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# A new paradigm for granite generation

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ABSTRACT: Ideas about granite generation have evolved considerably during the past two decades. The present paper lists the ideas which were accepted and later modified concerning the processes acting during the four stages of granite generation: melting, melt segregation and ascent, and emplacement. The active role of the mantle constitutes a fifth stage.

Fluid-assisted melting, deduced from metamorphic observations, was used to explain granite and granulite formation. Water seepage into meta-sedimentary rocks can produce granitic melt by decreasing melting temperature. $CO_2$ released by the mantle helps to transform rocks into granulites. However, dehydration melting is now considered to be the origin of most granitic melts, as confirmed by experimental melting. Hydrous minerals are involved, beginning with muscovites, followed by biotite at higher temperatures. At even deeper conditions, hornblende dehydration melting leads to calc-alkaline magmas.

Melt segregation was first attributed to compaction and gravity forces caused by the density contrast between melt and its matrix. This was found insufficient for magma segregation in the continental crust because magmas were transposed from mantle conditions (decompression melting) to crustal conditions (dehydration melting). Rheology of two-phase materials requires that melt segregation is discontinuous in time, occurring in successive bursts. Analogue and numerical models confirm the discontinuous melt segregation. Compaction and shear localisation interact non-linearly, so that melt segregates into tiny conduits. Melt segregation occurs at a low degree of melting.

Global diapiric ascent and fractional crystallisation in large convective batholiths have also been shown to be inadequate and at least partly erroneous. Diapiric ascent cannot overcome the crustal brittle-ductile transition. Fracture-induced ascent influences the neutral buoyancy level at which ascent should stop but does not. Non-random orientation of magma feeders within the ambient stress field indicates that deformation controls magma ascent.

Detailed gravity and structural analyses indicate that granite plutons are built from several magma injections, each of small size and with evolving chemical composition. Detailed mapping of the contact between successive magma batches documents either continuous feeding, leading to normal petrographic zoning, or over periods separated in time, commonly leading to reverse zoning. The local deformation field controls magma emplacement and influences the shape of plutons.

A typical source for granite magmas involves three components from the mantle, lower and intermediate crusts. The role of the mantle in driving and controlling essential crustal processes appears necessary in providing stress and heat, as well as specific episodes of time for granite generation. These mechanisms constitute a new paradigm for granite generation.

KEY WORDS: Ascent, emplacement, granite emplacement, mantle, melting, segregation.

The Earth is a thermal machine that progressively loses its energy with time. The heat generated by radiogenic elements in the core, mantle and crust is progressively lost through the convective motion of matter that drives plate tectonics (Bercovici 2003). It generates basaltic magma that adds material to the crust at Mid Ocean Ridges. In the continental crust, energy is lost through material transport during granitic intrusions and through deformation that manifests itself as seismic activity. Granitic magma formation is also a major way to transfer elements, such as Rb and Ba, but also Th and U, which are incompatible in mantle minerals, to the crust. It leads to a chemical differentiation of the continental crust with time.

Specific studies conducted on granitic bodies have documented internal structures, revealing the role of deformation. Joint structural and geophysical approaches identified patterns in granitic plutons, such as root zones from where magma was derived from the lower crust. At the same time, experimental studies on magma formation led to renewed ideas about crustal-derived magmas. They resulted from two different paths. On the one hand, new analytical techniques allowed more precise measurements involving a smaller sample volume. On the other, new concepts also modified scientists' interpretations of field observations (Castro *et al*. 1999). The renewed data sets led researchers to consider four stages of granite generation: melting, melt segregation, ascent and emplacement (Petford *et al*. 1997). Each stage has its own time and length scales, with an associated physical process. The four stages are embedded within a global context, which is mantle controlled.

The present paper suggests a new paradigm for granite generation. It is based on the author's own experience, gained through studies conducted on the several stages of granite generation. It also involves data and ideas which needed to be collected and put together in a synthetic form. It is organised in five major sections, according to the four stages and the global context. Definitions of the terms used throughout the paper relate to felsic igneous bodies. A granitic pluton is considered as a unit for granitic intrusion. Batholiths, such as the Sierra Nevada Batholith, are consequently built from several plutons. A single unit is built from several batches of magma with specific mineral facies or chemical compositions.

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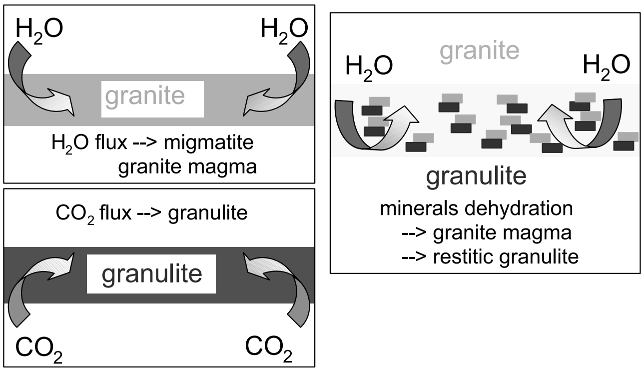


**Figure 1** Melting as viewed in the old (left) and new (right) paradigms: The old paradigm considered a generalised metamorphism, resulting in granitic melt generation by a flux of water and granulite production by a flux of $CO_2$. In the new paradigm, water liberated from hydrous minerals during dehydration melting is immediately reincorporated to granitic melts, leading to residual granulites.

## 1. Melting

### 1.1. The old paradigm

Melting is the initial stage of magma formation. It depends on the available heat from either a thickened crust and/or from the underlying mantle. Older models of granitic magma production, especially for crustal-derived magmas, were inspired from a generalised concept of metamorphism (Fig. 1). Crustal rocks are progressively buried, increasing the ambient temperature. A flux of $H_2O$ enhances melting by decreasing the liquidus temperature (Wyllie 1977). The parallel model of a $CO_2$ flux originated from the mantle results in granulite formation in the lower crust (Touret 1977).

Pioneering experiments (Tuttle & Bowen 1958) attributed granitic magma to $H_2O$-saturated or weakly under-saturated melting. However, experiments could not document water exceeding 2% in weight in granitic magmas (Maaløe & Wyllie 1975), weakening the idea of a water flux control. Andesitic magma, which forms a large part of subduction-related magmatism, could not derive from mantle peridotite because of the limited amount of water. It must derive from intermediate liquids formed from subducted sediments and mantle peridotite (Wyllie *et al.* 1976). In the continental crust, granitic magmas could be generated under regional metamorphism, provided that heat was there, resulting in $H_2O$-under-saturated granitic melt and migmatites (Wyllie *et al.* 1976). Hence, the model requires elevated temperature. A granitic magma with 2% water needs a temperature from 800 to 1000°C, under a pressure of 500–1000 MPa. Under such conditions, the continental crust can produce water-under-saturated liquids of granitic composition and granodioritic magma at higher temperatures (Wyllie 1977).

### 1.2. The new paradigm

In contradiction to a water-saturated model of melting, observations document that the lower crust has no free water, no hydrous minerals and too low a permeability (Yardley & Valley 1997). During experimental melting, dehydration breakdown of hydrous minerals (micas and amphiboles) results in a melt with granitic composition (Thompson 1982; Clemens 1990). The reactions liberate water that is immediately reincorporated to produce magma (Fig. 1).

Muscovite is the first mineral to melt above 700°C and melting is completed below 800°C in low-pressure conditions. Increasing pressure shifts the reaction to higher temperatures (Patiño Douce & Beard 1995). In both conditions, melt productivity remains weak, around 10–15%. The reaction also involves plagioclase and quartz, both in similar proportions and equivalent to the muscovite content. Biotite dehydration occurs between 820 and 950°C, with a weak pressure effect. The reaction involves biotite, as well as plagioclase and quartz in similar proportions, but each about three-quarters of the amount of biotite (Harris *et al.* 1995). The maximum amount of melt is about three-quarters of the initial volume of biotite in the source. Melts are peraluminous in composition with a K/Na ratio commonly greater than one (Whitney 1990).

Progressive dehydration melting of amphiboles starts at about 1000°C and 800 MPa (Rapp & Watson 1995). Reaction temperature slightly increases with pressure (Wyllie 1977). In vapour-absent conditions, hornblende and plagioclase react to form clinopyroxene and garnet plus melt. The melt is tonalitic in composition (Percival 1983) with K/Na commonly lower than one. At higher temperatures, amphiboles continue to react with plagioclase and clinopyroxene, producing more melt, up to 45%. At intermediate pressure, but lower than 1·0 GPa, the reaction consumes only hornblende and quartz, producing plagioclase and pyroxenes (Harlov 2002). In this case, the amount of hornblende should be greater than quartz (Patiño Douce & Beard 1995).

Mafic magmas derived from the mantle are necessary to generate the high temperature for hornblende dehydration melting. They also participate in granite formation in the continental crust. Mafic enclaves are ubiquitous in subduction-related granitic magmas, documenting several types of magma mixing (Didier & Barbarin 1991). Mixing takes place either deep in the crust, during the ascent or even after emplacement by a continuous feeding of mafic dykes (Collins *et al.* 2000; Weinberg *et al.* 2001). Three sources of magma have been suggested, involving the mantle, the intermediate crust and the lower crust (Collins 1996; Clarke *et al.* 2000). Balancing between the three sources may reproduce all types of granitic magmas. Trace element chemistry and isotopic signatures document the large mass transfer that has occurred between mafic and felsic magmas (Holden *et al.* 1991).

The dehydration reactions also imply the production of a restitic material of granulitic composition (Thompson 1982). A global estimate is that one volume of granitic magma and three volumes of granulitic restite result from the melting of four volumes of amphibolitic compositions (Wedepohl *et al.* 1991). Thus, dehydration melting has the double advantage of explaining granitic magma formation and crustal differentiation without the intervention of any external fluid flux.

## 2. Segregation

Two major points have long hindered comprehension of melt segregation. One was the requirement of a large volume of melt before segregation could take place. The second was the very incomplete understanding of the rheology of partially molten rocks (PMRs).

### 2.1. The old paradigm

Melting starts at mineral triple or quadruple junctions, where the activation energy is lowest (Jurewicz & Watson 1985). Once melt films are connected throughout the system, segregation starts, leading to the loss of cohesion of the source for about 20–30% melt. The commonly accepted model of melt segregation required a huge volume (10–20 km thick) of crust with highly variable melt fraction and a force to concentrate melts. Gravity-driven compaction is the preferred model (Fig. 2) of such force (Miller *et al.* 1988), in contrast to extensional fracturing (Shaw 1980).

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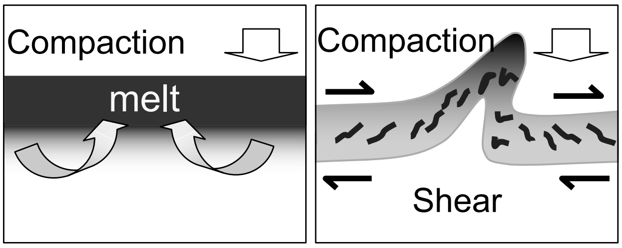


**Figure 2** Melt segregation: Within the old paradigm (left), a layer (in grey) with large melt fraction segregates by compaction (arrow). Melt can escape from its source region when overcoming a critical percentage. In the new paradigm (right), compaction and shear instabilities develop and interact, resulting in small melt packets (in grey) periodically escaping the system. The figure is adapted from experiments (Barraud *et al.* 2001).

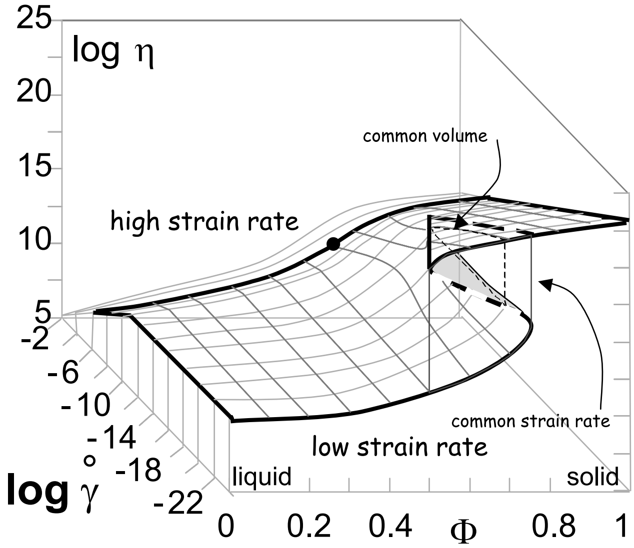


**Figure 3** Rheology of partially molten rocks: The three-dimensional diagram plots viscosity (η) as a function of strain rate (γ°) and a percentage of the solid fraction (φ). For high strain rate (e.g. tectonic strains, the viscosity contrast is the lowest and the transition is continuous between the two end-members. In contrast, at low strain rates, the curve adopts a cusp shape and three values of viscosity exist for a given value of solid fraction. Two are metastable, and the intermediate value does not exist in nature. Different paths exist, depending on whether strain develops under constant strain rate, constant volume or constant stress.

The binary description of PMR rheology implied a critical value of melt percentage at which the system shifts from a solid-like to a liquid-like rheology. A rheological critical melt percentage (RCMP) marks the transition between the two behaviours (Arzi 1978). The binary description was later transposed to the deformation of rock samples bearing a small amount of melt (Van der Molen & Paterson 1979). In such a binary model, melting is symmetric to crystallisation. The RCMP value has been fixed at 25–30% melt, corresponding to the range observed in migmatites (Arzi 1978; Miller *et al.* 1988).

Numerical models attributed magma segregation to compaction (McKenzie 1984; Brown *et al.* 1995). They were directly transposed from mantle conditions to the continental crust. Nevertheless, they resulted in compaction lengths of the order of 100 m (McKenzie 1984), which do not match field observations on migmatites.

## 2.2. The new paradigm

Detailed observations on melt segregation in anatectic rocks (Gleason *et al.* 1999; Sawyer 2000; Harlov 2002) or in crystallising magmas (Bouchez *et al.* 1992; Sawyer 2001) document the difficulty of finding adequate rocks in the temperature range corresponding to that of melting. Sawyer (2000) examined rocks in the granulitic and amphibolitic facies, as well as meta-pelites during incipient melting. Local replacement of mineral by melt (Sawyer 2000; Harlov 2002) and later connection of melt along grain boundaries document anisotropic segregation. New optical techniques able to investigate the nature of grain boundaries (Dimanov *et al.* 1999; Kohlstedt *et al.* 2000) document the presence of amorphous material at grain boundaries. They suggest that a melt film can be traced even for very low melt fractions (Wirth & Frantz 2000).

Those observations suggest new interpretations on the rheology of PMR. Obviously, a binary model is false, and melting is not symmetric to crystallisation (Burg & Vigneresse 2002). Hence, latent heat makes the liquidus temperature different from that of the solidus. In the new model, three stages are considered (Fig. 3): two end-members and a transitional stage bracketed by two thresholds (Vigneresse *et al.* 1996). The later are the liquid percolation threshold (LPT) and the melt escape threshold (MET) for melting. Identically, a rigidity percolation threshold (RPT) and particle locking threshold (PLT) exist for crystallisation. The individual threshold values are not dual to each other during melting and crystallisation. In the transitional stage, non-linear feedback loops develop which have a positive effect during melting, but a negative one during crystallisation (Burg & Vigneresse 2002). Instabilities result from the heterogeneous distribution of melt with temperature and place. The transitional stage greatly depends on strain rate. Whereas the melt presents a constant viscosity, the matrix follows a power law rheology. Decreasing strain rate exacerbates the viscosity contrast, which lead to two types of reaction (Fig. 3). At a high strain rate, such as during tectonics, the viscosity contrast between the two phases is the lowest. The transition is progressive between the two end-members and an effective viscosity can be assigned, whatever the melt content. Conversely, at lower strain rates, the contrast between the two phases is the greatest. Three situations are observed, depending on the melt content. At low and high melt content, the system adopts the rheology of the respective end-member. In between the thresholds, a complex situation develops during which the two phases, i.e. matrix and melt, coexist. Jumps continuously occur from the value of one phase to that of the other, precluding assignation of a bulk viscosity value to the system. Strain preferentially partitions in the melt (Vigneresse & Tikoff 1999), which segregates discontinuously, yielding successive bursts of melt. The effects differ on whether they develop at constant strain rate, constant stress or constant amount of melt (Fig. 3) (Vigneresse & Burg 2004).

Analogue models of melt segregation (Barraud *et al.* 2001; Bons *et al.* 2001) demonstrate segregation of discrete magma packets (Fig. 2). No quantitative interpretation is presently given for the discontinuous melt motion in those experiments. Numerical models using classical descriptions (Rabinowicz *et al.* 2001; Ricard *et al.* 2001) or using cellular automata (Vigneresse & Burg 2000) also document instabilities. They all consider low melt concentration located at grain boundaries (5–10% melt), which results in an equivalent porous flow. For granitic melts which segregate in crustal conditions, compaction is not efficient enough and horizontal shear must be added (Rabinowicz & Vigneresse 2004). A competition starts between instabilities initiated by compaction and by shear. They cooperate and lead to a compaction length that reduces to decimetres in scale, analogous to what is observed in migmatites (Rabinowicz & Vigneresse 2004). Melt extraction is discontinuous and successive cycles can develop during which a limited percentage (5%) of melt is expelled. The heat

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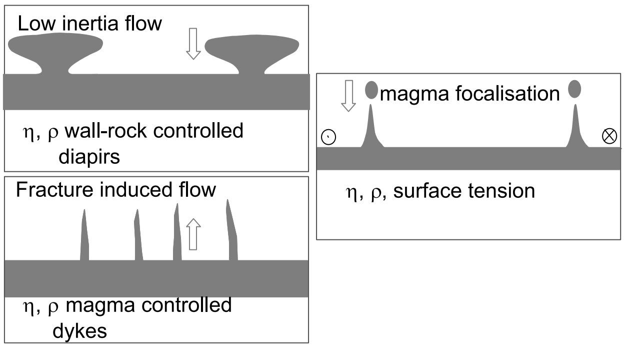


**Figure 4** Magma ascent toward the upper crust: In the old paradigm (left), the ascent develops through low inertia flow (diapirs) and fracture-induced flow (dykes). The viscosity and density values of the wall rock and of the magma, respectively, control the ascent in the two processes. In the new paradigm, rheological instabilities develop under compaction and shear, leading to individual bursts of melt and magma focalisation.

available for additional melting controls the cycle frequency. However, progressive melt removal also stiffens the crust, limiting the total number of cycles. The total amount of melt that can be extracted remains below 25% in volume (Rabinowicz & Vigneresse 2004).

## 3. Ascent

### 3.1. The old paradigm

Granitic magmas ascend from the lower crust where they form by melting to the middle crust where they build large plutons. In the old paradigm, there were two basic models (Fig. 4) of magma ascent: inertia flow or fracture propagation (Vigneresse & Clemens 2000). In both cases, a nearly liquid magma ascends to coalesce and form large plutons. Both models assume a neutral buoyancy level at which the density contrast between magma and its host rocks vanishes, stopping the ascent.

In diapiric ascent or Rayleigh-Taylor instability, a lighter and less viscous layer forms a dome, followed by the rise of a characteristic inverted-teardrop-shaped body. The viscosity contrast between magma and its host rocks drives the instability. Combined analogue and numerical experiments refined inertia flow models. A Stokes flow used either a rigid (Cruden 1988) or a soft sphere (Schmelling *et al.* 1988). Rayleigh-Taylor instabilities assumed either viscosity contrast (Weinberg 1996) or not (Weijemars 1986). Models in which both the ascending body and its host are viscous (Mahon *et al.* 1988) also examined the thermal effects. Most recent models consider a viscous intrusion into a ductile and/or brittle crust (Barnichon *et al.* 1999). Those models restrict diapirism to the lower plastic deforming crust. Former models had already demonstrated, though sometimes in a cryptic form, that diapirism is not compatible with a brittle upper crust (Mahon *et al.* 1988; Weinberg & Podladchickov 1994). In those models, a less dense and less viscous sphere, typically 5 km in radius, starts from a depth of about 30–40 km, and slows down its ascent at mid-crustal depth before stopping at a depth of between 14 and 21 km (see Vigneresse & Clemens 2000, fig. 1).

Fracture propagation, or dyking, represents an alternative for magma ascent. The magma viscosity drives the forces at the tip of a fracture, leading to the upward propagation of the fracture. The internal magma pressure must balance the elastic response of the rock. The first models (Weertman 1971) used an elastic crack with an internal pressure gradient caused by magma (Pollard & Muller 1976) or viscous liquid fill (Lister & Kerr 1991; Rubin 1997).

At present, magma ascent as a whole body (i.e. diapirism) is commonly considered to be unrealistic. Even salt diapirism is driven by tectonics, as evidenced by precise measurements of extrusion rate (Talbot *et al.* 2000). For magma, dyking is actually the preferred model. Nevertheless, it faces numerous problems relating the diameter of the feeding pipes to magma viscosity (Petford *et al.* 1993; Clemens *et al.* 1997). Small zones observed from gravity data inversion and interpreted as feeders for the pluton (Améglio *et al.* 1997) argue for dyking. Nevertheless, gravity data precision limits the detection of these feeders to those about 300 m in width at about 5 km in depth. Larger feeders should be detectable from gravity measurements, which obviously is not the case.

A neutral buoyancy level is not supported by field observations. A single pluton shows various types of magma, which all present with a varying composition and, hence, density. The apparent occurrence of magmas with different densities at the same level contradicts the existence of a neutral buoyancy level. Gravity measurements performed over granitic intrusion document negative or positive anomalies between the massif and its surroundings. The anomalies indicate that successive magma pulses stopped at a similar level, whatever their density (Vigneresse & Clemens 2000).

### 3.2. The new paradigm

In the new paradigm, deformation is intimately linked to granitic magma ascent (Fig. 4), providing the required forces to overcome the strength barrier at the ductile/brittle transition (Vigneresse 1995b). Nevertheless, magma ascent is not related to a specific type of deformation such as extensional tectonics. Numerous examples document granites associated with strike slip conditions or compressional thrusts (Schmidt *et al.* 1990; Kalakay *et al.* 2001). The best way to extract a melt from its matrix is by compression and shear (Vigneresse & Burg 2000). However, by itself, the ascent must be in a locally extensional near-field setting because of the lesser magnitude of the tectonic stresses magma must overcome (Vigneresse 1995a).

Complex relationships exist between granitic intrusion and faulting (Paterson & Schmidt 1999), excluding a genetic link between intrusion and faulting. Intrusions are randomly distributed with respect to faults, which precludes that faults could channel magma during the ascent (Paterson & Schmidt 1999). From gravity data inversion, the root zones of a pluton are not connected to adjacent faults, leading to a similar conclusion (Vigneresse 1995a). However, the shape of the walls in a pluton are controlled by adjacent faulting as with the Mortagne Massif, France (Guineberteau *et al.* 1987).

Differences in ascent are found in analogue models depending on whether they develop within a homogeneous crust or not. Laccoliths form during intrusion into a homogeneous crust (Roman-Berdiel *et al.* 1995), while wedge-shaped massifs develop when intruding a heterogeneous crust (Roman-Berdiel 1999). When the deformation pattern is complex (transtension or transpression), intrinsic non-linear relationships form loops between magma production and tectonics, driving mechanisms to complex situations (Saint Blanquat *et al.* 1998).

## 4. Emplacement

### 4.1. The old paradigm

In the old paradigm, the ascent of a large volume of magma as a whole requires a large space to be created that would be later filled with magma. The so-called 'room problem' has been a

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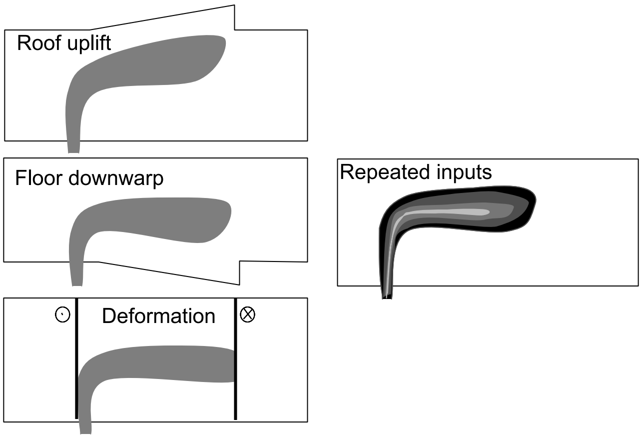


**Figure 5** Granitic magma emplacement in the intermediate crust: In the old paradigm (left), the so called room problem is solved by roof lifting (top), floor downwarping (centre) or deformation-assisted opening (bottom). In the new paradigm, repeated inputs of magma leave time for the surrounding rocks to relax and to accommodate magma emplacement.

continuous source of disputes during the past few decades (Read 1957; Paterson & Vernon 1995; Hutton 1997). Several mechanisms have been suggested which all contribute to create space for intrusion to take place. Wall assimilation has been advanced on the basis of isotopic and trace elements (Taylor 1980; De Paolo 1981). Large blocks of surrounding rocks are commonly observed which have fallen into the magma and present various degrees of assimilation. Room accommodation could also be realised by either uplifting (Fig. 5) the overlying rocks (Kerr & Pollard 1998) or depressing the floor (Fig. 5) to compensate magma ascent (Cruden 1998). Tectonic-aided fractures (Fig. 5) can also facilitate magma emplacement by inducing pull-apart-like structures which are rapidly filled by magma (Hutton 1982; Tikoff & Teyssier 1992).

In all those emplacement mechanisms, granitic magma acts passively. Nevertheless, it remains difficult to conceive how magma in a quasi-liquid state can develop enough strength to sustain those stresses induced by surrounding crustal rocks.

## 4.2. The new paradigm

A significant advance was initiated when structures were identified within granitic plutons (see review in Bouchez *et al.* 1997). Documenting internal structures, together with the determination of the three-dimensional (3D) shape of the massif by gravity data inversion, revealed that granite plutons recorded strain history (Guillet *et al.* 1985). Magma registers the very late strain increment (Benn 1994; Paterson *et al.* 1998). Present structural analyses commonly involve the anisotropy of magnetic susceptibility (AMS). The bulk shape of a granitic massif is obtained after mapping gravity anomalies, i.e. density changes caused by facies variations, over a pluton. Full 3D inversion is recommended (Vigneresse 1990). The shape of a massif at depth can later be correlated with its internal structures. Zones where the depth of the floor becomes deeper, provided they also correlate with steeply plunging magmatic lineations, are interpreted to be reflecting magma feeders (Vigneresse & Bouchez 1997).

Field observations, especially structures at the contact with wall rock, reveal that a forceful intrusion, with a major vertical movement of the magma, is often unlikely in granite plutons. In many cases, intrusions contain horizontal foliation and lineations close to, but orthogonal to, the walls of the granitic body. No flattening is observed within the granitic body when approaching the contact. By contrast, the wall rocks may show flattening approaching the granite (Brun *et al.* 1990), showing that magma inflated, pushing the host rocks during a large-scale horizontal movement of magma.

Granitic plutons have been classed by their shape into tabular- and wedge-shaped bodies (Améglio & Vigneresse 1999). A majority of batholiths belongs to the first category, whereas the second includes special cases often associated with tectonically controlled emplacement (Améglio *et al.* 1997). Tabular intrusions show a limited thickness ($5 \pm 2$ km in average), with their floors gently dipping towards a magma feeder, marked by a pronounced deepening. Within the massif, magma flows and regularly fills its room within the upper crust. Other examples describe magma as being intrusive within concordant structures and sub-parallel to host rocks anisotropy, for which the brittle-ductile transition is the best example (Vigneresse 1995a). In the second case, walls steeply plunge inward and the floor is deep (about 10 km at present) and flat, without any evidence of a localised deeper root zone. They fit models of tectonically assisted emplacement, the walls being often associated with shear zones.

The regional stress field developed by magma intrusion under partial crystallisation (Parsons *et al.* 1992; Vigneresse *et al.* 1999) explains the geometry of some granitic intrusion emplaced at high levels in the crust (Hogan *et al.* 1998). Magma intrudes along planes perpendicular to the minor stress component ($\sigma_3$), which are usually vertical planes in extensional or strike-slip environments. They represent planes of easy opening (Ramsay & Huber 1983). During crystallisation, the framework of touching crystals rapidly reaches 50% concentration which guarantees the rigidity of the body (Guyon *et al.* 1990). Above, the magma is able to sustain and to transmit normal stresses. It locally alters the nearby stress field (Parsons *et al.* 1992; Vigneresse *et al.* 1999), increasing the magnitude of the two minor stress components ($\sigma_3$ and $\sigma_2$). Two minor stress components may switch together when they have similar amplitudes. Step by step, opening along the initial vertical plane slows down and the stress pattern switches to another configuration (Vigneresse *et al.* 1999).

In a purely extensional stress field, the horizontal stress components are about one-quarter of the vertical load. Successive switches between the horizontal components may lead to the formation of vertical and cylindrical plutons (Fig. 6). The latter builds up from successive vertical sheets oriented along planes which rotate according to successive stress exchanges. This is the case for many 'anorogenic' granites, such as those in Namibia (Bowden & Turner 1974). In the case of transtension, switch occurs from strike-slip to extension, whereas the horizontal stress components switch from strike-slip condition to compression in cases of transpression. Elongated thin vertical sheets of magma result in transtensional situations, as observed in Southern Chile (Wilson & Grocott 1999). Batholiths built during transpression are wider, oriented along the shear direction (Fig. 6). They may develop intrinsic shear zones within the granitic magma; for instance, as observed in the Sierra Nevada (Saint Blanquat *et al.* 1998). When being emplaced during compression, massifs adopt a thin plate horizontal shape (Fig. 6).

Petrographic zoning in granitic plutons is a function of the rate at which successive batches of magma are emplaced (Hecht *et al.* 1997; Hecht & Vigneresse 1999). In the case of rapidly delivered magma batches, the earlier magma has no time to solidify. The later pulses of magma are emplaced into the core, pushing aside the former batch. If those batches consist of magma fractionated at depth, the resulting zonation of the pluton is normal, the latter material intrusive towards the core being more felsic. In contrast, when the rate of emplacement is slow, magma has time to cool down, reaching about 50% crystals or more. It becomes rigid before a new

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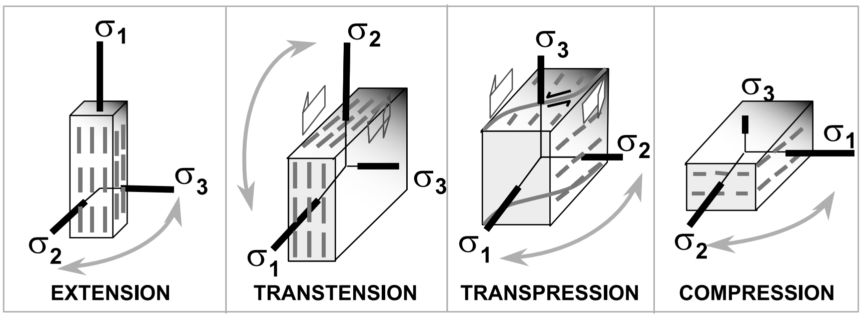


**Figure 6** Shape of granitic plutons depending on the switches between the stress components ($\sigma_1 \geq \sigma_2 \geq \sigma_3$) occurring in extensional, transtension, transpression and compressional environments (see text).

magma batch comes in. The later batch cannot develop enough force to overcome the yield strength of the now solidified magma. Thus, it is forced to intrude beside the earlier intrusion. A reverse zoning results, with the older magma in the core, and the younger, more evolved magma at the rims on both sides.

A granitic pluton is built from a sequence of successive intrusions (Fig. 5) with variously evolved magma (Vigneresse & Bouchez 1997; Petford *et al.* 2000). It has been verified by numerous examples on the scale of the large batholiths which form the Sierra Nevada (McNulty *et al.* 2000) and on a smaller scale in southern Spain (Aranguren *et al.* 1997). At an even smaller scale, in the Himalayan granites, magma heterogeneity is documented by the large variation in isotopic ratios (Deniel *et al.* 1987) and by the variable degree of melting shown by the distribution of rare-earth elements in zircons and monazites (Brouand *et al.* 1990).

In conclusion, granite magmatism takes place because host rocks allow emplacement within the crust. Discontinuous intrusions of magma leave time for the crust to relax the stress induced by magma intrusion, avoiding the room problem. The competition between the rate of magma emplacement and that of its cooling to reach rigidity determines the type of zoning in a granitic pluton. Both emplacement and zoning are dynamic.

## 5. The necessary mantle implication

### 5.1. The old paradigm

Granite magmas are, by essence, of crustal origin, as indicated by isotopic data. Therefore, initial isotopic strontium ratio ($I_{Sr}$) constitutes a good indicator of crustal contamination (Rollinson 1993). Hence, the mantle presents a relatively low and uniform $^{87}$Sr:$^{86}$Sr ratio that slowly increases linearly with time. In contrast, the ratio linearly evolves more rapidly from the time crustal rocks were formed and it starts from a value already higher compared to the mantle. The Sm:Nd ratio of the mantle is higher than that of the crust, implying that their isotopic ratios are in the same trend. However, the ratio for the crust and siliceous igneous rocks remains relatively uniform. The initial $^{143}$Nd:$^{144}$Nd ratio of the mantle linearly varies with age from the Bulk Earth composition. A coefficient, $\varepsilon_{Nd}$, represents the relative deviation of this ratio from the chondritic ratio. Starting with a null value of $\varepsilon_{Nd}$ at the formation of the Earth, mantle rocks display a linear increase with time, whereas crustal rocks show negative $\varepsilon_{Nd}$ values (Allègre & Ben Othman 1980).

The early finding of granitic magmas with high values of $I_{Sr}$ and negative $\varepsilon_{Nd}$ values indicated that a large amount of crust was necessary to explain granite genesis (Allègre & Ben Othman 1980). It also suggested that granites were mostly crustally derived.

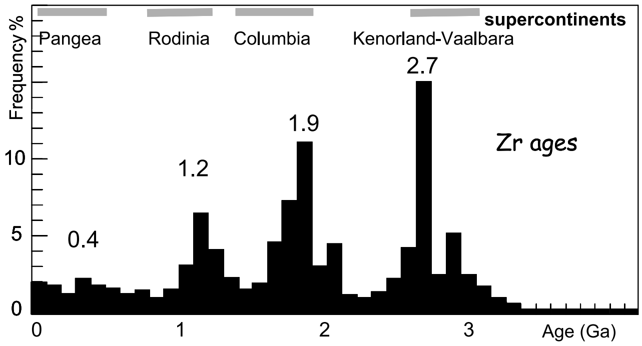


**Figure 7** Frequency diagram of zircon ages magmatism (redrawn from Condie 1998). The various peaks show the exacerbated magma production during amalgamation of supercontinents. It indirectly shows the role of mantle in producing granites by the increased amount of heat and stress generated during those periods.

### 5.2. The new paradigm

In the new paradigm, mantle acts as a material, heat and stress provider. Granites are obviously linked to plate tectonics (Pitcher 1993). They preferentially develop at plate margins, and show affinity with plate convergence and, to a lesser extent, with plate formation in the case of alkaline magmatism. Thus, it is not surprising to observe peaks in a frequency diagram of zircon ages of magmatic events (Condie 1998). These peaks reflect the abundant magma production linked to supercontinent cycles (Fig. 7) and document the role of the mantle during granite generation.

The importance of the mantle as a chemical reservoir is illustrated by the ubiquitous presence of mafic dykes and enclaves in a granitic magma (Didier & Barbarin 1991; Collins *et al.* 2000). Isotopic studies confirm the role of the mantle component with granitic rocks presenting positive $\varepsilon_{Nd}$. The diagram (Fig. 8) compiled for Hercynian and Caledonian granites from Europe and Australia, plus some granites from China (Jahn *et al.* 2000), demonstrates a balance of mantle and crustal components. Models of contamination indicate that most of data points can be explained by 25–66% of the crustal component, the volume of the mantle component thus consisting of three-quarters to one-third of the total material.

The mantle is also the heat engine that controls crustal melting. Mantle heat adds to the crustal heat produced by the radioactive decay of elements (U, Th and K). Common values of continental surface heat flow range from 40 to 120 mW m$^{-2}$, depending on crustal thickness and age of the geothermal province (Morgan 1984). The average value is about 60 mW m$^{-2}$. A common accepted balance is 10–15 mW m$^{-2}$ from the mantle, 5 mW m$^{-2}$ from the lower crust, which is notably depleted in both U and Th, the remainder from the uppermost crust. Increasing the thickness

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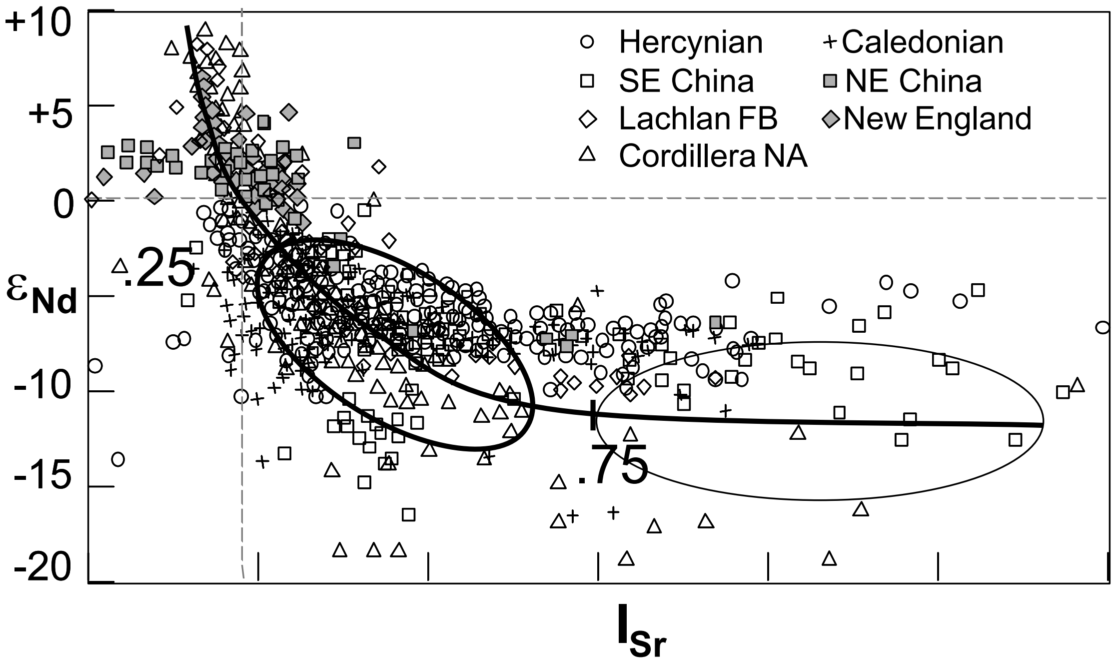


**Figure 8** Isotopic ratios ($I_{Sr}$ and $\varepsilon_{Nd}$) for Hercynian and Caledonian granites of Europe and south western Australia, north western Cordillera and from China (data from Jahn *et al*. 2000). Superimposed on the data sets are the field of the depleted mantle and upper crust. The field of the crustal component is shifted toward more negative $\varepsilon_{Nd}$ depending on the age of the protolith. Contamination by an older crustal reservoir (f. i. 1·0 Ga) shifts the values towards negative $\varepsilon_{Nd}$ ( − 10 units) and higher $I_{Sr}$. An average trend of mixing mantle and crustal components is indicated with their respective amount (0·25, 0·50 and 0·75). Most of the granitic magmas demonstrate the important contribution of the mantle (0·50–0–66).

of the heat-producing elements increases the surface heat flow, hence the geothermal gradient, which is 15–17 °C/km$^{-1}$ in standard conditions (Thompson 1999). In consequence, the temperature hardly reaches more than 525–600 °C at 35 km or at the Moho. It should be compared to the minimum temperature of 820 °C for biotite dehydration and 1000 °C for hornblende dehydration reactions (Patiño Douce & Beard 1995). Steady-state geotherms are not high enough to cause melting, attesting to the need for an additional heat source by either increasing crustal thickness or adding heat flux from the mantle. Doubling the average surface heat flow requires doubling of the crustal thickness or quadrupling the mantle heat flow.

Actually, high surface heat flow values (>100 mW m$^{-2}$) are measured in regions (e.g. the Alps, Himalayas and Andes) of thickened crust (Sclater *et al*. 1981). However, crustal thickening is not a mechanism likely to generate huge amounts of melt in a short time because increasing crustal thickness by thrusting takes time (Thompson 1999). Crustal erosion competes with heat diffusion and the crust returns to its normal thickness in about 100 Ma (England & Thompson 1984). No more than 20% melt is produced in about 70 Ma after crustal thickness doubling in cases of a standard geotherm and high (30 mW m$^{-2}$) mantle heat flow (Thompson 1999). In the case of a hot geotherm (40 mW m$^{-2}$), corresponding to tonalitic melts, melting starts 10 Ma after crustal thickening. A maximum (50%) of melt is produced 60 Ma later (Thompson 1999). Conversely, models in which the crust is thinned to half its initial thickness cannot achieve melting temperatures using a standard geotherm (Loosveld 1989) and melting is confined into the mantle. Only hot mode delamination, which would bring hotter mantle material to the Moho depth, can lead to high enough temperatures in the crust (Thompson 1999).

Other mechanisms for crustal melting require either advective or diffusive heat. Advective heat transport brings hot material from the mantle and rapidly heats up the crust. The time scale to induce melting is shorter by about one or two orders of magnitude compared to crustal thickening. However, latent heat effects restrict the melt volume. Six volumes of basalt are required to generate one volume of granitic magma (Bergantz 1995). The time for diffusive heat to progress varies 0·03 $L^2$ with distance (L) in kilometres and using a thermal diffusivity of $10^{-6}$ m$^2$ s$^{-1}$ (Crank 1975). When the heat source is at the base of the crust, it takes 3 Ma for heat to induce crustal melting at 22 km in depth, corresponding to about 600 MPa. When the heat source is at the base of the lithosphere, corresponding to heat released from the asthenosphere, the time delay may reach about 300 Ma. For instance, it corresponds to the formation of the so-called ‘anorogenic’ granites which lead to the 1·6–1·3 Ga formation of the rapakivi granites in Fennoscandia (Rämö & Haapala 1995). This situation requires that the continental plate remains essentially stationary over the heat source for this long. It can develop only in the case of a supercontinent cycle (Condie 1998), during which continental fragments amalgamate and remain stationary above a converging, and thus descending, mantle convective cell.

The mantle is also indirectly linked to granite generation as a stress provider to the continental crust. Because plate tectonics are partly driven by mantle convection, it generates permanent stresses within the plates (Zoback 1992). The stress pattern is characteristic of the continental crust. Stiffness of the thick and brittle upper crust allows large stress amplitude to develop. At the brittle/ductile transition, 100 MPa is necessary to induce extensional fractures in the continental crust and values of 600 MPa are common in cases of compressional environments (Vigneresse *et al*. 1999). Such high values contrast with stress that develops in a mantle environment. Stresses originating from the mantle control magma segregation, ascent and emplacement of granitic magmas in the crust. The high stress level in the continental crust precludes the transposition to crustal conditions of mechanisms acting into the mantle.

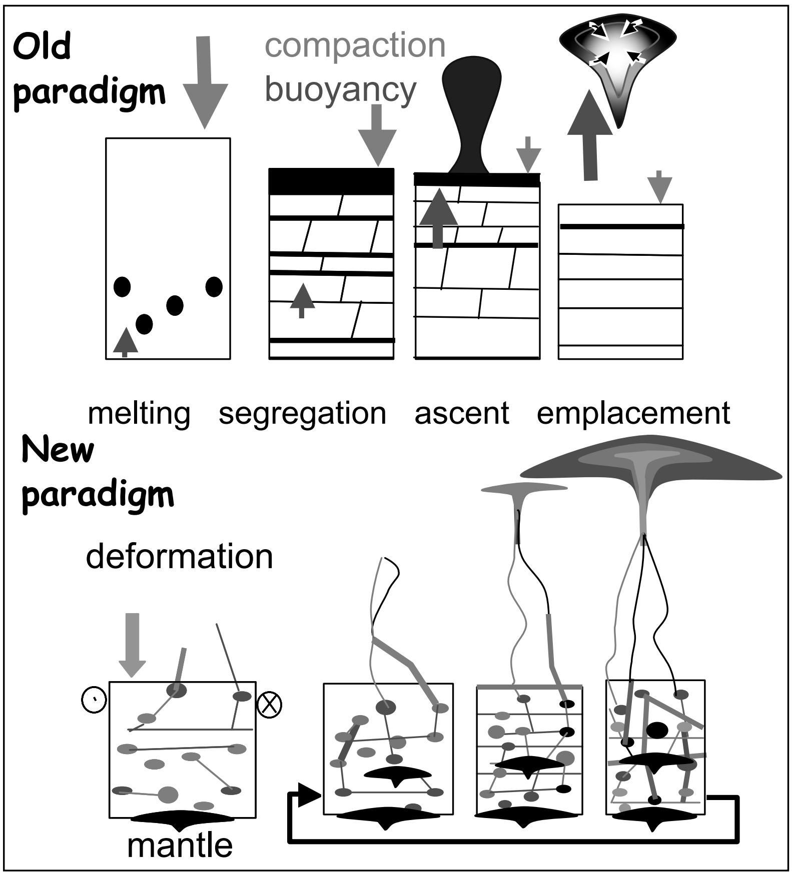


**Figure 9** MASH+AFC model (modified from Hildreth & Moorbath 1988) and the suggested new paradigm. In the old paradigm (top), four stages are identified: melting, segregation, ascent and emplacement. They develop successively, and each of them has its own characteristic length and time scale. In the new paradigm (bottom), the four phases are discontinuous, but they may develop at any time during the whole process. In addition, the external settings, represented by the mantle, act as a provider of heat and stress and as a chemical reservoir.

## 6. Global model

### 6.1. The old paradigm: MASH and AFC models

More accurate isotopic and trace element analytical methods allowed modelling of element variation during assimilation and fractional crystallisation or (AFC; De Paolo 1981; Powell 1984). The models were initiated with the equations describing isotope and trace elements behaviour (Shaw 1970) during fractionation in an open system or one periodically recharged with magma (O'Hara 1977). The most widely used AFC formalism describes the evolution of isotopes, and major and trace elements contained in a magma chamber affected by fractional crystallisation and wall rock assimilation at the same time (De Paolo 1981). Mixing with another magma may also be involved (Bergantz 2000). The necessary parameters are the distribution coefficients and the balance between the rate of assimilation to the rate of fractional crystallisation. When the ratio approaches infinite values, a simple binary model is approached. The variations of the distribution coefficients reflect the respective characters of compatible and incompatible elements within the given phase. Modifications to AFC have been introduced allowing magma replenishment (O'Hara & Matthews 1981), or decoupling between fractional crystallisation and assimilation (Cribb & Barton 1996). Energy-constrained assimilation and fractional crystallisation with replenishment has recently been suggested (Spera & Bohrson 2001). However, the basic assumption of a global ascent and emplacement of felsic magma with a volume similar to a pluton is still the basis of all AFC models.

The AFC models also require a coherent set of partition coefficients for felsic magmas (Bea *et al.* 1994). Their determination for major elements is ruled by Henry's laws, assuming equilibrium between melt and mineral formation (Rollinson 1993). This is valid for major elements contained within the major mineral phases. In contrast, incompatible elements (Th, Zr or REE) are incorporated as a separate phase in accessory minerals. The initial assumptions are consequently violated since elements are equally partitioned in the major minerals, but preferentially in accessory minerals. Thus, REE can hardly be used for modelling the genesis of granites (Bea 1996). It is necessary to check for which mineral assemblage a given set of partition coefficients is valid. For instance, the often-referred compilation of partition coefficients by Rollinson (1993) is only valid for basalts and basaltic andesites (Roberts & Clemens 1995). Determining a full set of partition coefficients for a granitic magma is a very difficult task and values commonly vary by several orders of magnitude for felsic melts. This has drastic effects on the fractional crystallisation curves.

Assimilation and fractional crystallisation (AFC) modelling is used on either trace element or isotopic data, but rarely on both at the same time (Roberts & Clemens 1995). When restricted to a set of elements, AFC modelling can always come out with a solution by modifying the respective rate of fractionation/assimilation or the degree of partial melting.

A global model for igneous intrusions is commonly designated as melting, assimilation, storage and homogenisation (MASH; Hildreth & Moorbath 1988). It applies to entire magma generation (Fig. 9a). Mantle and crustal magmas mix

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close to the mantle–crust boundary, establishing the characteristic chemical signature of magmas at the melting stage. The age and consequent chemical composition of the lower crustal component controls the isotopic variations, forming the assimilation stage. The depth at which mixing occurs controls the storage stage. Finally, AFC processes may substantially modify the composition of the ascending magmas, resulting in the homogenisation stage (Hildreth & Moorbath 1988). The MASH model is the first holistic approach to magma generation in convergent plate settings. Its major points consist of: (1) the importance of the crustally derived magma, with less involvement of mantle component; and (2) the role played by the mixing of magmas deep in the crust, in contrast to earlier ideas which implied an upper crustal magma chamber.

In conclusion, the current generally preferred model of granite generation assumes four separate stages which make up a global process (Fig. 9). A large amount of melt is required before segregation, and therefore, the bulk volume of buoyant magma drives magma ascent. A magma chamber is created by the input of magma that later differentiates and evolves chemically. The MASH model is basically static.

### 6.2. Toward a new paradigm

The new suggested paradigm is dynamic. Magma segregates by developing alternate periods of melt segregation with periods of matrix relaxation (Rabinowicz & Vigneresse 2004). Magma ascent also proceeds from a similar mechanism, discontinuous in space and time. Building a pluton consists of aggregating discrete pulses of magma, each having a unique chemical history and evolution (Bergantz 2000; Petford *et al.* 2000). Local mixing and chemical re-equilibration may take place, induced by the disturbance caused by the arrival of a new batch of magma into a more or less partially crystallised pool of magma. The discrete succession of magma batches interacts with the deformation field, locally altering the stress pattern. The intrusive body continuously modifies its internal shape and chemistry (Fig. 9), resulting in a petrographic zoning.

The discontinuous character of magma flow has important consequences for granite generation itself. It severely reduces the implications of results obtained from numerical models based on bulk magma chamber evolution. Most thermo-mechanical models of convective motion, crystallisation, cooling and chemical evolution caused by crystal-mush interactions assume a stationary process or a closed chemical system (see reviews in Cashman & Bergantz 1991; Ryan 1994). These models may remain valid, but on a much smaller scale. They could apply successively to the evolution of a body equivalent to a pluton in size, assuming that each pulse evolves individually. However, complex interactions and instabilities develop inherently with the decreasing time lag between magma batches.

The role of the mantle appears essential to all crustal processes. The mantle also certainly has an indirect role to play, since it brings the initial heat source or induces a stress pattern to the crust. It may also act as a counteractive parameter when its action switches off the controlling parameters.

## 7. Implications

The implications of the new paradigm are twofold. First, the changing context of several inputs of magma to build a granitic pluton certainly implies revisiting all models of chemical differentiation and fractionation. All schemes which incorporate a bulk *en masse* movement of magma should be reconsidered. Secondly, the successive inputs of magma, with varying chemical composition or temperature, have consequences on the way that the magma globally evolves. This should lead to a reconsideration of the very late stage of emplacement, which is often closely related to ore formation.

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