

a section, the herring-bone pattern may be ascribed to the superposition of fibrils.

Spherulites sectioned end-on show the fibril bundles with rectangular outlines and arranged in oblong groups. Within the groups the bundles are parallel oriented but adjacent groups are sub-parallel.

The individual fibrils could not be resolved in the microscope; their presence was indicated by their spherulitic extinction pattern. The fibrillar growth axis is  $c$  (001). The crust surfaces of the spherulites are covered with the blunt ends of protruding fibril bundles. The gravity measured in suspension in  $\text{CHBr}_3/\text{CCl}_4/\text{CH}_2\text{S}_2$  is 2.917 (1)  $\text{g}/\text{cm}^3$  (X-ray density 2.919  $\text{g}/\text{cm}^3$ ). Infra-red spectra were obtained with a Perkin-Elmer model 530 double-beam spectrophotometer using 1 mg samples in 350 mg KBr tablets. Spectra of white, pale green, and pale yellow-green prehnite were obtained in the range 200–4000  $\text{cm}^{-1}$ . The spectra exhibited no significant differences and are identical to the spectrum of prehnite given by Moenke (1962). Owing to the fibrillar nature of the prehnite no attempt was made to determine the refractive indices by immersion methods. However, Tröger (1959) has determined the optical properties of prehnite as a function of the  $\text{Fe}^{3+}$  mol. %, since a partial chemical analysis of the greenish-yellow prehnite gave  $\text{Ca}^{2+} = 19.4$  wt. %,  $\text{Na}^+ = \text{K}^+ = \text{Li}^+ = 0.01$  wt. % and  $\text{Fe}^{3+} = 2.0$  wt. %, the optical properties could be calculated on the basis of the Fe content. The results are  $\alpha$  1.627,  $\beta$  1.638,  $\gamma$  1.663,  $\text{SV}_y$  67°, and  $\Delta$  0.036.

*X-ray data.* A preliminary unit cell was derived

from single-crystal photographs. The refinement was done by least squares using the program PARAM (Steward *et al.*, 1976). The powder data are in agreement with the data for prehnite, PDF no. 7-333. The unit cell data are  $a$  4.628(1),  $b$  5.485(1),  $c$  18.478(4) Å, and cell volume 469.056 Å<sup>3</sup>.

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## A note on the occurrence of melilite in kimberlites and olivine melilitites

OLIVINE melilitites occur together with kimberlites in Siberia (Ukhanov, 1963) and Namaqualand, South Africa (Moore, 1976). Such field associations, coupled with the alkaline ultrabasic characteristics of the two rock types, suggest a close although as yet poorly understood genetic relationship. This is supported by striking similarities in the phase assemblages of kimberlites and olivine melilitites (Moore and Erlank, 1979). There are, however, a

number of marked differences in the mineral assemblages in the two rock types, including the rarity of melilite in kimberlites. It is suggested that the presence or absence of this mineral in under-saturated ultrabasic rocks can in part be understood in terms of melilite phase relations.

Representative analyses of melilitites from the Namaqualand–Bushmanland olivine melilitites are given in Table I. Melilitites from these pipes are

TABLE I. Representative microprobe analyses of melilites from the Namaqualand-Bushmanland volcanics

	Bies-1 melilite centre	Bies-1 melilite edge	ZH-14 melilite centre	ZH-14 melilite edge
SiO <sub>2</sub>	43.60	43.77	44.10	44.05
Al <sub>2</sub> O <sub>3</sub>	4.98	5.21	4.44	4.32
FeO	4.29	4.75	4.38	4.55
MnO	0.07	0.08	0.07	0.05
MgO	9.00	8.43	9.10	9.14
CaO	35.62	34.32	34.30	34.19
Na <sub>2</sub> O	2.69	3.25	3.37	3.37
K <sub>2</sub> O	0.17	0.16	0.16	0.15
Total	100.41	99.98	99.94	99.83
Number of cations for twenty-one oxygens				
Si	5.947	5.993	6.035	6.040
Al	0.802	0.841	0.717	0.698
Fe	0.489	0.544	0.502	0.522
Mn	0.008	0.009	0.008	0.006
Mg	1.829	1.720	1.856	1.860
Ca	5.206	5.035	5.029	5.024
Na	0.711	0.862	0.895	0.897
K	0.030	0.028	0.028	0.026
Mg/(Mg + Fe)	0.79	0.76	0.79	0.78
End-member proportions	Ge <sub>0.009</sub> Ak <sub>0.762</sub> SM <sub>0.228</sub>	Ge <sub>0.001</sub> Ak <sub>0.733</sub> SM <sub>0.266</sub>	Ge <sub>0.000</sub> Ak <sub>0.768</sub> SM <sub>0.232</sub>	Ge <sub>0.000</sub> Ak <sub>0.774</sub> SM <sub>0.226</sub>

similar to those in other alkaline volcanics in plotting close to the sodium melilite-akermanite join (Yoder, 1973).

Moore and Erlank (1979) pointed out that melilite is unstable at the solidus temperatures of basic igneous rocks, although it may be metastably preserved in volcanic rocks if cooling is rapid. However, if originally present, the phase would be partially or wholly replaced in magmas undergoing slow cooling. The marginal replacement of melilite by clinopyroxenes in the Namaqualand olivine melilitites is evidence of this reaction.

Kimberlites would be expected to cool rapidly as a result of the adiabatic cooling of an expanding gaseous phase during fluidized emplacement in the upper crust, and it is likely that the absence of melilite in these rocks results not from subsolidus decomposition of the phase but from non-crystallization. It is thus necessary to consider why melilite is generally an unstable phase in kimberlite magmas.

Pure akermanite is stable to 14 kbar under anhydrous conditions (Kushiro, 1964) (fig. 1) but stability is lowered to 10.2 kbar in the presence of excess H<sub>2</sub>O, and 8.7 kbar in the presence of excess CO<sub>2</sub> (fig. 2). Kimberlitic magmas are typically carbonate-rich, and melilite crystallization will therefore probably be inhibited in such liquids. Moreover, development of a fluidized system at low pressures as a result of expansion of a vapour phase would be expected to cause marked adia-

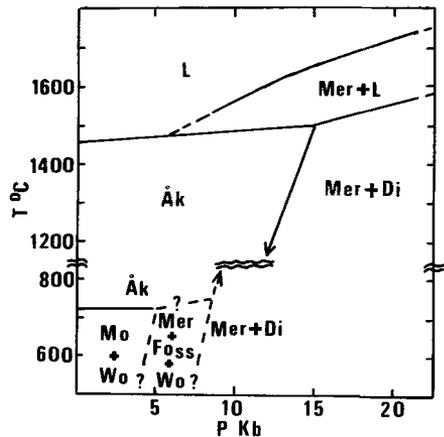


FIG. 1. Pressure-temperature plot of stability field of akermanite (Ak). Note that akermanite stability is limited to pressures less than 15 kbar under anhydrous conditions. The mineral is unstable at low temperatures and pressures. Mer = merwinite; Wo = wollastonite; Fo = fosterite; Mo = monticellite; L = liquid (after Kushiro, 1964).

batic cooling of the magma, leading to low emplacement temperatures—possibly as low as 350°C (McFadden and Jones, 1977). Kimberlite magmas would therefore follow a late-stage *P-T* path that is unusual for volcanic rocks and may never intersect the melilite stability field, as illustrated in fig. 2.

This suggests that the absence of melilite in kimberlites is not controlled by chemical factors alone, and that the phase could crystallize from these magmas under favourable  $P$ - $T$  conditions. It would be interesting to test this possibility experimentally, as this may provide further constraints on the late-stage  $P$ - $T$  histories of kimberlites.

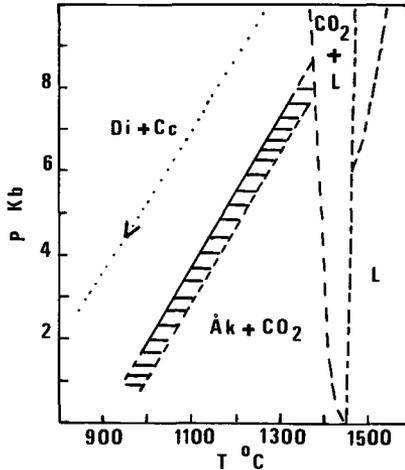


FIG. 2. Akermanite- $\text{CO}_2$  system (excess  $\text{CO}_2$ ). Cc = calcite. Other abbreviations as for fig. 1 (after Yoder, 1973). Note that akermanite stability is limited to pressures below 8.7 kbar under  $\text{CO}_2$ -saturated conditions. Dotted line shows a hypothetical late-stage  $P$ - $T$  path of a kimberlite magma.

The phase relations provide a framework for understanding variations in groundmass mineralogy in kimberlites. Skinner and Clement (1979) classify kimberlites on the abundance of five primary groundmass minerals—diopside, monticellite, phlogopite, calcite, and serpentine. The abundance of phlogopite is related to chemical factors such as the potassium concentration (M. J. Skinner, pers. comm.), while the abundance of calcite may be in part related to the amount of  $\text{CO}_2$

loss associated with fluidization. However, the abundance of diopside and monticellite will be controlled by the late-stage  $T$ - $P$  ( $-\text{CO}_2$ - $\text{H}_2\text{O}$ ) path relative to the monticellite and diopside phase fields. The same will apply to the occurrence of primary serpentine, which is only stable at temperatures below  $500^\circ\text{C}$  (Bowen and Tuttle, 1949). The groundmass mineralogy of kimberlites could therefore be potentially used to monitor the late-stage history of the kimberlite magma.

Moore and Erlank (1979) suggested that olivine melilitite magmas lose an immiscible carbonate liquid during ascent. As a result, they do not evolve a vapour phase and undergo the low-pressure adiabatic cooling which characterize kimberlites. Their late-stage  $T$ - $P$  paths, therefore, might not intersect the stability fields of minerals such as monticellite or serpentine, thus accounting for the rarity of these minerals in olivine melilitites.

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