantle power input itself is a transient pulse that sets up a positive feedback between crustal temperature, melt production, mechanical state, and intrusive volumes. As discussed above, the addition of heat from the intrusion and crystallization of mafic magma will result in silicic melt, with the melt production rate being a function of heat flow into the lower crustal protolith (Harris et al., 2000; Petford and Gallagher, 2001; Dufek and Bergantz, 2005). Thus, to a first order, felsic melt production can be seen as a function of the intrusion rate (flux) of basaltic magma into the crust. The pulse of mantle heat that drives flare-ups can be assumed to have a broadly Gaussian topology with the initial intrusion rate being low, rising to a peak, and then diminishing back to the background. Heat flow into the protolith will therefore rise and fall accordingly, resulting in a concomitant pattern in silicic melt production; an initial period of relatively low melt production (the waxing stage) that rises to a peak (the catastrophic stage) and then diminishes as the pulse dies off (the waning stage). The temperature profile of the crust will also change with time as the conductive geotherm is progressively elevated in response to the heat flow into the crust, melting, melt extraction and intrusion of silicic melts to higher levels resulting in elevation of the brittle–ductile

transition (de Silva et al., 2006a,b; de Silva and Gosnold, 2007-this volume). The net result is that early silicic melt accumulates in upper crustal environments that are relatively cold and brittle, favoring eruption over accumulation, while later melt accumulations are in a warmer more ductile upper crust that favors accumulation over eruption (e.g. Jellinek and DePaolo, 2003). As this system is tapped periodically, the pattern of increasing intensity and volumes of eruptions is produced. The pattern of diffuse low-volume volcanism early with later more focused larger-volume volcanism suggests that the conditions for accumulation of large volumes of silicic magma focus towards the center of the field with time. This may ultimately reflect the spatial pattern of intensity of the mantle power input as the rate of silicic magma production and intrusion is likely to be focused above the region of highest mantle heat flux.

The implications for cordilleran batholiths is that they should be composite with bodies of increasing volume developing with time that may be spatially zoned. A possible example of this progression can be found in the 85–93 Ma, 1200 km² Tuolumne Intrusive Suite, where production rate (based on preserved km²/myr) appears to have started slowly, peaked after 90 Ma, and dwindled by 85 Ma, with compositions progressively more felsic with time (Coleman et al., 2004, 2005).

4. A possible (and partial) model for crustal magmatism

Based on reasoning that has been presented so far, we assert that plutons and volcanoes provide information about magmatic systems that can be integrated. Plutons are composite, terminal images of magmatic systems, recording multiple intrusive events and late processes (some below the solidus), whereas volcanic rocks offer instantaneous snapshots of magmatic systems, mostly representing the peak of magmatic activity, and revealing the evolution of melt expelled from the plutonic realm. We proceed below to summarize our view of the main processes driving crustal magmatism (excluding environments of continental collision, where magma genesis is likely to be fundamentally different).

1. Basalt, the main driver of crustal magmatism (e.g., Hildreth, 1981), is produced by melting of the mantle, either due to decompression (e.g., Kushiro, 2001) or addition of fluids (Davies and Stevenson, 1992; Ulmer, 2001; Parman and Grove, 2004).
2. In areas where continental crust is absent (divergent margins, hot spots in ocean basins, young islands arcs), these basalts will rapidly move upward through the crust, and typically erupt with a fairly primitive

composition. Not much differentiation is induced, and silicic compositions are minor.

3. When continental crust is present, interaction of mafic magmas and non-basaltic crust most commonly leads to a significant production of silicic magmas (large mafic intrusions trapped in old, cold crust may be an exception to this rule). In both intraplate and arc settings, silicic magmas typically have hybrid characteristics (including both mantle and crustal components; e.g., Lipman et al., 1978; DePaolo, 1981; Farmer and DePaolo, 1983; Pitcher et al., 1985; Hildreth et al., 1991; DePaolo et al., 1992; Chazot and Bertrand, 1995; Knesel and Davidson, 1997). The total volume of silicic melt production will strongly depend on mantle power input; if power input is large (e.g., Yellowstone, Tertiary Yemen Province for intraplate setting and mature continental arcs for subduction zones), large quantities of silicic magmas can be generated. This is particularly obvious in cases of intense transient pulses of mantle power input, referred to as flare-ups (de Silva, 1989; Ducea, 2001); not only do large amounts of basalt intrude the crust during these events, yielding significant volumes of fractionated magma, but compressional forces acting at convergent margins typically impede ascent of primitive magmas to the surface (except as melt inclusions preserved in mafic crystal phases, Anderson, 1982), allowing basalts to contribute most of their thermal energy to crustal anatexis. These intense flare-up episodes lead to major crustal building events, with the emplacement of voluminous batholiths with intermediate characteristics between crust and mantle (Hildreth and Moorbath, 1988; Pitcher, 1993).
4. To account for the fairly evolved composition of the bulk crust (Rudnick, 1995), mafic plutonic residues must be recycled back to the mantle by delamination (Arndt and Goldstein, 1989; Kay and Mahlburg Kay, 1993; Rudnick, 1995; Jull and Kelemen, 2001; Dufek and Bergantz, 2005).
5. Buoyant silicic magmas will rise from their deep source regions into the upper crust and crystallize to a mush state in the relatively cold conditions that prevail at shallow depth. Once they reach a mush state, cooling by conduction will be slow (Koyaguchi and Kaneko, 1999), and they will be periodically reactivated by new input of hot, mafic magmas (Pallister et al., 1992; Murphy et al., 2000; Bachmann et al., 2002), leaving time for rhyolitic melt expulsion by microsettling and compaction (Bachmann and Bergantz, 2004). Evolved, crystal-poor rhyolites can then be periodically erupted, leaving large plutons with *subtle* cumulate affinities (as only a small fraction of the

interstitial rhyolitic melt in these mushes is expelled). Other magmas may solidify entirely without erupting, either due to external factors (deep storage, compressional environment, cold surroundings) or internal characteristics (low temperature, high H₂O content).

6. The magmatic system that develops in response to mantle power input advects heat through the whole crustal column. The evolving thermal and mechanical state of the crust during increasing mantle power input results in a positive feedback that results in accelerating silicic magma production and intrusion rate. This is manifested at the surface by increasing volumes (and progressive chemical evolution) of successive pulses of eruptions that presumably find plutonic roots as composite batholiths. The development of large cordilleran batholiths may be paralleled by surface ignimbrite flare-ups that represent the thermal and mechanical maturation of the crust in response to a transient increase in mantle power input (de Silva et al., 2006a,b).

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