Rapid ascent of granitoid magmas from the lower crust

A.D. Brandon *Department of Earth, Atmospheric, and Planetary Sciences, Massachusetts Institute of Technology, Cambridge, MA 02139-4307, USA.*

R.A. Creaser
T. Chacko *Department of Geology, University of Alberta, Edmonton, Alberta, T6G 2E3 Canada.*

Introduction

Ascent rates and mechanisms are important in controlling the compositional modification of silicic magma. When magmas undergo rapid ascent, contamination by country rock and fractional crystallization are minimal, and magma compositions are controlled by source characteristics, extent of melting, and melt/restite separation. During slow ascent, more modification can occur, and resultant magma compositions do not accurately reflect the nature of the source rock. Not only do magma ascent times constrain the possible geochemical histories, but they also delineate whether the method of migration is diapirism (~1 m/yr), or fracture propagation via dyke systems ($\gg$1 m/yr, Clemens and Mawer, 1992; Petford *et al.*, 1993). Recent studies based on numerical modeling advocate dyke transport of granitoid magma, and suggest that dykes can fill greater than 1000 km^3-size silicic plutons in less than 1000 years (Clemens and Mawer, 1992). Here, we use direct petrologic constraints to evaluate the transport rates and mechanisms of granitoid magmas

Mineralogical evidence

Evidence from pressure-sensitive mineral assemblages that record high pressure crystallization, have been under-utilized. Epidote is now known to be stable at > 6 kbar in granitoid magmas (Naney, 1983; Schmidt, 1992, 1993). Magmatic epidote formed at ~8 kbar has been documented in dacite dykes in the Colorado Front Range emplaced at 2 to 4 kbar, and is present because of the rapid ascent rates resulting in quench of the higher pressure mineral assemblage (Dawes and Evans, 1991). Magmatic epidote has been identified in silicic plutons emplaced at greater than 6 kbar (Zen and Hammarstrom, 1984) confirming its high pressure stability.

Magmatic epidote has been found in the mid-Cretaceous White Creek batholith in southeast British Columbia (Brandon and Lambert, 1994). It is a zoned pluton ranging from quartz monzodiorite on the outer edge, with granodiorite towards the center. Epidote from the White Creek batholith has compositional characteristics typical for magmatic epidote in granodiorites reported elsewhere. The very low TiO_2 (< 0.1 wt.%) and pistacite content (Ps = $[Fe^{3+}/(Fe^{3+} + Al)]$) of Ps_{27} to Ps_{29} are especially noteworthy. Magmatic epidote in granitoids has typically less than 0.2 wt.% TiO_2. Pistacite content for magmatic epidote falls within the range of ~23 to 29 in granitoids.

The White Creek batholith was emplaced into metapelites. The mineral assemblage of the contact aureole surrounding the batholith consists of quartz, muscovite, sillimanite, and staurolite, with an emplacement level of 2.3 to 3.8 kbar (7 to 11 km depth). Pressure calculations using two calibrations of the Al-hornblende geobarometers (Johnson Rutherford, 1989; Schmidt, 1992) for White Creek hornblendes gives pressures from 2.2 to 4.0 kbar, and hence overlapping the range of pressures indicated by the contact aureole assemblage. Occurrence of epidote in this shallow-level pluton suggests epidote was preserved by rapid magma transport from the source region.

Dissolution experiments and ascent rates

To quantitatively constrain magma transport rates from the deep crust, we have examined the low-pressure dissolution of epidote in granitoid magmas experimentally. The experiments show that epidote breaks down at low pressures (<6 kbar) and have reaction rims consisting of quenched melt, plagioclase and magnetite at temperatures of 750 to 780°C, confirming its high pressure stability. Reaction rim growth can be modeled by a parabolic rate law following Rubie and Brearley (1988), and is controlled by diffusivity of major elements across the reaction rim. Given the reaction rim widths in the experiments, effective diffusion coefficients range from 2.3 × 10^{-15} to 6.1 × 10^{-15} m^2/s at 750 to 780°C. Using these coefficients, at 780°C a 2 mm diameter epidote crystal will dissolve in <100 years with reaction rim melt fractions of $>10\%$. At 700°C, 30 degrees above the 3 kbar water-saturated granite solidus, a 2 mm diameter epidote will dissolve in <1200 years.

These constraints can be applied to epidote-bearing dacite dykes and the White Creek batholith. Epidote phenocrysts within the dykes have highly corroded rims indicative of dissolution prior to solidification of the dacite melt. Epidote resorption will begin at 6 kbar during ascent and continue until final solidification at 2 kbar, which equates to at least 14 km of epidote reactive transport. Minimum transport rates can be calculated for 2 mm wide epidote crystals which are preserved in the dykes and the batholith using the reaction rim growth data. Applying thermobarometric analysis of the phenocryst assemblage in the dykes, indicates equilibration at 800–880°C at 7 to 12 kbar (Dawes and Evans, 1991). If the dykes were at 800°C upon emplacement at 2 kbar, with greater than 10% melt in contact at the rims of the epidote, a minimum rate of 350 m/yr is calculated. For epidote within the White Creek batholith with $<10\%$ melt fractions, a minimum ascent rate of 1750 m/yr is obtained at 780°C. Therefore, rapid magma ascent is required for epidote preservation in low pressure granitoid rocks, supporting models for dyke transport (Clemens and Mawer, 1992).

Implications and conclusions

Geological and experimental evidence support a model where the presence of epidote in shallow-level granitoid plutons results from rapid magma transport via dykes. Rapid ascent rates of granitoid magma likely preclude large amounts of interaction with the surrounding country rocks. Therefore compositional zonations in plutons such as the White Creek batholith, probably reflect crystallization of independent magmas batches derived from a heterogeneous source region (Brandon and Lambert, 1994).

References

Brandon, A.D. and Lambert, R.StJ. (1994) *J. Petrol.*, **35**, 239–69.

Clemens, J.D. and Mawer, C.K. (1992) *Tectonophysics*, **204**, 339–60.

Dawes, R.L. and Evens, B.W. (1991) *Geol. Soc. Am. Bull.*, **103**, 1017–31.

Johnson, M.C. and Rutherford, M.J. (1989) *Geology*, **17**, 837–41.

Naney, M.T. (1983) *Amer. J. Sci.*, **283**, 993–1033.

Petford, N., Kerr, R.C. and Lister, J.R. (1993) *Geology*, **21**, 845–8.

Rubie, D.C. and Brearley, A.J. (1988). A model for disequilibrium melting during metamorphism. In *High-temperature Metamorphism and Crustal Anatexis*, J.R. Ashworth and M. Brown, Eds., Unwin Hyman, London: 57–86.

Schmidt, M.W. (1992) *Contrib. Mineral. Petrol.*, **110**, 304-310.

Schmidt, M.W. (1993) *Amer. J. Sci.*, **293**, 1011–60.

Zen, E. and Hammarstrom, J.M. (1984) *Geology*, **12**, 515–8.