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Boron content and isotopic composition of tektites and impact glasses: Constraints on source regions

MARC CHAUSSIDON[1] and CHRISTIAN KOEBERL[2,*]
[1]CRPG-CNRS, BP 20, 54501 Vandoeuvre-Les-Nancy, France
[2]Institute of Geochemistry, University of Vienna, Dr.-Karl-Lueger-Ring 1, A-1010 Vienna, Austria

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Abstract—Abundances of Li, Be, and B, as well as boron isotopic compositions, were determined in twenty-seven tektite and impact glass samples, using an ion microprobe. Samples included tektites from the Australasian, North American, and Ivory Coast strewn fields, and Aouelloul and Darwin impact glasses. Variations of B abundance and isotopic composition in a flanged australite were also studied. δ^{11}B variations of only a few permil were found within the australite flange. The isotopic composition shows no correlation with the B contents or with the distance from the rim of the flange. The mean δ^{11}B value for the flanged australite is very similar to that of Muong-Nong type tektites (−1.9 ± 1.9‰). Thus, vapor fractionation has been unimportant during tektite formation. This is supported by the observation that B contents and the δ^{11}B values of the different samples from the Australasian tektite strewn field are not correlated with each other.

Most tektites show a rather limited range of δ^{11}B values (−9.3 ± 1.5 to +2.7 ± 1.5‰), which is small compared to the range observed for common terrestrial rocks (−30 to +40‰). The B abundance and isotopic data can be used to place constraints on the tektite source rocks. Australasian tektites have high B and Li abundances; only clay-rich sediments, such as pelagic and neritic sediments, as well as river and deltaic sediments have B contents (up to 100 ppm) and δ^{11}B values that are in agreement with the range shown by Australasian tektites (−4.9 to +1.4‰). ^{10}Be and Rb-Sr data indicate continental crustal source rocks and exclude pelagic and neritic sediments. However, deltaic sediments, e.g., from the Mekong river, which are of continental crustal origin, agree with ^{10}Be, Rb-Sr, and B data, and support a possible source locality close to the coast of SE Indochina in the South China Sea. On the other hand, one bediasite sample has a very high δ^{11}B value of +15.1 ± 2.1‰, requiring the presence of marine carbonates or evaporites among the source rocks. This finding supports an offshore locality for the North American tektite source crater.

INTRODUCTION

Tektites are natural glasses, which are found on earth in four distinct strewn fields (e.g., Barnes, 1963), i.e., the North American (35 Ma), Central European (15 Ma), Ivory Coast (1.1 Ma), and Australasian (0.77 Ma) tektite strewn fields (numbers in parentheses are the radiometrically determined age of tektites in each strewn field). Geochemical studies have shown unambiguously that tektites were derived by melting due to hypervelocity impact from terrestrial upper crustal rocks (see, e.g., Taylor, 1973; Koeberl, 1986, 1992, 1995; Blum et al., 1992). For two of the four tektite strewn fields (Ivory Coast, Central European), the possible source craters are known with a great degree of certainty (the Bosumtwi, and Ries craters, respectively), on the basis of chemical, isotopic, and age data. No crater identifications, or connections with known craters, have been made for the North American or Australasian strewn fields, although there are a number of suggestions and some possible candidates for both (see below).

There is now a renewed interest in the study of tektites and related impact glasses, because of their crucial role in understanding certain aspects of impact cratering. Impact glasses occur at several impact craters all over the world, and impact glasses have been found to be associated with the Cretaceous-Tertiary boundary (e.g., Sigurdsson et al., 1991; Koeberl and Sigurdsson, 1992). In this respect, the behavior of volatile elements and compounds (e.g., CO_2, SO_2, water) is of interest, because of possible climatic effects of a meteorite impact on the Earth.

Although tektites have first been found on land, microtektites (usually <1 mm in diameter) were found in deep-sea sediments at three of the four strewn fields. The occurrence and distribution of microtektites helps to define the geographical extent of the strewn fields. Tektites (on land) occur in three main groups: (1) normal or splash-form tektites, (2) aerodynamically shaped tektites, and (3) Muong Nong-type (or layered) tektites. The aerodynamic ablation results from partial remelting of the tektite glass during atmospheric reentry after it was ejected outside the terrestrial atmosphere and solidified through quenching. Such aerodynamically shaped tektites are known mainly from the Australasian strewn field, primarily as flanged-button australites. The shapes of splash-form tektites (spheres, droplets, teardrops, dumbbells, etc., or fragments thereof) are not aerodynamical forms, but result mostly from the solidification of rotating liquids. Muong Nong-type tektites, named after a type-locality in Laos, are usually considerably larger than normal tektites and are of chunky, blocky appearance. They have a layered structure with abundant vesicles, and some of them contain mineral inclusions (e.g., quartz, zircon, chromite, rutile,

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[*] Author to whom correspondence should be addressed.

corundum, cristobalite; see, e.g., Glass and Barlow, 1979). They are mostly found in Indochina, associated with the Australasian strewn field, although recently a few rare examples from other strewn fields were recognized as well. However, in the present paper we refer only to Muong Nong-type tektites from the Australasian strewn field.

The petrography, as well as the chemical and isotopic composition of tektites, is in complete agreement with an origin by impact melting of terrestrial rocks. Recently, osmium isotope studies have shown that a very small, but clear extraterrestrial signature is present in at least some tektites (Koeberl and Shirey, 1993). While there are still a few open questions regarding the exact mechanism of tektite formation, some basic facts are clear. The production of tektites requires specific impact conditions, because otherwise many more tektite strewn fields would exist on earth. Tektites are produced by nonequilibrium shock melting of surficial rocks, and the superheated melts may undergo a plasma phase, during which they are subjected to partial reduction. The melts are transported through the atmosphere (probably in the wake of the expanding vapor cloud), quenched, and distributed over a geographically extended area—the strewn field.

Chemical and isotopic studies have been essential in helping to determine the type of source rocks for each tektite strewn field (see below). An important observation is the high ^{10}Be content of Australasian and Ivory Coast tektites, which is evidence for a derivation from terrestrial surface rocks, and precludes an origin from greater depth in the crater (e.g., Pal et al., 1982; Middleton and Klein, 1987; see below). For the Central European and the Ivory Coast strewn fields, the source rocks are well constrained, because the source craters are known. The same degree of certainty does not exist for the North American and the Australasian tektite strewn fields, and, thus, geochemical constraints on the composition of the target rocks are helpful in determining the location of the source crater.

The study of the content and isotope composition of B in tektites can help to constrain the target rocks, from which tektites have formed, for three main reasons: (1) the extraterrestrial B contribution to the composition of tektites is very small because meteorites have very low B abundances, compared to most surface terrestrial rocks, and the meteoritic contribution to the tektite composition is always $\leqslant 1$ wt%; (2) terrestrial rocks exhibit large variations in their B contents, ranging from very low contents (0.1–2 ppm in oceanic basalts) to several 100 ppm in sediments and up to several wt% in some differentiated crustal rocks and minerals; and (3) the isotopic composition of B shows large variations in the global geochemical cycles, which leads to, for example, significant differences between marine and nonmarine sediments, and, within marine sediments, between clay minerals or carbonates. Thus, the determination of the boron isotopic characteristics of tektites, together with other chemical data, should help to differentiate between several different target rock types.

In addition, the study of the abundances and isotopic composition of B in tektites, especially in different types of tektites found in the same strewn field, can provide useful information on vapor fractionation effects. Such effects could occur in the aerodynamically ablated tektites, e.g., in the flanges of an australite. Earlier determinations of boron concentrations in tektites and impact glasses were reported by, e.g., Mills (1968), Taylor and Kaye (1969), Chapman and Scheiber (1969), and Matthies and Koeberl (1991). Here, we present the first survey of B isotope compositions in a total of twenty-seven samples of tektites from three different strewn fields, the internal variations in a flanged australite, and of impact glasses from the Aouelloul and Darwin craters. These measurements were made by ion microprobe, which offers the advantage of small-scale in situ analysis. This is a necessity for the study of tektites, which can be very heterogeneous on a micrometer scale.

SAMPLE DESCRIPTION AND ANALYTICAL METHODS

Samples and Sample Localities

We have studied various tektite samples, mainly from the Australasian strewn field, as well as some samples from the North American and Ivory Coast strewn fields. The sample suite includes eleven Muong Nong-type indochinites, two splash-form indochinites (thailandites), and one flanged australite. For more detailed sample descriptions and other data on the Muong Nong-type tektites, see Koeberl (1992) (samples 8301–8317) and Glass and Koeberl (1989) (samples 5424, X-102, X-103, and F-16).

Muong Nong-type tektites show major (and most trace) element abundances that are almost identical to those or splash-form tektites, but they are chemically less homogeneous than normal tektites, and there are some differences in composition between the layers. However, the contents of water and some volatile elements (e.g., B, Zn, the halogens) are higher in Muong Nong-type tektites, compared to splash form tektites (e.g., Koeberl and Beran, 1988; Matthies and Koeberl, 1991; Koeberl, 1992). The volatile element "enrichment" in Muong Nong-type tektites probably results from less intense heating (i.e., shorter heating event and lower temperatures), compared with splash form tektites. They formed from the same or a very similar source material, but were less thoroughly heated and have, therefore, lost less of their volatile elements than the splash form tektites. Volatile element contents in Muong Nong-type tektites are comparable to, or only slightly lower than, the ones in upper crustal sedimentary rocks (see discussion in Koeberl, 1992), while splash-form tektites show considerable depletions.

We also analyzed three bediasite samples from the North American strewn field, 8401, 8402, and 8501 (for details on samples and chemical analyses, see Weinke and Koeberl, 1985; Koeberl, 1988), and two Ivory Coast tektites, 2069 and 2114 (see Koeberl and Shirey, 1993; Chamberlain et al., 1993; C. Koeberl et al., unpubl. data). In addition, three impact glasses from the Aouelloul impact crater, Mauritania, and five impact glasses from the Darwin crater, Tasmania, Australia, were studied.

Aouelloul impact crater

This simple, bowl-shaped 3.1 Ma old crater has a diameter of about 390 m and is exposed mainly in Ordovician Zli sandstone, with some Oujeft sandstone present. No unequivocal mineralogical evidence of shock metamorphism has yet been described from the crater rocks. However, planar fluid inclusion trails (presumably remnants of planar deformation features) in form of distinct sets in quartz grains were recently found in Zli sandstone from the crater rim (C. Koeberl and W. U. Reimold, unpubl. data). Numerous fragments of a dark inhomogeneous impact glass, which contains abundant schlieren and partly digested quartz and feldspar grains, have been found mainly outside the crater rim, and, to a lesser degree, on the inner slope (e.g., Campbell-Smith and Hey, 1952). The composition of the glass is similar to that of the Zli sandstone, but a few, mainly siderophile, elements show some enrichments in the glass (e.g., Koeberl and Auer, 1991), which was used to suggest that the crater was formed by an iron meteorite projectile (Morgan et al., 1975). For the present study, we analyzed three Aouelloul impact glasses (samples AOL-10, −15, and −30). One Zli sandstone sample was analyzed as well.

TABLE 1 . Major element compositions of tektites and impact glasses (in wt%) (1)

sample	SiO_2	Al_2O_3	FeO(2)	MgO	CaO	Na_2O	K_2O	TiO_2	Total
A : Australasian strewn field									
1 : Muong Nong-type tektites									
8301	75.43	11.42	4.50	1.52	1.18	1.33	2.62	0.71	98.71
8306	77.01	11.13	3.87	1.56	1.27	1.27	2.37	0.70	99.18
8309	82.57	8.50	3.11	1.13	0.99	1.19	2.32	0.41	100.22
8310	82.26	8.78	3.13	1.25	1.22	1.27	2.26	0.47	100.64
8311	76.66	10.97	3.87	1.50	1.52	1.29	2.51	0.66	98.98
8317	75.78	10.88	4.59	1.54	1.36	1.21	2.58	0.72	98.66
5424	71.62	13.70	4.64	2.37	2.86	1.80	2.58	0.98	100.55
X102	73.97	12.49	4.25	1.65	1.28	1.35	2.56	0.65	98.20
X103	69.53	14.82	5.13	2.43	2.95	1.74	2.79	0.89	100.28
F-16	76.44	11.14	4.60	1.60	1.24	1.14	2.54	0.22	98.92
2 : Splash form tektites (thailandites)									
8202	73.75	13.06	4.08	1.89	1.65	1.36	2.52	0.74	99.05
8203	74.65	12.35	4.33	1.90	1.79	1.03	2.19	0.70	98.94
3 : Flanged tektites (australites)									
USNM 3463 (3)	69.83	14.25	4.89	2.35	3.42	0.96	2.29	0.81	98.80
B : North American strewn field (bediasites)									
8401	79.39	12.22	2.62	0.47	0.53	1.47	2.27	0.61	99.58
8402	76.07	13.21	3.58	0.73	0.81	1.80	2.30	0.80	99.30
8501	77.26	13.09	3.48	0.77	0.49	1.73	2.42	0.64	98.88
C : Ivory Coast strewn field									
2069	68.18	16.52	6.51	3.86	1.28	1.92	1.70	0.47	100.44
2114	68.33	16.17	6.41	3.85	1.27	2.39	2.06	0.51	100.99
D : Aouelloul impact glasses									
10	85.96	6.59	2.27	1.14	0.38	0.57	2.73	0.53	100.17
15	86.30	6.75	2.50	1.19	0.33	0.36	2.14	0.38	99.95
30	87.04	6.44	2.48	1.18	0.36	0.29	2.12	0.49	100.40
E : Darwin impact glasses									
8702	80.15	10.51	4.07	0.79	0.00	0.03	2.61	0.63	98.79
8705	87.21	7.21	2.17	0.60	0.03	0.00	1.71	0.56	99.49
8709	88.37	7.04	1.16	0.83	0.00	0.01	1.80	0.61	99.82
8711	87.53	7.24	2.11	0.44	0.00	0.01	1.80	0.51	99.64
8712	85.49	6.96	2.70	1.02	0.03	0.00	1.96	0.43	98.59

(1) major element compositions analysed with a Camebax electron microprobe at the Université de Nancy I (J. M. Claude, analyst).

(2) : all Fe as FeO.

(3) : This composition is the mean of 19 analyses of the rim (see Fig 2 and Appendix 1).

Darwin impact crater

This about 0.74 Ma old ≤1 km-diameter-structure, situated about 26 km SSE of Queenstown, Tasmania, is not well exposed and, therefore, difficult to study (Fudali and Ford, 1979). Impact glasses have been found in this area long before the structure was identified as an impact crater. The chemical composition of the glasses is very similar to that of terrestrial sediments and can be reproduced by mixtures of local target rocks, such as quartzites and shales (see Meisel et al., 1990, for details on the composition of Darwin glass and target rocks). For the present study, we analyzed five Darwin impact glasses (samples DG 8702, 8705, 8709, 8711, and 8712).

Analytical Methods

The contents of Li, Be, and B, and the isotopic compositions of B were determined with a Cameca ims 3f ion microprobe at CRPG-CNRS (Nancy, France). The procedures used are the same as the ones previously described for peridotites and oceanic basaltic glasses (Chaussidon and Libourel, 1993; Chaussidon and Jambon, 1994). The samples were polished, cleaned in alcohol and coated with gold, and sputtered by an O-primary beam. The analyzed volume corresponds in general to a hole about 10 μm in diameter and 2–3 μm deep. Care was taken to obtain values for the glass and to avoid measuring any mineral inclusions. Positive secondary ions were analyzed at a mass resolution of about 1800, and an offset of -60 ± 10 V was applied for Li, Be, and B concentration measurements. The detection limits are about 0.01 ppm for Li and B and about 0.1 ppm for Be. The Li, Be, and B contents were determined with a precision of about ±5 rel% in the range 20 ppm to more and about ±10 rel% for contents < 20 ppm. The $\delta^{11}B$ values were determined with an accuracy of ±1.5‰ (1 sigma) and calculated relative to the NBS 958 standard (see Table 2). Precision is about ±1 to ±1.5‰. In addition, the major element compositions of all samples were measured by standard electron microprobe techniques near the spots also analyzed by ion microprobe.

RESULTS

Lithium, Beryllium, and Boron contents of Tektites and Impact Glasses

The major element compositions and the Li, Be, and B contents of all tektite and impact glass samples are given in Tables 1 and 2. The major element composition of the tektites and impact glasses analyzed here shows a range of SiO_2 contents between 68 and 89 wt%. Similarly, Li and B contents are very variable, from 1.8–101.4 ppm and from 2.0–82.1 ppm, respectively. In contrast, the Be contents are less variable, ranging from 0.8 to 2.9 ppm. The range of Li, Be, and B contents is consistent with previously published analyses of Muong Nong-type tektites (Koeberl, 1992; Matthies and Koeberl, 1991), but the present data show a strong positive correlation between the Li and B contents for nearly all the

TABLE 2 . Li, Be and B contents and $\delta^{11}B$ values of tektites and impact glasses

sample	ppm Li	ppm Be	ppm B	$\delta^{11}B$(1)
A : Australasian strewn field				
1 : Muong Nong-type tektites				
8301	50.3	2.3	43.5	-1.6
8301 (2)	-	-	-	-0.3
8306	61.1	2.6	51.8	-2.6
8306 (2)	-	-	-	-0.6
8309	52.0	2.0	35.2	-6.1
8309 (2)	-	-	-	-4.8
8310	60.3	2.4	53.9	-0.5
8310 (2)	-	-	-	-1.0
8311	46.9	2.3	46.1	-4.9
8311 (2)	53.1	2.4	49.6	-4.8
8317	42.2	2.1	46.0	-2.2
8317 (2)	-	-	-	-2.5
5424	87.6	2.9	77.8	-1.6
5424 (2)	101.4	2.9	82.1	-
X102	12.0	2.9	55.3	-0.6
X102(2)	-	-	-	-1.1
X103	78.1	2.6	69.1	-1.3
X103(2)	79.6	2.7	70.0	-
F-16	49.0	2.1	44.5	+0.4
F-16 (2)	-	-	-	+1.2
2 : Splash form tektites (thailandites)				
8202	52.8	2.8	20.2	-1.4
8203	65.5	2.8	15.0	-3.9
3 : Flanged tektites (australites)				
USNM 3463	-	-	24.7(3)	+1.4(3)
USNM 3463	-	-		+0.4(3)
USNM 3463	-	-	-	+0.2(3)
B : North American strewn field (bediasites)				
8401	3.9	2.5	3.0	+2.7
8402	4.9	2.3	8.3	+15.1(4)
8501	4.3	2.2	6.4	-1.3
C : Ivory Coast strewn field				
2069	10.4	1.6	7.9	-4.8
2114	5.7	0.8	2.0	-1.6
D : Aouelloul impact glasses				
10	2.6	1.2	4.9	-6.0
15	3.1	1.5	6.9	-4.4
30	3.1	1.5	3.4	-5.0
E : Darwin impact glasses				
8702	3.6	2.4	18.8	-3.7
8705	2.9	2.1	9.0	-8.6
8705 (2)	-	-	-	-9.3
8709	2.5	1.7	4.7	-2.5
8711	2.7	2.7	5.0	-6.2
8712	1.8	1.2	16.7	-1.4

(1) : $\delta^{11}B$ values are given in ‰ (±1.5‰, 1 sigma) relative to the NBS 958 standard ($^{11}B/^{10}B_{NBS\ 958} = 4.04558$) with : $\delta^{11}B = 1000 \times ((^{11}B/^{10}B_{sample} / ^{11}B/^{10}B_{NBS\ 958}) - 1)$.

(2) : Same sample than previous line but analysed at a different spot.

(3) : B contents and $\delta^{11}B$ values are averages (24.7 ppm ±3.6 [24 points], 1.4±1.7‰ [14 points], 0.4±1.2‰ [10 points] and 0.2±3.4‰ [22 points]) for 3 traverses of ≈2.5 mm across the glass.

(4) this $\delta^{11}B$ value is the mean of 5 analyses at different spots (15.1±2.1‰).

samples investigated (Fig. 1), which was not obvious in previous studies. Contrary to Li, Be contents are not correlated with B contents, except for Muong-Nong type tektites (solid circles in Fig.1).

The results of the present study cannot be directly compared with the previous analyses (Koeberl, 1992). Differences are likely to be due to the observation (Koeberl, 1992) that Muong Nong-type tektites show a significant heterogeneity at the micrometer scale. The major element compositions of the Muong Nong-type tektites listed by Koeberl (1992) were also obtained by electron microprobe analysis, but on different parts of the samples. However, the Li, Be, and B contents of Koeberl (1992) were obtained on bulk samples, while the present in situ ion probe analyses can be directly correlated with the major element compositions obtained at almost the same location on the samples. Nevertheless, the B contents found in the present study agree reasonably well with data by Matthies and Koeberl (1991), sometimes for the same samples. For example, we find 44.5 ppm B for the Muong Nong-type tektite F-16 and 20.2 ppm for the thailandite 8202, compared with 44 and 17 ppm by Matthies and Koeberl (1991).

Significant differences in the contents of Li and B are present between samples from the different tektite strewn fields. Samples from the Australasian strewn field have higher Li and B contents by a factor of up to 10, compared to North American and Ivory Coast tektites, which all have Li and B contents < 20 ppm (Table 2). Within the Australasian strewn field, the Muong-Nong type tektites have higher Li and B contents than thailandites and australites (Figs. 1 and 2), while they all have similar Be contents. This relationship was already observed before (e.g., Koeberl, 1986, 1992; Koeberl and Beran, 1988; see also Introduction). The differences between tektites from different strewn fields can reflect (1) mixing of heterogeneous source rocks in various proportions during the impact, and/or (2) volatilization processes during impact and melting. As tektites from the three strewn fields studied here result from three different source areas (source craters), such differences are, in fact, expected.

In comparison, the impact glasses from the Aouelloul and Darwin craters show low Li and B contents, ranging from 1.8–3.1 ppm, and from 3.4–18.8 ppm, respectively. Previously obtained Li data for the Aouelloul glass (Koeberl and Auer, 1991; C. Koeberl et al., unpubl. data) show a range of about 2.9–6 ppm Li and agree reasonably well with the present analyses. No B data were previously available for Aouelloul glass. For Darwin glasses, Matthies and Koeberl (1991) found a range of about 9–13 ppm B, which agrees with a range of about 5–19 ppm in the present study. In contrast, a (so far unexplained) larger difference exists between previous (bulk) and present Li data for the Darwin glasses (Meisel et al., 1990), as well as for Be contents in both Aouelloul and Darwin glasses, which are consistently higher in the present study.

Boron Isotope Composition of Tektites and Impact Glasses

All the tektites and impact glasses analyzed show a rather limited range (but no uniform value) of their boron isotopic compositions (δ^{11}B values ranging between −9.3 ± 1.5‰ and +2.7 ± 1.5‰), with the single exception of one bediasite sample, which has a δ^{11}B value of +15.1 ± 2.1‰ (Table 2). This latter value cannot be due to an artefact or erroneous measurement, as this sample has been remeasured several times (see Table 2). The average values are very similar to the only four previously published values, which showed a range between −6.6 ± 1.5‰ and −5.1 ± 1.5‰ for two indochinite, one rizalite, and one bediasite sample (Agyei and McMullen, 1978). The average of our δ^{11}B values for the samples from the Australasian and the North American strewn fields are, however, a few permil higher than the previously published values. This difference is probably not significant, because Muong-Nong type tektites for instance show δ^{11}B variations of up to about 5‰ between different samples, while, with one exception, no significant variations (in excess of 1.5‰, equivalent to the analytical accuracy) have been found within a single sample (Table 2). The only sample that showed small δ^{11}B variations within a single sample (of ±3‰) was the flanged australite (USNM 3463), where such variations were found within the flange, but are apparently not correlated with the B contents.

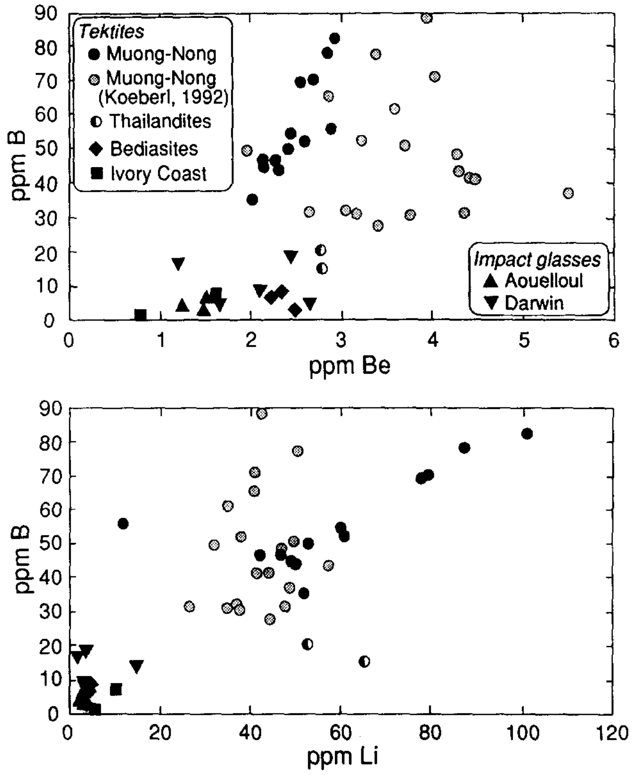

FIG. 1. Relationships between the Li, Be, and B contents for tektites and impact glasses. Muong Nong-type tektites and thailandites are enriched in Li and in B, compared to all other samples and show a correlation between Li, Be, and B (see text).

No significant systematic difference in δ^{11}B is observed between samples from the different tektite strewn fields, between different components of the same strewn field, and between tektites and impact glasses. However, the lowest δ^{11}B values are found for the Darwin impact glasses, while the bediasites have the highest ones. These differences are likely to be the result of different target rock compositions (see

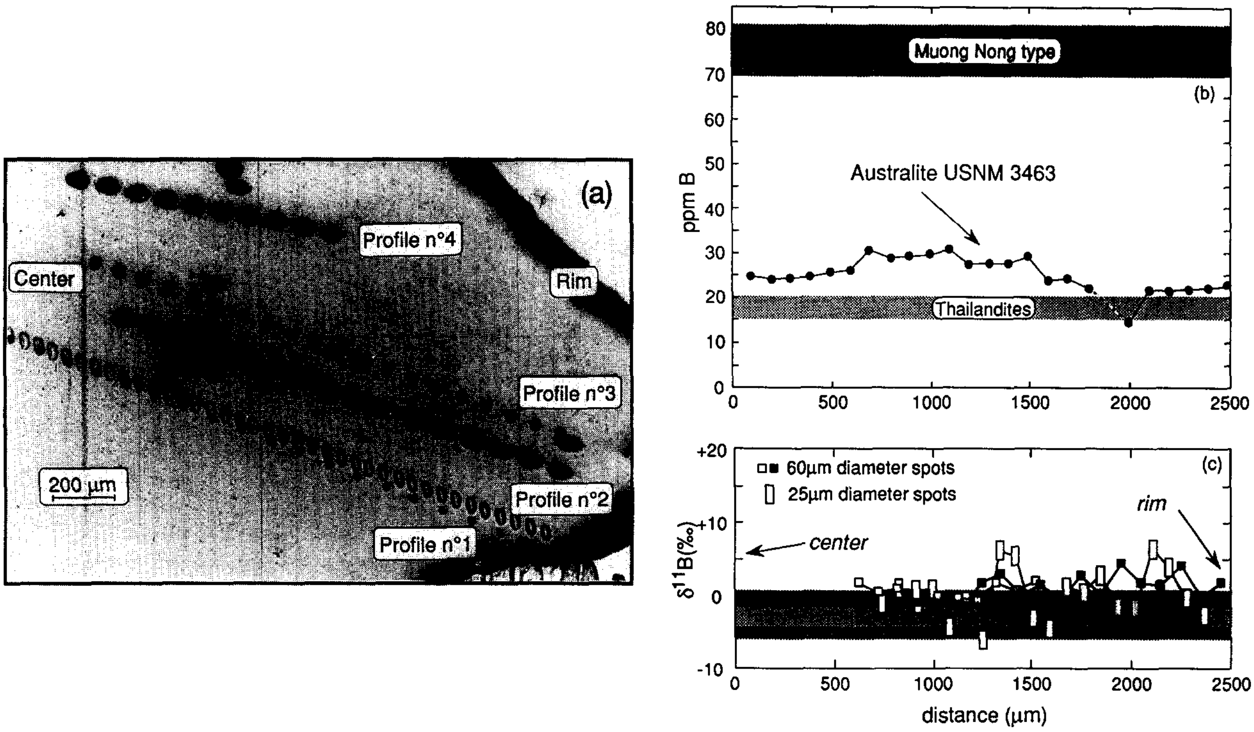

FIG. 2. Variations of the B contents and $\delta^{11}B$ values in the rim of the australite USNM 3463. (a) Photo showing the rim due to the partial remelting in the high atmosphere of the tektite: B contents were obtained along profile # 1, while $\delta^{11}B$ values were measured along the three other profiles, # 2 to # 4. The diameter of the analyzed area varies between about 5 µm for the B concentration profiles, about 60 µm for isotopic profiles # 2 and 4, to 25 µm for isotopic profile # 3. (b) Boron concentration profile. Fields are shown for Muong Nong-type tektites and thailandites, which belong to the same strewn field. (c) $\delta^{11}B$ profiles. A higher $\delta^{11}B$ variability is observed for the profile with the smaller analytical spots, suggesting some heterogeneity at the micrometer scale. The mean $\delta^{11}B$ of australite USNM 3463 is only slightly different from the values of Muong Nong type tektites, implying a that the B isotopic composition was not fractionated during the remelting process.

below). No obvious correlation exists between the $\delta^{11}B$ values and the B contents or any other elemental contents, but a slight decrease of the $\delta^{11}B$ values with increasing SiO_2 contents or decrease of B contents seems to be present when all the samples are considered together (Fig. 3a,b).

DISCUSSION

Behavior of Boron during Impact Melting

Two processes may influence the elemental and isotopic abundance of boron. First, volatile elements are volatilized to various degrees from the source rocks during impact melting. This process occurs when impact melt is formed, and the degree of volatilization is dependent on the temperature. Second, elements in the hot gas may undergo mass-dependent vapor fractionation. The present results allow to constrain two parameters: (1) the volatility of B during impact melting and mixing of terrestrial rocks, and (2) the presence of possible vapor fractionation of the boron isotopes.

It seems that the variations of the B contents (and, to some degree, also of the Li contents) of the Muong-Nong type tektites, compared to those of the splash-form tektites from the same strewn field, are primarily controlled by volatilization processes (Koeberl, 1992). For B, the influence of mixing of various heterogeneous source rocks during the impact is probably less important than volatility effects. This is shown by the poor correlation between the B contents and major element contents (Tables 1, 2).

Beryllium is considered to be a refractory element, similar to Al, while B is generally considered to be a volatile element. Boron has been observed in some volcanic emanations (Nomura et al., 1982) and its calculated condensation temperature (as boride or borate) in the solar nebula is about 700 K (Cameron et al., 1973). Such a temperature of 700 K would imply a depletion factor of B in the Earth, compared to chondrites of about 0.08 (Higgins and Shaw, 1984). This value is lower than those derived from bulk Earth (silicate) abundances that are estimated at about 6.5 ± 0.8 atoms B per 10^6 atoms Si (i.e., depleted by a factor of 0.27, relative to chondritic abundances), or a study of mantle xenoliths (Higgins and Shaw, 1984) and at 3.1 ± 1.2 atoms B per 106 atoms Si (depletion factor of 0.15), and from the study of oceanic basalts (Chaussidon and Jambon, 1994). These data indicate a condensation temperature for B ranging between 1200 K (similar to that of Li) and 800 K, making it only moderately volatile. In fact, the volatilization behavior of B (and, to a lesser extent, of Be and Li) during impact melting is probably strongly dependent on the form and speciation, in which B occurs in the target rocks. The high Li and B contents of the Muong-Nong type tektites suggest that these elements were structurally bound to silicates, such as clay minerals or tourmaline, but did not occur in the form of oxides, hydroxides, and/or carbonates or evaporites.

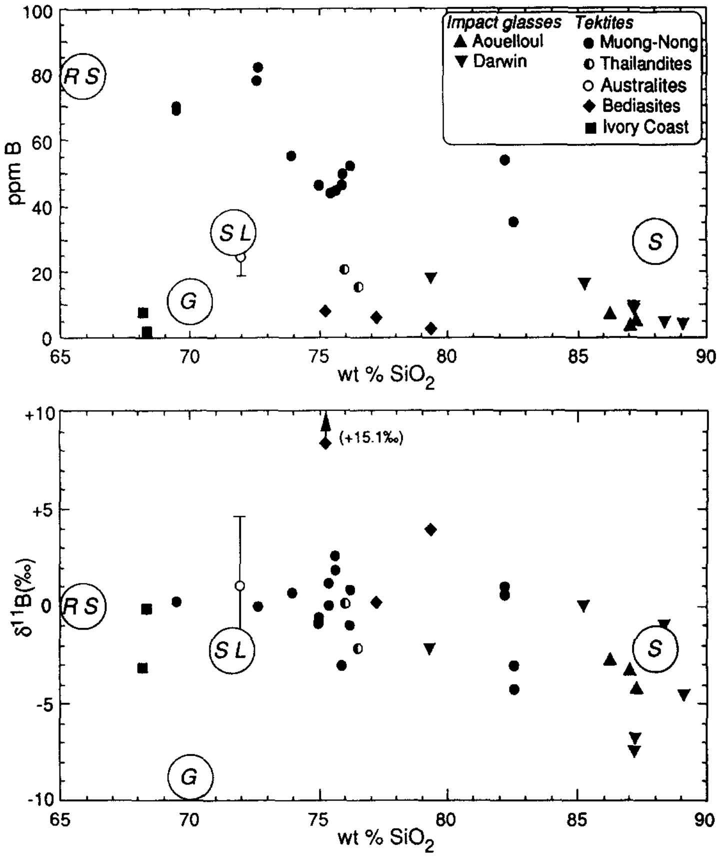

FIG. 3. Variation of the B contents and $\delta^{11}B$ values of tektites and impact glasses vs. their SiO_2 contents. The four points represent average values of some continental rocks of interest (see text for values): RS for river sediments, G for granites, SL for soils and loess, and S for sandstones. A higher sandstone component is likely for the impact glasses, while river sediments are likely to be part of the source rocks of the Australasian tektites (see text).

Thailandites and australites have lower B contents than Muong-Nong type tektites, which can be attributed to a larger degree of volatilization of B from the source rocks than in Muong-Nong type, as already noted before (e.g., Koeberl, 1986, 1992; Koeberl and Beran, 1988). This is also shown in Fig. 1, as the thailandites have similar Li and Be contents as the Muong Nong-type tektites, but different B contents. Also, there is no statistically significant difference in the $\delta^{11}B$ values between the splash-form and Muong Nong-type tektites (Table 2), indicating that mass-dependent vapor fractionation was minimal. The small $\delta^{11}B$ variations of a few permil that are present between several Muong Nong-type tektites (Table 2) can probably be attributed to a small boron isotope variability of the impacted rocks. No significant boron isotope fractionation due to B loss by mass-dependent vapor fractionation is observed within the Muong Nong-type and splash-form tektites.

Volatilization has most likely affected the composition of most tektites in the Australasian strewn field. However, it is important to constrain the extent of volatilization and vapor fractionation from the melt, which are, in fact, two different effects. There seems little doubt that the more volatile elements (e.g., the halogens, Cu, Zn, Ga, As, Se, Pb) were volatilized to various degrees from the source rocks upon melting, largely depending on the peak temperature. This is indicated because the Muong Nong-type tektites, suspected to have experienced the lowest temperatures of all tektites, still have a much higher complement of these elements than the splash-form tektites (Koeberl, 1992). However, it was also proposed that compositional differences of tektites can by explained by volatilization of silica at high temperatures and oxygen fugacities (Walter, 1967; Walter and Clayton, 1967). However, Molini-Vesko et al. (1982) have shown that the silicon isotope systematics in tektites do not agree with what would be expected if selective silica volatilization has taken place, and that, therefore, vapor fractionation may not have played a significant role. Esat and Taylor (1986) and Esat (1988) obtained similar results from the study of magnesium isotopes in microtektites. This conclusion was also reinforced by the measurements of Humayun et al. (1994), who determined the potassium isotopic composition of tektites. They found that tektites have potassium isotopic compositions that are indistinguishable from those of terrestrial rocks, and that there is no evidence for vapor fractionation.

To further test for possible vapor fractionation effects, we measured the B contents and isotopic compositions along several profiles across the rim of one flanged australite (Fig. 2a). Delano (1992) has found that, on a micrometer scale, the core of a flanged australite has a more heterogeneous composition than the rim. Delano (1992) suggested that this is due to the second heating event, which remelted the flange material and led to a greater homogeneity of the latter. It is conceivable that, because of the second intense heating pulse, the flange of an australite would provide the best sample to look for vapor fractionation effects in tektite material.

The results, shown in Figs. 2b,c, indicate δ^{11}B variations of only a few permil. These variations are apparently not related to the B contents (Fig. 2b) and show no significant correlation with the distance from the rim of the flange. The mean δ^{11}B value for this australite is just slightly higher than that of Muong-Nong type tektites ($-1.9 \pm 1.9‰$), but overlap within errors bars. The averages calculated from three different profiles give mean δ^{11}B values of $+1.4 \pm 1.7‰$, $+0.4 \pm 1.2‰$, and $+0.2 \pm 3.4‰$ (Table 2). This implies that, if present at all, any boron isotope fractionation between vapor and melt ($\delta_{\text{vapor-melt}} \approx 1000 \times \ln \alpha_{\text{vapor-melt}}$) is close to 0‰, and less than 3‰ (in absolute value). In fact, assuming a Rayleigh-type isotopic distillation, $\delta_{\text{vapor-melt}}$ in excess of 3‰ would produce systematic differences in δ^{11}B between Muong-Nong type tektites and australites of more than 4‰ (in absolute value) assuming a B loss of 50% (see Table 2). Thus, while a comparison between Muong Nong type and splash form tektites indicate some elemental B fractionation, the isotopic fractionation of B is minimal ($\delta_{\text{vapor-melt}}$ between 0 and 3‰). This may be due either to the high temperature of the melt, or because B occurs mostly in the B^{III} speciation in the Si-rich melt (Chakraborty et al., 1993). Also, no correlation exists within the Australasian strewn field between the B contents and the δ^{11}B values (Fig. 4). This conclusion, that vapor fractionation has not played a major role during tektite formation, is in agreement with earlier observations (e.g., Molini-Vesko et al., 1982; Esat and Taylor, 1986; Humayun et al., 1994). However, the assumption has to be made that any outer australite rim (which may show more severe fractionation) was not ablated during atmospheric passage.

Constraints on Source Materials of Tektites

The range of δ^{11}B variations found in tektites, between $-9.3 \pm 1.5‰$ and $+2.7 \pm 1.5‰$, is somewhat restricted when compared to the range observed for common terrestrial rocks (between -30 and $+40‰$, Table 3). This narrow range cannot be explained by the incorporation in the tektites and the impact glasses of an "extraterrestrial B component" because (1) meteorites have very low B contents between 0.2 and 2 ppm (Weller et al., 1978 ; Shaw et al., 1988) and (2) their δ^{11}B values are probably very variable, as shown by a range of -50 to $+40‰$ found for single chondrules in carbonaceous and ordinary chondrites (Chaussidon and Robert, 1994). On the contrary, only terrestrial rocks can account for the high B contents (between 2 and 82 ppm) found in tektites.

If we accept our previous conclusion, namely that no significant vapor fractionation occurred during the impact, the B contents and δ^{11}B values of tektites and impact glasses can be used to constrain the composition and nature of their source rocks. This seems more easily possible for the Australasian tektites, the bediasites, and the Darwin impact glasses, which exhibit a larger variation in their B-δ^{11}B relationship. On the other hand, the Ivory Coast tektites and the Aouelloul impact glasses show lesser variation in their B contents and δ^{11}B values, and the small number of samples analyzed does not show a clear trend.

Regarding the source rocks of the Australasian tektites, we need to explain the relatively high B and Li contents in these tektites, which are much higher than in any other tektite group (Table 2; Fig. 4). However, the δ^{11}B values of the Australasian tektites are not significantly different from those of other tektites. While soils and loess cannot be excluded as source rocks, the high B contents (up to 82 ppm) and the range of δ^{11}B values (-4.9 to $+1.4‰$) of Australasian tektites suggest that marine pelagic and neritic sediments, as well as river and delta sediments (Table 3; Fig. 4), are likely source rocks. Such

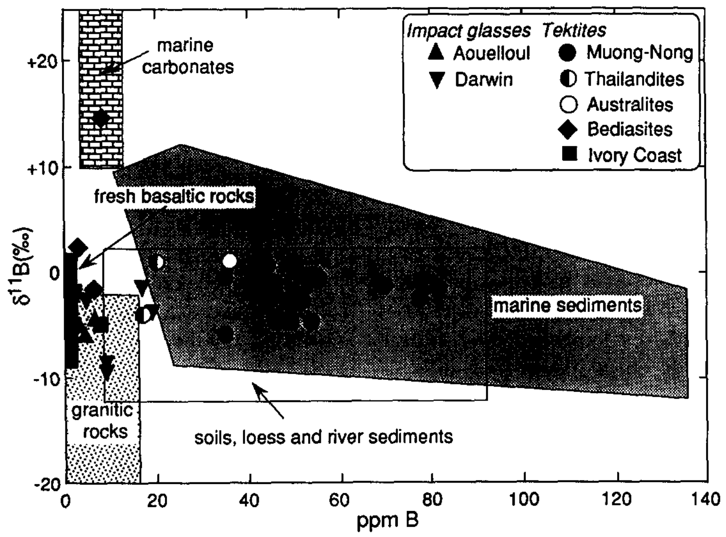

FIG. 4. Boron contents and δ^{11}B values of tektites and impact glasses compared to the ranges of terrestrial rocks. Sandstones are not shown, but would plot in the left side of the field for soils, loess and river sediments. A sedimentary component (soils, river sediments, sandstones, and/or marine sediments) is required in the source materials of most of the samples, except for the Ivory Coast tektites. Marine carbonates must be present among the source rocks of bediasites.

TABLE 3 . B contents and isotopic compositions of some terrestrial rocks

Rock type	ppm B	δ^{11}B (‰)
Magmatic rocks		
granitic rocks	1 to ≈150[1]	-30 to -2† [2]
massive sulfide deposits and tourmalinites	n. g.	-22.8 to +18.3† [3]
"average" granitic crust	10 [1]	-13 to -8 [2]
fresh oceanic basalts	0.5 to 2 [4]	-12 to -2 [4]
fresh island arc basalts	3 to 64 [5,6]	-6.8 to +7.3 [5,6]
altered oceanic crust	9 to 81 [7]	+0.1 to +12.5 [7]
Non-marine sedimentary rocks		
soils and loess	8 to 76 [8,9]	-12.4 to -0.7 [8,9]
river and delta sediments	64 to 94 [8]	-6.7 to +1.9 [8]
non-marine evaporites	n. g.	-21.9 to +0.3 [10]
Marine sedimentary rocks		
marine evaporites	n. g.	+18.2 to +31.7 [10]
marine carbonates	15 to 26 [9]	+8.0 to +10.5 [9]
modern pelagic sediments	95 to 132 [9]	-6.2 to -1.8 [9]
modern neritic sediments	70 to 85 [9]	-5.4 to -2.2 [9]
ancient marine sediments	0.1 to 131 [9]	-17.0 to -5.6 [9]
Seawater	4.5 [7]	+39.5 [7]

† : δ^{11}B values are those for tourmalines which contain most of the B present in the rock.
n. g. : not given : the B contents of these rocks are extremely variable, between a few tens of ppm B to several wt% B_2O_3, for pure tourmalinites or pure borate deposits.
Data are from [1] HARDER (1970), [2] CHAUSSIDON and ALBAREDE (1992), [3] PALMER and SLACK (1989), [4] CHAUSSIDON and JAMBON (1994) and CHAUSSIDON and MARTY (1994), [5] PALMER (1991), [6] ISHIKAWA and NAKAMURA (1994), [7] SPIVACK and EDMOND (1987), [8] SPIVACK et al. (1987), [9] ISHIKAWA and NAKAMURA (1993), [10] SWIHART et al. (1986).

clay-rich sediments are the most likely source rocks from the B systematics, though of course a few rare continental rock types, for example, some B-rich pegmatites or tourmalinites and/or some continental evaporites, could have similar B contents and δ^{11}B values. However, such rocks are rather unlikely as major source rocks of the tektites. The B contents of soils and loess might be too low to explain the high B abundances in Muong Nong type tektites, especially if some B loss (up to 50%) has to be taken into account (the high B end of the soils, loess, and river sediments field in Fig. 4 is defined by the river sediments). The high Li contents of the Muong Nong-type and splash-form tektites are also consistent with the presence of neritic or deltaic sediments, as these sediments may contain up to 96 ppm Li (Ishikawa and Nakamura, 1993).

This is an interesting result, as it seems to place some constraints on possible locations for the source crater of the Australasian tektite strewn field. So far, no source crater is known, although many proposals for possible source craters were made and later discounted (see, e.g., Koeberl, 1992, 1995, for a review). Some of the more recent proposals include a possible offshore impact location (about 175 km to the east of the Vietnam seashore; Schnetzler et al., 1988), a location in the Laotian mountains (Schnetzler, 1992), or the about 100 km long and up to 35 km wide Lake Tonle Sap in Cambodia (Hartung and Koeberl, 1994). None of these locations has so far been confirmed. The discovery of impact debris in deep-sea cores near the coast of Indochina as well as the fact that the quantity of both impact debris and microtektites in cores all over the Australasian strewn field increases towards Indochina (Glass and Wu, 1993) supports a location in or near the southeastern part of Indochina.

As the search for a source crater for the Australasian field has not yet been successful, geochemical arguments regarding the source are very important. Blum et al. (1992) studied the Sm-Nd and Rb-Sr isotope systematics of the Australasian tektites and found that their source material was derived from a Proterozoic crustal terrane, which underwent its last major Rb-Sr fractionation event around 175 Ma ago. This age is interpreted as the time of the deposition of the sedimentary target rocks of probably Jurassic age, which occur in the central and SE part of Indochina. Thus, from the chemical and isotopic data, there is no doubt that the source rocks of the Australasian tektites were not marine, but continental rocks.

Further relevant clues can be obtained from the high ^{10}Be abundances in Australasian tektites, which can only have been introduced from sediments that have absorbed ^{10}Be that was produced in the terrestrial atmosphere (e.g., Pal et al., 1982; Middleton and Klein, 1987). The high ^{10}Be contents again indicate a continental source. The study of ^{10}Be is also important for the understanding of the target stratigraphy. The concentration of ^{10}Be in the environment is a strict function of the depth from the surface; only rocks (or sediments) close to the surface contain appreciable amount of ^{10}Be. Blum et al. (1992) have calculated that mixing of a 200 m column of bedrock into the surficial cover that contains the ^{10}Be explains the ^{10}Be concentrations observed in Australasian tektites. The absolute concentrations of ^{10}Be in the various types of Australasian tektites allow another important conclusion. The ^{10}Be concentrations in australites are higher than those in indochinites and philippinites, which, in turn, are higher than those in Muong Nong-type tektites (e.g., Pal et al., 1982; Raisbeck et al., 1988; J. Klein, pers. commun., 1992; also, see discussion in Koeberl, 1992, 1995). This allows the conclusion that australites were made from material very close to or at the surface, while the Muong Nong-type tektites originated from a slightly deeper unit, in agreement with impact mechanics and compositional data.

The contents of ^{10}Be in pelagic sediments are very low, and they also have low accumulation rates. However, the largest ^{10}Be inventories exist near the continental margins, where accumulation rates are 10 cm per year or higher, and large ^{10}Be concentrations can be found at greater depths than on land (J. Klein, 1992, pers. commun.). Eroded material from a large area of the continental crust is deposited in the form of river and deltaic sediments (which have continental crustal composition). The Mekong river, one of the largest and most sediment-rich rivers in the world, deposits large quantities of sediment in the shallow South China Sea. In conclusion, the B abundances and isotopic ratios found in Australasian tektites are in agreement with a source locality slightly offshore, maybe close to the coast of southeastern Indochina, which would contain continental crustal debris with the right B and ^{10}Be contents and Rb-Sr, Sm-Nd, and boron isotopic composition.

In contrast with the δ^{11}B values of other tektites and impact glasses, one bediasite sample has a very high δ^{11}B value of +15.1 ± 2.1‰. This value requires the presence of some marine carbonates and/or evaporites in the source rocks. Some rocks altered at low temperature by seawater cannot be excluded either (Table 3). In fact, seawater is rich in B (4.5 ppm) and has a strongly positive δ^{11}B value of +39.5‰ (Spivack and Edmond, 1987), resulting in increased B contents and positive δ^{11}B values for magmatic rocks altered at low temperature in the marine environment (Spivack and Edmond, 1987). In contrast, diagenetic processes in sediments and/or high temperature hydrothermal alteration of silicate rocks, even in the marine environment, are likely to significantly decrease (by a few to 10 permil) the original δ^{11}B values of the rocks (Ishikawa and Nakamura, 1993; Vengosh et al., 1991). This is caused by the preferential incorporation of B in tetrahedral coordination in solids rather than in fluids, with tetrahedral B being depleted in ^{11}B compared to trihedral B (Kakihana et al.,1977). Therefore, the high δ^{11}B value found in the bediasite sample points, most probably, to carbonates and/or evaporites in the source rocks. This is quite similar to the case of some the high-Ca impact glass spherules found at the K/T boundary in Haiti, which have B contents between 10 and 29 ppm and high δ^{11}B values of up to +11 ± 2‰ (M. Chaussidon et al., unpubl. data). Our results are in agreement with an offshore source crater for the North American strewn field; for example the proposed Chesapeake Bay or Toms Canyon crater structures, which are both under water, on the continental shelf (Poag et al., 1992, 1994; Powars et al., 1993).

The B contents and δ^{11}B values of the Ivory Coast tektite samples are distinct from those of the Australasian tektites (and also from the North American tektites), and are consistent with low-B source rocks, such as average granites and Precambrian metasediments, which are the source rocks at the Bosumtwi crater in Ghana (see Introduction). A B content of 17 ppm was measured for a sample of Zli sandstone from the Aouellou crater. The low B contents of the Aouelloul impact glasses are due to the low B content of the target sandstones and some volatilization (Fig. 3). The δ^{11}B values of the Aouelloul impact glasses are in agreement with a average continental source rock.

Finally, the Darwin impact glasses are characterized by intermediate B contents, between 5 and 19ppm, and by the lowest δ^{11}B values (to −9.3 ± 1.5‰) of all the tektites and the impact glasses analyzed (Table 2). Such B contents and low δ^{11}B values can be present in some soils or loess sediments or in some sandstone sediments (Table 3). Indeed, quartzites (and shales) are present among the target rocks (Meisel et al., 1991). Sandstones and quartzites have bulk B contents between 5 and 70 ppm B (Harder, 1974) and although bulk boron isotope analyses are not available, their δ^{11}B values are probably intermediate between that of the detrital components (i.e., tourmalines from granitic rocks, clustering between −13 and −8‰, Table 3) and that of the clay-rich matrix (between −12‰ and −2‰, depending on the origin of the clay, Table 3). This is supported by the observation of tourmaline inclusions, which were found in some Darwin glasses (Reid and Cohen, 1962).

Mean Boron Isotope Composition of Upper Crustal Rocks

A general conclusion from previous geochemical studies was that tektites have, in general, a major and trace element composition that is very similar to that of the post-Archean upper continental crust (e.g., Taylor, 1973; Koeberl, 1986). However, the mean B contents and δ^{11}B values of the tektites and impact glasses (31 ± 26 ppm and −0.2 ± 5.0‰, respectively) are significantly different from those estimated for the average continental crust (10 ppm and between −8 and −13‰, Harder 1974; Chaussidon and Albarède, 1992). However, they are similar to those of continental sedimentary rocks, such as soils or river sediments (Spivack et al., 1987). Assuming that boron isotopes are not strongly fractionated during impact and melting of terrestrial rocks (see previous section), this difference could imply that the surface crustal rocks impacted are not representative of the "bulk continental crust" and have higher B contents and δ^{11}B values than deep crustal rocks. Indeed, the lower crust has very low B contents of 2 ± 1 ppm (Leeman et al., 1992), at variance values for continental and marine sedimentary rocks (Table 3). In addition, as the B depletion of the lower crust is being attributed to metamorphic and dehydration reactions, the δ^{11}B value of the lower crust is likely to be modified to lower values during this process, because of the preferential partitioning of ^{11}B compared to ^{10}B during fluid-rock interaction at low temperatures (Vengosh, 1992).

SUMMARY AND CONCLUSIONS

We used an ion microprobe to study the abundances of Li, Be, and B, as well as the boron isotope compositions, in twenty-seven tektite and impact glass samples. The tektite samples were from the Australasian, North American, and Ivory Coast strewn fields, while the impact glasses originated from the Aouelloul and Darwin craters. In addition, the internal variations of the abundance and isotopic composition of B in a flanged australite was studied, in order to place constraints on possible mass-dependent isotope fractionation.

Our main results and conclusions from these analyses are:

(1) We presented a detailed study of the abundances and isotopic composition of B in tektites, especially in different types of tektites found in the same strewn field. Also, we provided first results for internal variations of the distribution of B and changes of the δ^{11}B value. The results for our elemental abundance determinations are, for most samples, similar to bulk values that have previously been reported in the literature. However, we found a positive correlation between the Li and B and the B and Be abundances among the Muong Nong-type indochinite samples, which has not been recognized before.

(2) The B contents and isotopic compositions along several profiles across the rim of one flanged australite were also measured. A flanged tektite was subjected to a second intense heating pulse, and may, therefore, show vapor fractionation effects in tektite material. However, we found δ^{11}B variations of only a few permil, which show no significant correlation with the B contents or the distance from the rim of the flange. The mean δ^{11}B value for the flanged australite is indistinguishable from that of Muong-Nong type tektites (−1.9 ± 1.9‰). Thus, we conclude that vapor fractionation has not played a major role during tektite formation. It is, however, possible that surface material of the tektites was removed during atmospheric passage as the surface layers were heated and melted, erasing any traces of local isotopic fractionation.

(3) The B contents and the δ^{11}B values of the different samples from the Australasian tektite strewn field are not correlated with each other. This supports the conclusion that vapor fractionation was not an important process.

(4) The study of the B abundances and isotopic composition in tektites can be used to constrain the type of target rocks from which they have formed. We found that tektites show a rather limited range of δ^{11}B values (−9.3 ± 1.5‰, with an average of +2.7 ± 1.5‰). This range is small compared to the range observed for common terrestrial rocks (i.e., about −30 and +40‰). The B abundance and isotopic data can only be explained by terrestrial source rocks, as meteorites have very low B contents, and much more variable δ^{11}B values.

(5) Tektites from the Australasian strewn field have relatively high B and Li abundances and δ^{11}B values of (−4.9 to +1.4‰), which are consistent with values for pelagic and neritic sediments, as well as river and delta sediments data. Such clay-rich sediments were most likely among the Australasian tektite source rocks. However, ^{10}Be and Rb-Sr data clearly point to continental crustal source rocks, excluding pelagic, and probably, neritic sediments as a source. River and deltaic sediments, however, are continental crustal material with a higher clay content that have acquired B from the seawater and would satisfy the ^{10}Be, Rb-Sr, and B data. The Mekong river delta deposits large quantities of sediment in the shallow South China Sea, leading a fast deposition rate of ^{10}Be-rich sediments. Thus, an offshore source locality close to the coast of southeastern Indochina is favored by the present data.

(6) In contrast to other tektites, one bediasite sample has a very high δ^{11}B value of +15.1 ± 2.1‰, requiring the presence of marine carbonates or evaporites among the source rocks. This finding supports an offshore locality for the North American tektite source crater, such as those proposed at the Chesapeake Bay and the continental shelf off New Jersey.

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APPENDIX A: Major element compositions of the rim of australite USNM 3463 in wt% (1)

distance(2)(µm)	SiO_2	Al_2O_3	FeO(3)	MgO	CaO	Na_2O	K_2O	TiO_2	MnO	Total
650	69.04	14.11	5.15	2.37	3.42	1.20	2.41	0.73	0.11	98.54
800	68.87	14.45	4.88	2.22	3.46	1.12	2.46	0.87	0.10	98.43
900	69.20	14.44	5.23	2.34	3.74	1.04	2.29	0.79	0.08	99.15
1050	71.09	13.49	4.44	2.06	3.17	0.82	2.24	0.79	0.16	98.26
1100	69.05	14.37	4.65	2.34	3.52	1.20	2.50	0.75	0.00	98.38
1250	69.26	14.54	4.75	2.47	3.20	0.92	2.10	0.88	0.09	98.21
1300	69.36	14.07	5.05	2.41	3.50	0.95	2.53	0.85	0.10	98.82
1420	70.02	14.31	4.97	2.33	3.50	0.94	2.24	0.87	0.03	99.21
1500	70.49	14.17	4.48	2.23	3.24	0.86	2.14	0.85	0.11	98.57
1550	71.16	13.91	4.78	2.29	3.30	0.96	2.38	0.79	0.04	99.61
1590	69.37	13.87	5.27	2.53	3.53	0.98	2.12	0.74	0.17	98.58
1700	69.57	14.66	5.10	2.52	3.22	0.95	2.06	0.76	0.13	98.97
1900	70.81	13.89	4.50	2.17	3.31	0.97	2.51	0.83	0.07	99.06
1920	70.56	14.32	5.11	2.24	2.73	0.98	2.21	0.80	0.00	98.95
2100	70.21	14.36	4.86	2.36	3.39	0.92	2.19	0.86	0.08	99.23
2250	69.42	14.16	5.23	2.39	3.60	1.00	2.50	0.90	0.05	99.25
2270	70.01	14.15	4.77	2.32	3.35	0.87	2.18	0.72	0.01	98.38
2440	69.47	14.50	4.90	2.50	3.63	0.79	2.09	0.80	0.06	98.74
2480	69.86	14.30	4.87	2.31	3.54	1.06	2.42	0.82	0.02	99.20

(1) major element compositions analysed with a Cameca SX 50 electron microprobe at the Université de Nancy I (J. M. Claude, analyst).

(2) : distance in µm from a point taken as a reference (see Fig 2).

(3) : all Fe as FeO.

APPENDIX B : B contents and isotopic compositions of the rim of the australite USNM 3463

profile n°1 (1)		profile n°2		profile n°3		profile n°4	
d(2)(µm)	ppm B(3)	d (µm)	$\delta^{11}B$(‰)	d (µm)	$\delta^{11}B$(‰)	d (µm)	$\delta^{11}B$(‰)
100	24.4	1150	-1.6	735	-1.0	600	+1.6
200	23.6	1250	+1.7	820	+1.0	700	+0.4
300	23.8	1350	+3.0	905	+0.6	800	+0.8
400	24.4	1450	+0.2	990	+0.8	900	-1.9
500	25.2	1550	+1.3	1075	-4.3	1000	-0.1
600	25.5	1650	-0.7	1160	-0.6	1100	-0.4
700	29.8	1750	+2.7	1245	-6.0	1200	-0.7
800	28.2	1850	+0.0	1330	+6.1	1300	+1.6
900	28.5	1950	+4.2	1415	+5.3	1400	+0.6
1000	28.8	2050	+1.8	1500	-3.0	1500	+1.9
1100	30.1	2150	+1.5	1585	-4.6		
1200	27.0	2250	+4.1	1670	+1.1		
1300	27.4	2350	0.0	1755	+0.3		
1400	27.0	2450	+1.6	1840	+2.8		
1500	28.6			1925	-1.3		
1600	23.3			2010	-1.6		
1700	23.7			2095	+6.4		
1800	21.7			2180	+3.9		
1900	-			2265	-0.4		
2000	14.3			2350	-2.6		
2100	21.3						
2200	21.3						
2300	21.5						
2400	21.7						
2500	22.4						

(1) : the location of the four profiles in sample USNM 3463 is shown in Fig 2.

(2) : d = distance in µm from a point taken as a reference (see Fig 2).

(3) : the B contents are given for an average SiO_2 content of 69.83 wt%.