

Dolomite-apatite inclusion in chrome-diopside crystal, Bellsbank kimberlite, South Africa

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Abstract

A chrome-diopside crystal from the De Bryn and Martin mine of the Bellsbank kimberlite, South Africa, contains oriented mica inclusions and dolomite-rich blebs. One bleb contains a euhedral inclusion of apatite (F 1.4, Cl 0.9, SrO 0.9, total rare earths 1.5 wt.%). The dolomite blebs (FeO 0.7, MnO 0.6, SrO 0.45 wt.%) are chemically and texturally distinct from veins of fibrous dolomite (FeO 2.5 wt.%). Serpentine, barite (SrO 0.1-0.7 wt.%), chrome-spinel and pyrite also occur. We propose that H₂O-CO₂-bearing peridotite melted partially during diapiric uprise, and that diopside, phlogopite, Fe-poor dolomite, apatite, and perhaps chrome-spinel and barite crystallized simultaneously at a depth less than ~100 km. Differential expansion during transport to the surface in a kimberlite magma allowed veining by Fe-rich dolomite and partial alteration to serpentine.

Introduction

Although properties of magmas from the Earth's mantle can be interpreted in terms of experimental phase equilibria in the peridotite-H₂O-CO₂ system (summarized in Wyllie, 1979, 1980), and although micas and amphiboles with primary textural relationships to other minerals in peridotite xenoliths have been transported by kimberlites and alkali basalts, there is little direct evidence on carbonate minerals from the mantle. Dolomite or magnesite (Kushiro *et al.*, 1975) should be stable in pressure-temperature fields outlined by pyroxene thermobarometry (Boyd, 1973) of lherzolite xenoliths from kimberlites, but coarse grains of these Mg-rich carbonates in textural equilibrium with silicates have not been found. Many xenoliths contain calcite, especially in veins, but the textural relationships appear to have developed during and after transport to the surface. McGetchin and Besancon (1973) gave microprobe analyses of carbonate inclusions in two garnet megacrysts from diatremes in Arizona. We now describe an inclusion composed of dolomite, apatite, mica, Cr-spinel, barite, and serpentine in a Cr-diopside crystal from the Bellsbank kimberlite, S. Africa.

Specimen description

A 1 × 1 × 1.5 cm single crystal of Cr-diopside was collected by JVS from the concentrate at the De Bryn and Martin mine in the Bellsbank kimberlite during

the First International Conference on Kimberlites. Phlogopite crystals up to 3 × 0.5 mm occur with their basal cleavage in the (010) plane of the diopside. In polished thin section, dozens of rounded blebs of iron-poor dolomite lie randomly scattered throughout the diopside. Veins of iron-rich dolomite pervade the crystal, and pyrite and serpentine are present in them. Many dolomite blebs appear to have undergone secondary alteration, especially where a vein of Fe-rich dolomite intersects them or passes nearby. However, some appear to be almost unaffected, and emphasis is placed on one bleb (Fig. 1) which contains a euhedral prism of apatite (6 × 25 μm) surrounded by iron-poor dolomite. The periphery of the bleb is occupied mostly by serpentine with slightly curved cleavage lamellae. Other blebs contain oriented mica grains and small spinels typically <10 μm wide. One bleb is composed of barite with minor serpentine. Most carbonate is optically homogeneous but one polygranular dolomite-mica inclusion was found. It is impossible to give reliable modal proportions of the minerals in the blebs which appear to be primary, but dolomite amounts to about two-thirds, and phlogopite to about one-quarter. Apatite is rare, and a crystal large enough to analyze was found in only one bleb.

Electron microprobe analyses

Analyses (Table 1) were obtained by a combination of energy-dispersive and crystal spectrometer



Fig. 1. Photomicrograph of $6 \times 25 \mu\text{m}$ apatite crystal in a dolomite bleb, itself in a chrome-diopside megacryst. The apparent inclusion in the apatite crystal is also apatite. Serpentine borders the dolomite. Transmitted light, partially crossed polars.

techniques, distinguished respectively by either one or two figures after the decimal point. Operating conditions and standards for EDS analyses were listed in Smith *et al.* (1978), except that Ca and Mg analyses in dolomite were referred to a dolomite standard. Replicate analyses on the same spot showed no change of composition. Spectrometer analyses for minor elements in silicates and carbonates used a beam current of $0.7 \mu\text{A}$ and the following standards: Ti, Cr: Corning V glass; Mn, Ba: Corning W glass; Sr, Ni, Rb: Corning X glass; Al, Si, Ca, Na: Di85Jd15 glass; Fe: olivine P140; K: Asbestos microcline; F: fluorphlogopite; Cl: scapolite XXXIV. Detection levels are $\sim 0.02 \text{ wt.}\%$ oxide. A wide beam was used to avoid damage. Special conditions (Smith *et al.*, 1979) were used for Rb and Ba in mica. Apatite analyses were made at 25kV and $0.5 \mu\text{A}$. The above standards were used except for Ca, P, F from Durango apatite (Jarosewich *et al.*, 1979) and rare earths from glasses prepared by Drake and Weill (1973). A detection level of $\sim 0.03 \text{ wt.}\%$ oxide was obtained for the rare earths.

Mineral chemistry

The composition of the clinopyroxene crystal (Table 1) falls in the chrome-diopside group, which

is the most common type for lherzolite inclusions in kimberlite (Stephens and Dawson, 1977). Large single crystals of this group have been described only from kimberlite pipes in Colorado (Egler *et al.*, 1979) and Orapa, Botswana (Shee and Gurney, 1979). Analyses of the large oriented crystals of mica and of the mica in dolomite blebs agree within experimental error, and fall in the range for phlogopite megacrysts picked out of kimberlites (Dawson and Smith, 1975; Smith *et al.*, 1979). The Cl content (0.68 wt.%) is much higher than the mean value of 0.05 wt.% for phlogopite megacrysts (Smith *et al.*, 1980), but some megacrysts do have similar Cl. The chrome spinel has too much Fe to fit the trends of peridotitic or kimberlite groundmass spinels (Haggerty, 1979). Titanium is too high to match with analyses of spinels in kimberlite groundmass given in Smith *et al.* (1978).

The Fe-poor dolomite blebs (MgCO_3 , 50, CaCO_3 , 48, FeCO_3 , 0.9, MnCO_3 , 0.8, SrCO_3 , 0.4 mol.%) are compositionally similar to the dolomite (MgCO_3 , 46, CaCO_3 , 51, FeCO_3 , 2, SrCO_3 , 0.3) found by McGetchin and Besancon (1973) in a chrome pyrope megacryst from the Red Mesa kimberlite, Arizona. The

Table 1. Electron microprobe analyses of 73-113-li

	1	2	3	4	5	6	7	8
P_2O_5	n.a.	n.a.	n.a.	n.a.	n.a.	39.1	n.a.	n.a.
SiO_2	55.4	42.1	n.a.	0.04	0.24	0.9	44.5	43.6
TiO_2	0.23	1.06	5	0.00	0.00	n.a.	n.a.	n.a.
Al_2O_3	2.30	12.4	5	0.00	0.00	n.a.	1.0	1.2
Cr_2O_3	2.23	0.55	48	n.a.	n.a.	n.a.	n.a.	n.a.
FeO	2.26	3.88	30	0.73	2.5	n.a.	2.3	1.6
MnO	0.07	0.02	n.a.	0.62	0.41	n.a.	n.a.	n.a.
MgO	15.3	24.9	11	21.9	19.4	n.a.	38.8	39.2
NiO	0.04	0.20	n.a.	0.00	0.00	n.a.	n.a.	n.a.
CaO	20.8	0.02	n.a.	29.2	30.3	52.1	0.2	0.3
Na_2O	2.44	0.09	n.a.	0.03	0.03	n.a.	n.a.	n.a.
K_2O	n.a.	10.8	n.a.	0.00	0.00	n.a.	n.a.	0.2
F	n.a.	0.97	n.a.	n.a.	n.a.	1.41	n.a.	n.a.
Cl	n.a.	0.68	n.a.	n.a.	n.a.	0.87	n.a.	n.a.
Sum	101.07	97.25*	99	52.97	52.88	95.89*	86.8	86.1

1. Chrome-diopside.
2. Mica. Sum includes $\text{BaO} = 0.048 \text{ wt.}\%$ $\text{Rb}_2\text{O} = 0.090 \text{ wt.}\%$.
3. Chrome-spinel. Ti, Al and Mg vary by 20% on different grains.
4. Dolomite inclusions in clinopyroxene. Sum includes 0.45 wt.% SrO.
5. Dolomite vein cutting through clinopyroxene.
6. Apatite. Sum includes 0.86 wt.% SrO, 0.35 La_2O_3 , 0.64 Ce_2O_3 , 0.16 Pr_2O_3 , 0.19 Nd_2O_3 , 0.03 Sm_2O_3 , and 0.08 Gd_2O_3 .
- 7,8. Serpentine: associated with barite and dolomite blebs and as vermicular serpentine, respectively. Barite with 0.11-0.66 wt.% SrO and pyrite were also found.

* sum corrected for O=F, Cl
n.a. not analyzed

Red Mesa pyrope contained a clinopyroxene inclusion with higher TiO_2 and Al_2O_3 (0.64 and 4.0 wt.%, respectively) and lower Cr_2O_3 and Na_2O (1.3 and 2.0 wt.%) than the Bellsbank clinopyroxene. Dolomite in the veins contains three times as much FeO as dolomite in the blebs.

Apatite in the dolomite bleb is very much richer in Cl (0.87 wt.%) than apatite crystals (<0.01 wt.%) in secondary veins crossing peridotite xenoliths (Smith *et al.*, 1980). Other analyzed apatites from mantle-derived xenoliths have high Cl (~1 wt.%; Frey and Green, 1974; 0.4 wt.%; Wass, 1979).

Serpentine is a common alteration product in peridotites. Analyses (Table 1) show that serpentines associated with dolomite or barite blebs are richer in iron than vermicular serpentine. Both serpentine types are much poorer in iron than kimberlitic serpentine minerals reported in Smith *et al.* (1978).

Discussion

Although most of the dolomite blebs have ill-defined margins because of either interaction with Fe-rich dolomite veins or alteration to serpentine, a few blebs are almost intact. A primary origin for these blebs is supported by (a) the presence of a euhedral apatite crystal larger and much richer in Cl than tiny Cl-poor apatite crystals found in secondary veins in peridotite xenoliths, (b) the presence of mica crystals with the same composition as the large oriented inclusions, and (c) the lower Fe content than for vein dolomite.

Barite is commonly found in kimberlite groundmass (*e.g.*, Smith *et al.*, 1978) but has not been reported to occur in a peridotitic paragenesis. Because barite and dolomite blebs exhibit similar textural relationships with the other minerals, a primary origin for barite is possible.

Because the oriented mica inclusions have the same chemistry as mica megacrysts in kimberlites, and because the mica inclusions have much higher NiO and Cl than for secondary-textured micas (DeLaney *et al.*, 1980), the possibility of direct crystallization from a liquid at depth is explored. However, a thorough study of diopside megacrysts, especially from the Bellsbank kimberlite, is needed before the dolomite-bearing diopside crystal is assigned to a definite paragenesis. Marginal alteration of blebs can be explained by solutions percolating into microcracks caused by differential expansion of host diopside and guest inclusions during ascent from considerable depth.

The absence of orthopyroxene or garnet precludes

estimation of temperature or pressure by the procedures used for lherzolites, but limited experimental data on carbonate-silicate reactions may provide constraints. Eggler *et al.* (1976) calculated the position of the reaction diopside + dolomite = calcite + forsterite + vapor, which limits the stability of the left-hand side of the reaction to 1230° at 25 kbar, 1000° at 15 kbar and 730°C at 5 kbar. Lowering the activity of CO_2 would decrease the thermal stability of the carbonate-pyroxene assemblage. The effect of adding just phlogopite to this assemblage has not been experimentally determined, but Eggler and Wendlandt (1979) showed that dolomite-phlogopite-lherzolite is stable at 30 kbar for temperatures less than ~1100°C over a wide range of vapor composition, and Wyllie (1980; his Fig. 2) presented a schematic field of stability for dolomite-phlogopite-peridotite and buffered vapor with a maximum limit between 40 kbar, 1100° and 27 kbar, 1200° and a minimum between ~30 kbar, 800° and 18 kbar, 1000°C.

To conclude, we suggest that partial melting of H_2O , CO_2 -bearing peridotite occurred during diapiric uprise of mantle (Wyllie, 1980; his Fig. 8), and that low-Fe dolomite, phlogopite, diopside, apatite, and barite(?) precipitated from a partial melt at a depth less than ~100 km. Serpentinization and permeation by Fe-rich dolomite occurred as a result of transport by kimberlite to the Earth's surface. Perhaps differential expansion between carbonate and silicate minerals causes fracturing of carbonate-peridotite xenoliths during ascent, and perhaps only a few carbonate inclusions in silicate minerals and diamond will survive to provide information on the carbonate mineralogy of the upper mantle. The electron microprobe analyses demonstrate that both dolomite and apatite, and perhaps barite, must be considered as reservoirs for strontium in the upper mantle, in addition to diopside and mica, and study of implications for magma generation and geochemical models is in progress. Ion microprobe analyses of trace elements will be made when appropriate calibrations have been obtained.

Acknowledgments

We thank I. Baltuska, R. Draus, O. Draughn, and I. M. Steele for technical help, and the National Science Foundation for grant EAR 77-02711. Helpful criticism from A. L. Boettcher, F. R. Boyd, D. H. Eggler, D. Smith, S. Y. Wass, P. J. Wyllie, and H. S. Yoder is appreciated. This study was made possible by the hard work of the organizers of the First International Kimberlite Conference.

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*Manuscript received, May 27, 1980;
accepted for publication, September 29, 1980.*