Silicic magmas in the Izu-Bonin oceanic arc and implications for crustal evolution

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Crustal structure and chemistry of volcanoes

Rhyolite magmas occur to a greater or lesser degree in most volcanoes along the Izu-Bonin volcanic front from Hakone to Torishima, a distance of 550 km. Their origin, together with that of dacitic magmas, in oceanic arcs is a matter of considerable interest and debate (*e.g.* Tamura and Wysozanski, 2006). Recent studies of along-arc crustal structures show a clear correlation between the average chemistry (basaltic vs rhyolitic) of the Quaternary volcanoes and the average seismic velocity of the underlying arc crust (Kodaira *et al*., 2007). Thus it appears that the origin of arc magmas and their chemical variations are directly related to the origin and evolution of the plutonic and volcanic arc crust. In the case of the Izu-Bonin arc, where basaltic volcanoes tend to stand on the lower velocity (or more continental-like) crust (Kodaira *et al*., 2007), this is surprising considering the complex history of the 50 Ma crust. We will show the relationship between the chemical compositions of rhyolites and the average crustal velocity and thickness of middle crust beneath the volcanoes, and discuss the origin of rhyolites along the Izu-Bonin volcanic front by examining their chemical variations.

Conclusions

A number of conclusions are reached. 1) Remelting of middle crust and the formation of rhyolite magmas may have occurred both in the basaltic and rhyolitic volcanoes. 2) Basalt volcanoes consume newly produced middle crust at higher temperatures to produce rhyolite magmas. 3) Rhyolite volcanoes may have no mantle roots. The melts that feed these volcanoes originate, instead, in the middle crust. Lateral intrusions of basaltic dikes cause low temperature melting in the old (Oligocene?) middle crust, generating the rhyolitic magmas.

References

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Forsterite-rich and Ca-poor olivine phenocryst crystallized by polybaric melting in a subduction zone

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The role of compositionally diverse components in basaltic lavas has been a long-standing problem about the ultimate origin of crystal grains. In convergent plate margin settings, the origin of forsterite-rich olivine grains with vastly different CaO contents within the same lava has been debated, and their coexistence may indicate several genetic populations in a single magma system (e.g. Kamenetsky *et al*., 2006). The presence of these forsterite-rich olivine grains with lower-forsterite olivine grains also suggests a short residence time of these crystals in the magma chamber.

In this study, we will present the detailed magmatic processes of Laguneta volcano that erupted in an extensional, behind the volcanic front (BVF) graben area in El Salvador, Central America. The Laguneta lava is olivine-tholeiitic to alkaline in composition and has the highest MgO contents among the Salvadorian BVF lavas. Incompatible trace element compositions, characterized by the enrichment of fluid-mobile elements, indicate the involvement of slab-derived flux, although the degree of its involvement was much lower than that of volcanic front lavas. The high LREE/HREE ratio and excess ^{230}Th reveal the presence of garnet in the residue. Petrological evidence, presence of forsterite-rich (up to Fo93) and low-Ca (down to 0.07 wt. %) olivine 'phenocrysts', on the other hand, indicate that the parental magma was segregated from cpx and garnet-free residue. Chemical composition and zoning profiles of olivine phenocrysts, and the chemical composition of spinel inclusions, demonstrate in detail the characteristics of multiple magma batches, their mixing order, and their magmatic residence time after the formation of olivine phenocrysts. We conclude that the forsterite-rich and Ca-poor olivine phenocrysts in the lava was crystallized from the final drops of mantle-derived magma fractions that formed by the decompression melting of metasomatized sub-arc peridotite. The extensional tectonic setting near the volcanic front facilitated rapid magma ascent and extraction.

Reference

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