Boron isotope variations in OIB and BABB

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Introduction

B is potentially a very good tracer of mantle geodynamic because surface reservoirs are enriched in B compared to mantle rocks and have very different $\delta^{11}B$ values between continental crust (mean between 13 and 8‰, Chaussidon and Albarède, 1992) and oceanic crust which has interacted with seawater (between 0 and +15‰, Spivack and Edmond, 1987). One of the key points of the B geochemical cycle is its behavior during subduction processes and the fraction which is returned to surface reservoirs through arc magmatism. A recent study of the B isotope composition of mid oceanic ridge basalt (MORB) glasses has shown a restricted range of variations, between -6.5 ± 2‰ and -1.2 ± 2‰, for N-MORB and E-MORB (Chaussidon and Jambon, 1994). These data suggest that less than 2% in mass of the B hosted in the subducted plate is reinjected in the upper mantle.

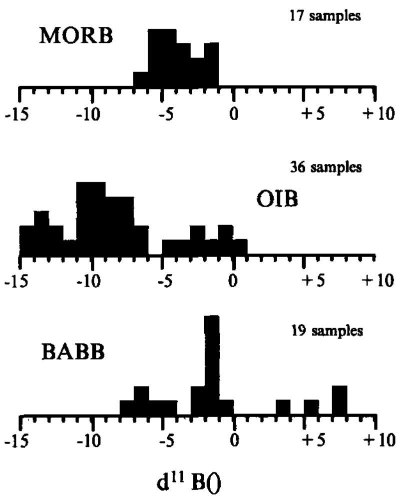

FIG. 1 : histogram showing $\delta^{11}B$ values for MORB OIB and BABB.

Samples and analytical techniques

All samples studied are fresh basaltic glasses from the Manus basin, the Lau basin and the Mariana for BABB and from Loihi seamount, the Galapagos and St Helena islands for OIB. In addition, a few primary melt inclusions from Iceland are considered. The $\delta^{11}B$ values and B contents were determined with a Cameca ims 3f ion microprobe at CRPG (Nancy) following procedures previously described (Chaussidon and Libourel, 1993 ; Chaussidon and Jambon, 1994). Analytical uncertainty is ± 2‰ (1 sigma) for the $\delta^{11}B$ values, and $\approx \pm 10\%$ for B contents.

Results

All results are shown in Fig. 1 together with some preliminary data from Iceland (Gurenko and Chaussidon, in prep). OIB have clearly $\delta^{11}B$ values which extend to much lower values than MORB, with a mode at $\approx -10 \pm 3$‰. On the contrary, BABB have more positive $\delta^{11}B$ values, from -8 up to $+7$‰. These systematic differences of B isotopes between OIB, MORB and BABB are similar to that shown by He isotopes, but contrary to the variations of some other major isotopic systems (e.g. O, Sr, Nd) for which no clear difference between BABB and OIB are present. OIB with high $^3He/^4He$ ratios (>15) have low $\delta^{11}B$ values of -11 ± 2‰, while BABB with lower

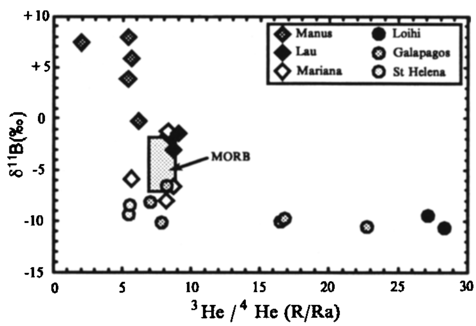

FIG. 2 : $\delta^{11}B$ values versus $^3He/^4He$ ratios.

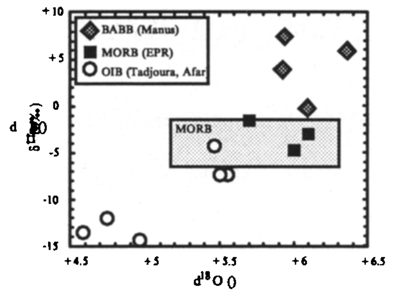

FIG. 3 : $\delta^{11}B$-$\delta^{18}O$ relationship for OIB from Tadjoura and BABB from Manus basin.

$^3He/^4He$ ratios (between 2.1 and 9.2) have very variable $\delta^{11}B$ from values similar to MORB up to +8‰ (Fig. 2). BABB samples with very high $\delta^{11}B$ values are also very rich in B with contents up to ≈10ppm. In addition, for some samples, $\delta^{11}B$ values are correlated with other isotopic systems, as with $\delta^{18}O$ values for instance (Fig. 3).

Discussion

Two types of processes can explain most of the $\delta^{11}B$ variations found in OIB and BAAB: (1) late stage interaction between mantle melts and the oceanic crust and (2) within mantle heterogeneities. Assimilation of oceanic crust hydrothermalized with seawater at low temperature increases the $\delta^{11}B$ values of melts (e.g. suite of OIB samples from Kilauea, Chaussidon and Jambon, 1994). On the contrary, the present OIB samples from Tadjoura show very low $\delta^{11}B$ values associated with low $\delta^{18}O$ values (Fig. 3) which could result of assimilation of crust hydrothermalized at high temperature (Barrat *et al.*, 1993). This process could also produce low $\delta^{11}B$ values because B isotopic fractionations between hydrothermal fluids and rocks are positive (Palmer and Sturchio, 1990). Thus, assimilation processes occurring in the oceanic crust might produce very variable $\delta^{11}B$ values.

Assuming that melting and fractionation processes are unable to produce the B isotope variations observed because B is strongly incompatible and will remain in the melt (Chaussidon and Libourel, 1993), three main processes can cause the range of $\delta^{11}B$ observed for mantle melts. (1) The $\delta^{11}B$ variations observed in the mantle result of an addition of subducted material to a mantle (depleted and enriched) with a $\delta^{11}B$ value of $\approx -12\pm2$‰. The B content of the mantle being very low (≈0.065 ppm for the depleted mantle), additions of low amounts of recycled B with high $\delta^{11}B$ values (between 0 and +10‰) can produce the differences observed between OIB and MORB (Chaussidon and Jambon, 1994). The process could however be more complicated since the $\delta^{11}B$ value of the altered oceanic crust is likely to be variable with depth. In fact, B with positive $\delta^{11}B$ values is likely to be concentrated in low temperature secondary phases in the top of the subducted slab, while deeper portions of the oceanic crust could gain negative $\delta^{11}B$ values during high temperature exchanges with metamorphic fluids. (2) MORB melts are systematically enriched in ^{11}B (from -12 ± 2‰ to -3.5 ± 2.5‰) because of late stage interaction with seawater or seawater altered oceanic crust. This process could be general and occur via assimilation of roof rocks in MORB magma chambers. However, $^{40}Ar/^{36}Ar$ ratios in MORB are generally high (up to 28 000) clearly showing that this process is not occurring for MORB in most cases. (3) Primitive variations in $\delta^{11}B$ values dating from the differentiation or the accretion of the Earth are present in the mantle. Our knowledge of B isotopic fractionations at high temperature and of the natural variability of $\delta^{11}B$ values in chondrites being very poor, these processes cannot be at the moment critically evaluated.

Conclusion

$\delta^{11}B$ values show rather systematic variations between OIB (mode at -10 ± 3‰), MORB (between -6.5 ± 2 and -1.2 ± 2‰) and BABB (between ≈−8 and ≈+7‰). Interaction of mantle melts with the oceanic lithosphere (shown for different OIB localities) and within mantle heterogeneities, probably due to the addition to the mantle of B isotopically fractionated in the oceanic crust, are at present the two more obvious processes controlling the $\delta^{11}B$ of mantle melts.

References

Barrat, J.A., Jahn,B.M., Fourcade, S. and Joron, J.L. (1993) *Geochim. Cosmochim. Acta*, **57**, 2291–302.

Chaussidon, M. and Jambon, A. (1994) *Earth Planet. Sci. Lett.*, **121**, 277–91.

Chaussidon, M. and Libourel, G. (1993) *Geochim. Cosmochim. Acta*, **57**, 5053–62.

Chaussidon, M. and Albarède F. (1992) *Earth Planet. Sci. Lett.*, **108**, 229–41.

Gurenko, A. and Chaussidon, M. *Boron isotope composition of Icelandic primary melts* (in prep).

Palmer, M.R. and Sturchio, N.C. (1990) *Geochim. Cosmochim. Acta*, **54**, 2811–15.

Spivack, A.J. and Edmond, J.M. (1987) *Geochim. Cosmochim. Acta*, **51**, 1033–43.