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Oxygen isotope geochemistry of high-Mg andesites and adakites: Implications for slab melting

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The melting of subducted lithosphere contributed to the formation of continental crust during the Archean, but is now restricted to modern hot subduction zones, and generates adakites and high-Mg andesites. Upper oceanic crust (the top ca. 1-3 km) is higher in δ^{18} O than common mantle rocks (ca. 11-14% vs. 5.5%), whereas lower ocean crust is often lower (ca. 2-6%). Thus, δ^{18} O values of slab melts are indicative of the place of their segregation. We find δ^{18} O values of olivine phenocrysts from >35 adakites and high-Mg andesites (Aleutians, Kamchatka, Setouchi, Fiji, C. America, Cascades, and Andean Austral zone) are consistently ¹⁸O enriched and range from 5.0 to 6.5%, with a majority within the 5.2-5.7 range—higher than $\delta^{18}O_{olivine}$ in NMORBs (5.0-5.2‰), and in typical IAB (5.2±0.2%). However, deviations in δ^{18} O from typical terrestrial basalts are subtle, requiring that amounts of subducted upper oceanic crust in their sources are small and/or that high- δ^{18} O upper crust and low- δ^{18} O lower crust counter-balance one another. Relationships between δ^{18} O and other geochemical properties provide evidence for the causes of subtle ¹⁸O enrichments in adakites. First, those with trace element signatures suggesting abundant slab fluid components in their sources are modestly elevated (0.1 to 0.5%; e.g., Shasta in the Cascades) whereas those that appear to sample unusually large amounts of sediment melt (Setouchi, Japan) are more strongly elevated (0.7 to 1.1%). Second, where significant lower crust amount of lower crust are involved (e.g. Kamchatka) adakites are 1-2% elevated. Thirdly, the 'end-member' case slab melts from C. Aleutians and Cook Island (Andes) are NMORB-like to 0.1-0.2% depleted. Finally, there is an overall correlation of increasing δ^{18} O with decreasing Sr/Y, La/Yb, and increasing 87Sr/86Sr.

We suggest a model whereby adakites sample large (ca. > 10 %) amounts of subducted ocean lithosphere, and the $\delta^{18}O$ of that slab component depends on the proportion coming from thin but very high- $\delta^{18}O$ sediments/pillow basalts on the top of the plate vs. the thick layer of underlying, hydrothermally-altered and moderately low- $\delta^{18}O$ gabbros. A chanellized vs. porous flow during melt segregation is preferred.

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Boron isotopic composition of Hawaiian shield lavas

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Geochemical variations in Hawaiian lavas have led geochemists to the conclusion that recycled subducted materials are involved in the source of the Hawaiian magma. If this is the case, the B isotopic ratios of the subducted component should retain a memory of low-temperature fractionation processes that occurred in ancient near sea-floor surface and ancient subduction systems because B isotope values are not fractionated by high-temperature magmatic processes. However, the primitive Hawaiian source mantle is believed to have homogeneous δ^{11} B values of -10 ‰ relative to NIST SRM951. On the other hand, the large variation of δ^{11} B values observed in Hawaiian volcanic glasses determined by secondary ion mass spectrometry (SIMS) that extend towards higher values are thought to have been caused by assimilation process during magma ascent [1].

In this study, B isotopic ratios of fresh tholeiite and picrite lavas from three representative Hawaiian shields; Mauna Loa (N = 6), Kilauea (N = 11); and the Makapuu-stage of Koolau [2] (N = 8) were determined by thermal ionization mass spectrometry (TIMS) after removing secondary absorbed contaminants. The mean $\delta^{11}B$ values systematically decrease in the order: Mauna Loa (-3.3 %), Kilauea (-4.0 %), and Koolau (-5.1 ‰). The variation of $\delta^{11}B$ values obtained from each volcano shows nearly the constant values (2σ values of ~ \pm 0.6 %, comparable to the range of analytical reproducibility. The whole range of newly determined $\delta^{11}B$ values is much smaller than that previously determined by SIMS (-3.8 ± 4.0 to +0.6 ± 4.0 % for Kilauea (N = 4) and - 10.6 ± 3.0 to -4.0 ± 2.0 % for Loihi (N = 6) [e.g. 1, 3]) and is comparable to the analytical error of SIMS analyses of individual samples.

The Sr, Nd, and Pb isotopic compositions of these lavas show large variations in each shield. Hence, the homogeneous $\delta^{11}B$ values in each shield suggest that the different source components must have only small differences in $\delta^{11}B$. However, the systematic difference of the $\delta^{11}B$ between the three shields suggests that on a larger scale both in space and time the different source components have at least detectable small differences in $\delta^{11}B$.

References

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