

## Formation of orthopyroxene–Fe–Ti oxide symplectites in Precambrian intrusives, Rogaland, southwestern Norway

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### ABSTRACT

Petrographic and mineral chemical data are presented for orthopyroxene–Fe–Ti oxide symplectites in three intrusive bodies from southwestern Norway: the Bjerkreim-Sokndal lopolith, the Hydra anorthosite, and the Lyngdal hyperite. Detailed studies of the Bjerkreim-Sokndal lopolith indicate that symplectites are not common, occurring in only six thin sections out of 1037 studied. Only one of these samples contains olivine, and the symplectites clearly do not replace or form pseudomorphs after olivine. The symplectites in the Hydra anorthosite and Lyngdal hyperite do appear to replace olivine. The proportions of orthopyroxene and Fe–Ti oxide in the symplectites are fairly constant and range from 73:27 to 66:34 (vol%). Microprobe data demonstrate that the orthopyroxenes do not have compositions appropriate for crystallization at or above the solidus. Many of the pyroxenes in these intrusives have re-equilibrated during subsolidus cooling, but the orthopyroxenes in the symplectites show no exsolution and thus must have formed subsolidus. Pyroxene thermometry indicates temperatures of formation of 700–800 °C at estimated pressures of 5 kbar or less (upper amphibolite–granulite facies conditions). The Fe–Ti oxide in most of the symplectites is titaniferous magnetite that shows complex exsolution phenomena indicative of extensive re-equilibration during subsolidus cooling. It is virtually impossible to reconstruct the original compositions or to apply oxide thermometry.

The symplectites do not result from discontinuous precipitation from a supersaturated solid solution, eutectoidal breakdown of a pre-existing phase or isochemical replacement of olivine. It is suggested that the occurrence of symplectites reflects the overall re-equilibration of the original magmatic mineral assemblage during high-grade retrogressive metamorphism. They form only in the vicinity of orthomagmatic Fe–Ti oxide. The precise mechanism of formation is obscure, but appears to involve co-operative nucleation, predominantly at grain boundaries, and sympathetic growth via a bridging mechanism. The development of oriented intergrowths is governed by the requirements of minimal interfacial and strain energy. Diffusion mainly occurs along grain boundaries or advancing interfaces and is probably aided by a thin film of intergranular fluid. The supercooling necessary for symplectites to form probably reflects the high energy barriers to solid-state nucleation and growth.

### INTRODUCTION

Fine-grained intergrowths (symplectites) are relatively common in intrusive igneous rocks, but there is still much controversy about their origin. This is especially true in the case of orthopyroxene–opaque oxide (magnetite and/or ilmenite) symplectites that occur in gabbroic–noritic plutonic and hypabyssal rocks. In essence, the controversy centers upon the question of whether such symplectites form by subsolidus processes or whether they form directly from a melt by coprecipitation.

That symplectitic intergrowths can form by subsolidus processes is irrefutably demonstrated by their occurrence in metamorphic rocks, and convincing arguments have

been presented that the orthopyroxene–opaque oxide symplectites in some plutonic rocks formed after solidification was completed (Frodesen, 1968; Starmer, 1969; Esbensen, 1978; Van Lamoen, 1979). These arguments are largely based upon petrographic or textural evidence such as the tendency for the intergrowths to replace olivine in reaction coronas and their occurrence in rocks that contain ample other evidence for subsolidus re-equilibration. Attempts to model the formation of the symplectites quantitatively (see especially Van Lamoen, 1979) indicate that this is not a simple isochemical process, but the consequences of this conclusion have not, to date, been adequately explored.

It is extremely difficult to prove that orthopyroxene–

opaque oxide symplectites in gabbroic-noritic intrusions formed by coprecipitation from a melt. The main reason for this is that the long cooling histories of relatively large magma bodies, coupled with the release of volatiles in the last stages of solidification (see Morse, 1980, p. 328) and the possible access of circulating meteoric water (Taylor and Forester, 1979), allow continuous chemical and textural re-equilibration of the original mineral assemblage. It may thus be virtually impossible to distinguish between minerals and textures formed by crystallization in residual melt pockets and those formed later in the presence of deuteric and/or meteoric fluids.

The particular model adopted for the formation of symplectites can profoundly influence interpretation of the cooling histories of plutonic rocks and hence influence interpretation of the thermal history of the crust. For example, experience gained by metallurgical engineers suggests that symplectites can only grow from melts if the latter are supercooled so that if a supersolidus origin for symplectites is adopted, an explanation of how supercooling develops in magmas crystallizing at depth within the crust is required. Constitutional supercooling (Chalmers, 1964; Chadwick, 1972) could explain symplectite formation, but it is necessary to demonstrate that constitutional supercooling indeed occurs during crystallization of magmas in the plutonic environment. If the symplectites form by subsolidus reactions, then given diffusion data for relevant components, together with estimates of the conditions of formation, it should be possible to estimate the length of time required for formation. Clearly, however, it is first necessary to examine models for the origin of symplectites, and this is the primary goal of the present paper.

In this paper we report the results of a detailed investigation into the occurrence and mineral chemistry of orthopyroxene-magnetite-ilmenite symplectites in intrusive Proterozoic igneous rocks from Rogaland and Vest Agder, southwestern Norway. Initially, we concentrated upon the symplectites in the Bjerkreim-Sokndal lopolith, and most of the data presented are for samples from this intrusion. Subsequently, we have also studied symplectites in the Hydra anorthosite massifs and in the hyperites (gabbro-norites) from the Lyngdal area, and we include observations on samples from these intrusions. All of the rocks studied have re-equilibrated to a greater or lesser degree during prolonged subsolidus cooling.

### The Bjerkreim-Sokndal lopolith

The Bjerkreim-Sokndal lopolith is a large layered intrusion (total volume unknown) associated with anorthosite massifs (Fig. 1). Geochronological studies of the lopolith and the surrounding metamorphic envelope indicate intrusion ages between ~1050 m.y. and ~950 m.y. (Pasteels et al., 1979; Priem, 1980; Wielens et al., 1981). The spread in age is thought to be real and to reflect a complex intrusive history.

In this paper we give only a brief summary of the geology of the Bjerkreim-Sokndal intrusion. A more de-

tailed description, including new observations and data, will be presented elsewhere (Voncken and Barton, in prep.). The lower part of the intrusion (hereafter referred to as the anorthositic-leuconoritic phase) consists of at least five major rhythmic units, each of which has anorthosite at the base and grades upward via leuconorite to norite (the latter is not present in the two lowermost rhythmic units). The five rhythmic units are thought to result from fractional crystallization with superimposed magma mixing (Duchesne, 1972a; Hermans et al., 1975; Duchesne and Demaiffe, 1978). The rocks of the lower rhythmic units (RI, RII, and the lower parts of RIII and RIV) consist essentially of plagioclase, orthopyroxene, and ilmenite. Clinopyroxene, titaniferous magnetite, and apatite appear as major components in the upper parts of RIII and RIV, whereas K-feldspar initially appears as a component of antiperthite in RV. Olivine occurs only sporadically near the base of RIV, in what appears to represent a separate, later intrusion.

The upper part of the intrusion is mangeritic to quartz mangeritic (P. Michot, 1960, 1968; J. Michot, 1960; J. Michot and P. Michot, 1968) or pyroxene monzonitic to syenitic (Hermans et al., 1975) in composition. In the present paper we refer to the upper part of the intrusion as the quartz monzonitic phase (cf. Rietmeijer, 1979). Duchesne (1972a, 1978) and Duchesne and Demaiffe (1978) have proposed that this phase represents the residual liquid, possibly contaminated with crustal material (Pasteels et al., 1979), from crystallization of the lower layered series. Rietmeijer (1979), however, has found evidence for a structural discontinuity between the upper and lower parts of the lopolith that is reflected in the chemistry of the pyroxenes. This discontinuity suggests a time gap between the intrusion of the anorthositic-leuconoritic phase and the quartz monzonitic phase that is difficult, though not impossible, to reconcile with a model of continuous magmatic differentiation as envisioned by Duchesne (1972a, 1978) and Duchesne and Demaiffe (1978).

### Distribution of orthopyroxene-magnetite-ilmenite symplectites in the Bjerkreim-Sokndal lopolith

Representative samples were examined of all of the major rock types occurring from the base to the top of the lopolith. Out of a total of 1037 thin sections studied, symplectites were found in only six. Five of the six samples were from the uppermost part of RV, the sixth was from an olivine-bearing sample near the base of RIV. Thus, with one exception, the symplectites occur at a similar stratigraphic height and the exception occurs in what appears to be a separate, later intrusion.

### PETROGRAPHIC SUMMARY

The rocks in which the symplectites occur are mostly monzonorites or biotite norites (top of RV). The olivine-bearing sample is a leuconorite. The textures of these rocks are highly variable, even within a single thin section, but are mostly granular or slightly porphyritic or porphyro-

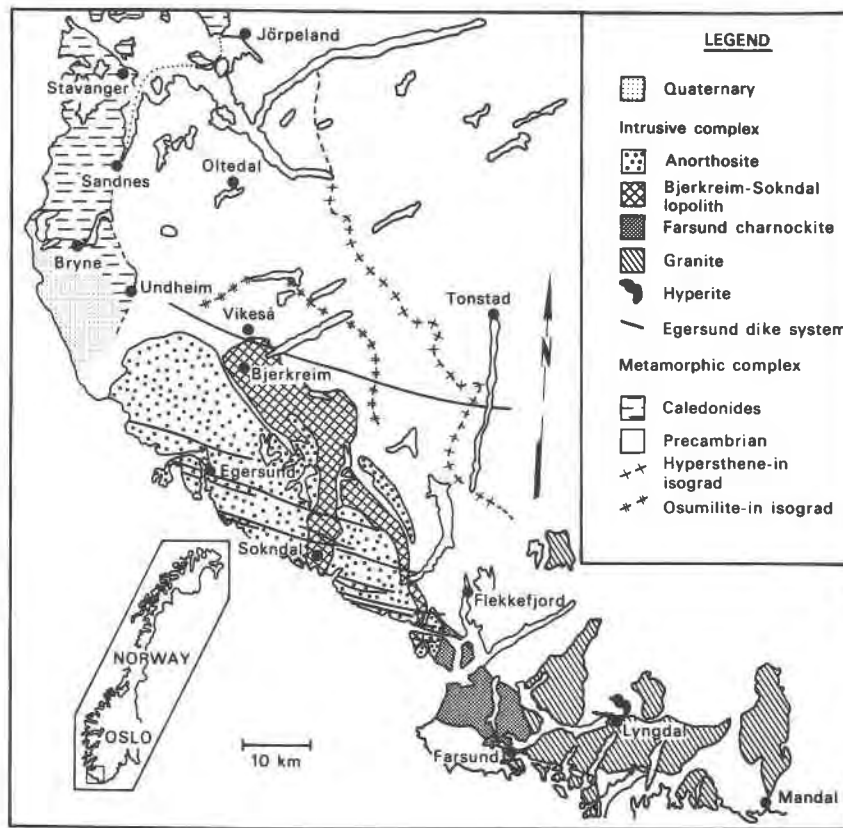


Fig. 1. Geologic map of Rogaland and Vest Agder showing the distribution of the major rock types. The positions of the hypersthene-in and osumilite-in isograds are taken from Hermans et al. (1975). The Hydra anorthosite is the small intrusive SSW of Flekkefjord. Hyperite occurs also to the south of Lyngdal in a small intrusive body that is not shown on the map.

clastic owing to the occurrence of large (up to 4 mm) anhedral feldspar crystals set in a finer-grained matrix. Cumulate textures are absent. In all samples there is evidence of deformation, i.e., the minerals show undulose extinction, strain lamellae, kink bands, and the development of subgrain boundaries. Some samples show granulation.

Plagioclase is the dominant mineral (30–70 vol%) and shows minor zoning (<4% An). The range in plagioclase composition for all samples is  $An_{52}$  to  $An_{29}$ . In two of the monzonorites (R186 and R486), the plagioclase shows complex exsolution phenomena and intergrowth forms. In addition to antiperthite (not present in all samples), patch-perthite and irregular intergrowths of plagioclase and alkali feldspar occur. The feldspars have not been studied in detail, but all stages of exsolution—from antiperthite through patch-perthite to complete segregation of two phases—are present.

Orthopyroxene (10–30%) occurs in various forms: in intergrowths with oxide minerals (described more fully below), as relatively large (mostly <1 mm, rarely up to 5 mm) subhedral crystals that show exsolution and are rimmed by clinopyroxene, as rims around clinopyroxene and olivine and as small granular crystals associated with nonexsolved equigranular grains of feldspar.

Inverted pigeonite was identified as an accessory phase in only one sample (R486).

All samples contain 5–15% oxides (titaniferous magnetite and ilmenite) as discrete grains or as inclusions in pyroxenes and as components of symplectites. The oxides show complex microstructures that undoubtedly reflect subsolidus re-equilibration.

Clinopyroxene occurs in five samples; it is absent from one of the norites (R195). It occurs as subhedral crystals ( $\leq 1$  mm) that show exsolution and are rimmed by orthopyroxene, as rims around orthopyroxene crystals, and as small granular crystals (cf. description of orthopyroxene). In general, the exsolution shown by the larger subhedral pyroxenes is similar to that described in detail by Rietmeijer (1979) and Rietmeijer and Champness (1982) for other samples from the Bjerkreim-Sokndal lopolith and need not be further described here beyond noting that exsolution is occasionally bleblike and irregular and that it may coarsen toward and coalesce with an external rim of exsolved material (cf. Bohlen and Essene, 1978). The pyroxenes also contain platelets of ilmenite oriented along the (010) and (001) planes.

Other, less abundant, minerals include olivine, apatite, spinel, sulfides, biotite, amphibole, and K-feldspar. In the one sample in which it occurs (GA248), olivine ( $Fe_{70}$ ) is

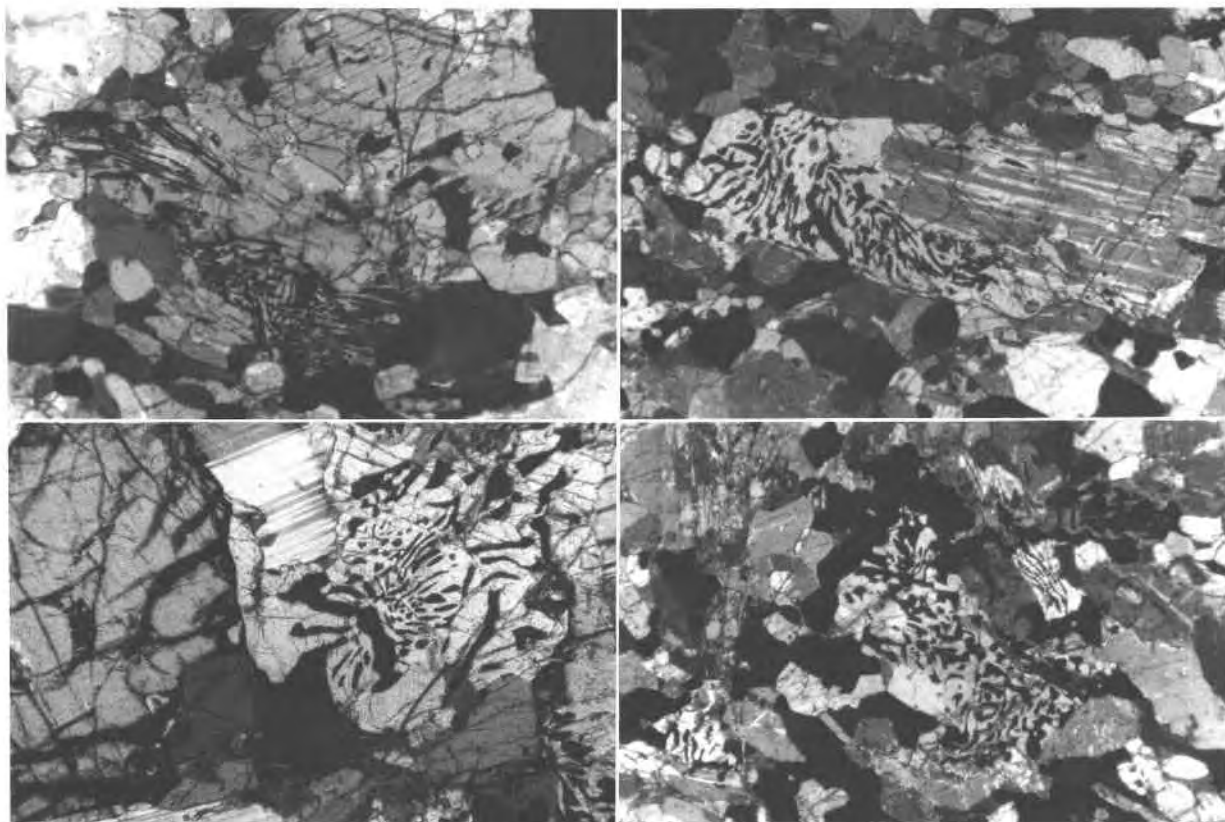


Fig. 2. (Upper left) Irregular symplectite. The irregular outline suggests that this intergrowth is not a pseudomorph after olivine. (Upper right) A rim of symplectite around clinopyroxene. (Lower left) Symplectite replacing olivine in the Hydra anorthosite. (Lower right) Association of symplectite and orthomagmatic Fe-Ti oxide. Field of view in all photographs is  $1.4 \times 2.8$  mm.

rounded with irregular, resorbed margins and is rimmed by orthopyroxene. Spinel occurs as discrete grains or lamellae in titaniferous magnetite. Sulfides occur in globules suggestive of formation by liquid immiscibility. Biotite and amphibole rim discrete grains of Fe-Ti oxide and pyroxene and also form small, euhedral crystals. K-feldspar is found as discrete, polygonal crystals and in complex intergrowths with plagioclase. Both sanidine and orthoclase are present.

#### Description of the symplectites

The symplectites are irregular in shape and in most cases do not appear to be pseudomorphs (Fig. 2a). They form small areas in the core of an orthopyroxene crystal or complete areas 1–2 mm in diameter. They sometimes occur as rims around clinopyroxene (Fig. 2b). Two types of intergrowth occur: lamellar and vermicular, the former commonly grading outward into the latter. The width of the Fe-Ti oxide component is mostly 5–10  $\mu\text{m}$  but can reach 30  $\mu\text{m}$ . No grain boundary is observed between discrete grains and symplectitic Fe-Ti oxide; optical orientation and microstructures are continuous between the two. Evidently, the symplectitic and discrete Fe-Ti oxides have re-equilibrated to a similar degree. The orthopyroxene component of the symplectites contains no exsolution lamellae and is identical to the orthopyroxene that

forms rims around olivine and clinopyroxene or occurs as discrete granules.

The Fe-Ti oxide component of the lamellar symplectites often appears to be oriented such that its (111) plane is parallel to the (100) plane of the associated pyroxene. Detailed studies show, however, that these intergrowths are irregular in three dimensions and that the Fe-Ti oxide lamellae are not continuous along their length. The Fe-Ti oxide in the vermicular symplectites is occasionally skeletal. In some cases, the Fe-Ti oxides radiate from a small apatite crystal. Exsolution textures in the Fe-Ti oxides are complex. In all but one sample (R195) the Fe-Ti oxide in the symplectites is predominantly titaniferous magnetite with thin exsolution lamellae of ilmenite oriented parallel to (111) and extremely small lenses of spinel oriented parallel to (100) and (111). Within a single titaniferous magnetite vermicule, 2–3 generations of ilmenite lamellae and 1–2 generations of spinel lenses can be recognized. Microstructures in the  $\text{Fe}_2\text{O}_3$ -rich ilmenites are more simple. Occasional spinel lenses are observed, and in one sample (R195) abundant lamella of hematite occur.

The proportions of orthopyroxene and Fe-Ti oxide in the symplectites has been determined by line counting. For each sample, 3–5 areas were measured, and over 4000 lines were counted for each area. The precision of the

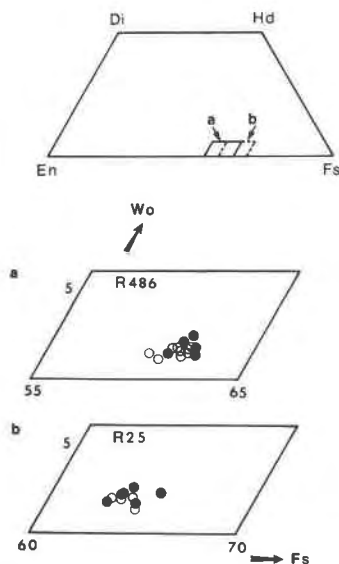


Fig. 3. Comparison of the compositions of symplectitic (open circles) and nonsymplectitic (filled circles) in two samples from the Bjerkreim-Sokndal lopolith.

measurements varies from 1% to 5% (relative). The proportions of orthopyroxene and Fe-Ti oxide range from 73:27 to 66:34 (vol%), i.e., the proportions of the phases are relatively constant.

In most cases, the symplectites are perfectly fresh, but occasionally they are replaced by other phases—mainly biotite and brown pleochroic amphibole. In some cases, the original symplectite is now represented by a relict amphibole-magnetite or biotite-magnetite symplectite.

#### Symplectites in the Hydra anorthosite and Lyngdal hyperite

Symplectites are not common in the Hydra anorthosite or in the Lyngdal hyperite. In the former, they occur only

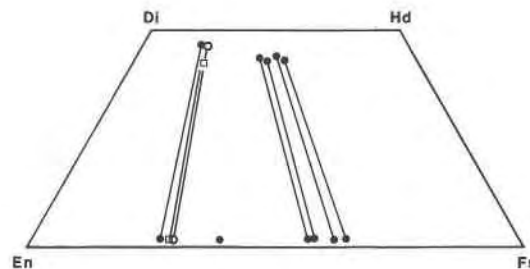


Fig. 4. Compositions of pyroxenes in the symplectite-bearing samples plotted in the conventional quadrilateral. Solid symbols = samples from the Bjerkreim-Sokndal lopolith; open circles = sample from the Hydra anorthosite; squares = sample from the Lyngdal hyperite.

in rare olivine-bearing leuconorites or in monzonorites close to the margins of the intrusion (H. P. Drent, oral comm., 1982) whereas in the latter they have been found in rare olivine-bearing biotite norites. Most of the symplectites differ from those in the Bjerkreim-Sokndal lopolith in that they clearly replace or, in extreme cases, form pseudomorphs after olivine (Fig. 2c). In those cases where replacement is incomplete, it is apparent that the development of symplectites is restricted to the vicinity of orthomagmatic Fe-Ti oxide (Fig. 2d). Where grains of orthomagmatic oxide are absent, the olivine is corroded and is rimmed by orthopyroxene. In other respects, however, the symplectites (and also the discrete pyroxenes, Fe-Ti oxides, etc.) in these intrusions are similar to those in the Bjerkreim-Sokndal lopolith. In addition to the orthopyroxene-Fe-Ti oxide symplectites, the hyperites contain biotite-orthopyroxene and clinopyroxene-albite intergrowths.

#### Mineral chemistry

All analyses were obtained using conventional wavelength- and energy-dispersive electron-microprobe tech-

TABLE 1. Average compositions of pyroxenes in the symplectite-bearing samples

	GA248		R195		R53		R186		R486	
	opx	cpx	opx	opx	cpx	opx	cpx	opx	cpx	
SiO <sub>2</sub>	52.8	51.6	52.0	51.0	51.2	49.7	50.7	50.2	51.5	
TiO <sub>2</sub>	0.15	0.60	0.18	n.d.	0.20	0.10	0.10	n.d.	0.51	
Al <sub>2</sub> O <sub>3</sub>	2.55	3.63	1.40	0.52	1.35	0.88	1.88	0.43	1.30	
FeO*	16.8	7.08	23.2	33.2	15.9	33.4	15.9	36.2	16.8	
MgO	26.2	14.4	21.5	14.4	10.6	14.2	10.4	12.5	9.3	
MnO	0.28	0.15	0.40	0.64	0.31	0.73	0.58	0.60	0.37	
CaO	0.65	22.7	0.53	0.93	20.3	1.03	20.2	0.78	21.1	
Na <sub>2</sub> O	n.d.	0.68	n.d.	n.d.	0.40	n.d.	n.d.	n.d.	n.d.	
Total	99.4	100.8	99.2	100.7	100.3	100.0	99.8	100.7	100.9	
Number of cations on the basis of six oxygens										
Si	1.927	1.899	1.961	1.991	1.962	1.963	1.953	1.989	1.970	
Ti	0.004	0.017	0.005	n.d.	0.006	0.003	0.003	n.d.	0.015	
Al	0.110	0.157	0.062	0.024	0.061	0.041	0.085	0.020	0.059	
Fe*	0.513	0.218	0.732	1.084	0.510	1.103	0.512	1.200	0.537	
Mg	1.426	0.790	1.209	0.838	0.605	0.836	0.597	0.738	0.530	
Mn	0.009	0.005	0.013	0.021	0.010	0.024	0.019	0.020	0.012	
Ca	0.025	0.895	0.021	0.039	0.833	0.044	0.833	0.033	0.864	
Na	n.d.	0.049	n.d.	n.d.	0.030	n.d.	n.d.	n.d.	n.d.	
Total	4.014	4.030	4.003	3.997	4.017	4.014	4.002	4.001	3.987	

Note: GA249, R195, R53, R186, R486, R25—Bjerkreim-Sokndal lopolith; H1—Hydra anorthosite; L1—Lyngdal hyperite; n.d. = not detected.

\* Total Fe reported as FeO or Fe<sup>2+</sup>.

niques and are fully corrected. Most analyses were performed with a focused (0.5 to 2  $\mu\text{m}$ ) electron beam, but a broad, defocused spot (>50- $\mu\text{m}$  diameter) was used to obtain bulk analyses of some symplectites and exsolved pyroxenes.

In each sample, the orthopyroxenes show a restricted range of composition, with variations of the order of 3 mol% En and 1.5 mol% Wo. There are no systematic compositional differences between the orthopyroxene in the symplectites and the host orthopyroxene in large grains of low-Ca pyroxene, the orthopyroxene that forms rims on clinopyroxene or olivine, and the discrete granular orthopyroxene (Fig. 3). Likewise, most *clinopyroxenes* in each sample show a restricted compositional range, corresponding to about 3 mol% Wo and 2 mol% En, and for this reason, average pyroxene compositions are given in Table 1 and are plotted in the conventional Ca-Mg-Fe diagram in Figure 4.

There are, however, significant differences between pyroxenes analyzed with a focused beam and those analyzed with a broad beam. Reintegrated analyses of pyroxenes from the Bjerkreim-Sokndal lopolith will be reported elsewhere, but from Figure 4 it is obvious that the low-Ca pyroxenes showing exsolution were originally much richer in Ca than the symplectitic low-Ca pyroxenes and also that the high-Ca pyroxenes that show exsolution were originally much poorer in Ca than the clinopyroxene that occurs as granular crystals and rims around the larger orthopyroxene crystals (Fig. 4).

Considering all samples, both the orthopyroxenes and the clinopyroxenes show a wide range in  $\text{Mg}/(\text{Mg} + \Sigma\text{Fe}^{2+})$ ,<sup>1</sup> from 0.73 to 0.53 and 0.78 to 0.44, respectively. In the case of the Bjerkreim-Sokndal lopolith, this variation can be correlated with location in the intrusion,  $\text{Mg}/$

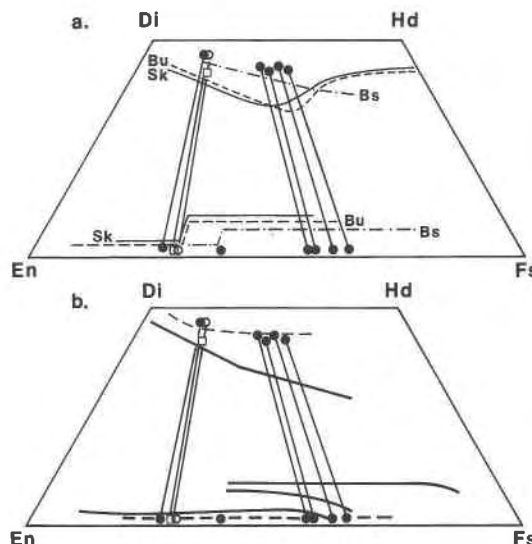


Fig. 5. Comparison of pyroxene compositions in symplectite-bearing samples with pyroxene crystallization trends in other rocks. (a) Data for other intrusive bodies; Sk = Skaergaard; Bu = Bushveld; BS = Bjerkreim-Sokndal (data from Duchesne, 1972b). (b) Generalized solidus (solid lines) and subsolidus (dashed lines) crystallization trends from Huebner (1980). Note the close correspondence in the compositions of the high-Ca pyroxenes and Duchesne's crystallization trend and Huebner's subsolidus crystallization trend. Symbols as in Fig. 4.

( $\text{Mg} + \Sigma\text{Fe}^{2+}$ ) decreasing with increasing stratigraphic height. This trend is generally similar to that in basic layered intrusions such as Skaergaard and to that in the Marcy anorthosite massif (Ashwal, 1982). The orthopyroxenes are characterized by relatively low  $\text{Al}_2\text{O}_3$  (<3.0 wt%),  $\text{TiO}_2$  (<0.3 wt%) and  $\text{Fe}_2\text{O}_3$  (<2.6 wt%). The latter was calculated by assuming that the sum of the cations is four on a six oxygen basis. The clinopyroxenes contain slightly higher amounts of these oxides:  $\text{Al}_2\text{O}_3$  up to 3.7 wt%,  $\text{TiO}_2$  up to 0.8 wt% and  $\text{Fe}_2\text{O}_3$  up to 3.8 wt%. The low-Ca contents of the orthopyroxenes are remarkable, especially considering the range in  $\text{Mg}/(\text{Mg} + \Sigma\text{Fe}^{2+})$ . This is illustrated in Figure 5 where the compositions of the pyroxenes are compared with the solidus trends for the orthopyroxenes in the Skaergaard and Bushveld intrusions and with the solidus trend established for the Bjerkreim-Sokndal low-Ca pyroxenes by Duchesne (1972b). The clinopyroxenes are more calcic than the solidus high-Ca pyroxenes in the Skaergaard and Bushveld intrusions but are compositionally similar to supposed solidus high-Ca pyroxenes from the Bjerkreim-Sokndal lopolith (Duchesne, 1972b). The compositions of the orthopyroxenes and clinopyroxenes agree well with Huebner's (1980) generalized subsolidus trends for quadrilateral pyroxenes (Fig. 5).

In contrast, the reintegrated bulk compositions of exsolved pyroxenes in the Bjerkreim-Sokndal lopolith, Hydra anorthosite, and Lyngdal hyperite are similar to those of solidus pyroxenes in other layered intrusions and to Huebner's (1980) solidus trends for quadrilateral py-

<sup>1</sup> Total Fe reported as  $\text{Fe}^{2+}$ .

TABLE 1—Continued

R25		H1		L1	
opx	cpx	opx	cpx	opx	cpx
49.4	50.9	52.2	50.9	53.0	52.4
0.10	0.18	n.d.	0.59	0.11	0.29
0.65	1.35	2.93	3.70	2.36	1.96
37.2	18.2	19.0	7.72	18.7	7.80
11.4	8.98	24.9	14.0	25.8	14.3
0.78	0.38	0.39	n.d.	0.24	0.14
0.88	20.2	0.56	22.1	0.42	21.9
n.d.	0.50	n.d.	0.64	n.d.	0.86
100.4	100.7	100.0	99.7	100.6	99.7
1.979	1.964	1.915	1.899	1.926	1.953
0.003	0.005	n.d.	0.017	0.003	0.008
0.031	0.061	0.127	0.163	0.101	0.086
1.246	0.587	0.583	0.241	0.568	0.243
0.681	0.516	1.362	0.778	1.398	0.795
0.026	0.012	0.012	n.d.	0.007	0.004
0.038	0.835	0.022	0.883	0.016	0.784
n.d.	0.037	n.d.	0.046	n.d.	0.062
4.003	4.019	4.021	4.027	4.020	4.026

TABLE 2. Temperature estimates (in °C) based upon the compositions of coexisting pyroxenes

	Wood and Banno (1973)	Wells (1977)	Lindsley* (1983)
GA248	832	793	700
R53	825	885	720
R186	843	914	810
R486	811	868	710
R25	803	858	700
H1	855	842	780
L1	844	821	751

\* Temperatures based on 5-kbar isothermal section.

roxenes. Nevertheless, from the petrographic descriptions, it should be clear that the analyses do not necessarily represent the compositions of pyroxenes crystallized from a melt (i.e., some external granule exsolution has occurred). It follows that the analyses of calcic pyroxenes reported by Duchesne (1972b) cannot represent solidus compositions.

The temperatures at which the pyroxenes equilibrated can be calculated from the relationships given by Wood and Banno (1973) and Wells (1977), and estimated by comparison of the high-Ca pyroxene compositions and the solvus isotherms presented by Lindsley (1983). They are listed in Table 2. Despite the uncertainties involved in the application of these methods [see, for example, Bohlen and Essene's (1979) discussion of the thermometers of Wood and Banno and of Wells], it seems safe to conclude that most of the pyroxenes, including those in the symplectites, equilibrated at temperatures <900 °C, mostly in the range 700–800 °C. It is noteworthy that there is no relationship between equilibration tempera-

ture and Mg/(Mg +  $\Sigma\text{Fe}^{2+}$ ) of the pyroxenes, i.e., with stratigraphic height in the intrusion. These relatively low temperatures may be compared with those obtained for the bulk analyses of the exsolved pyroxenes, which are  $\geq 1000$  °C. For those pyroxenes, equilibration temperatures correlate with Mg/(Mg +  $\Sigma\text{Fe}^{2+}$ ).

Most of the samples contain two types of Fe-Ti oxide: a member of the hematite-ilmenite series and a member of the ulvöspinel-magnetite series. In one sample from the lopolith (R53), titaniferous magnetite is the dominant Fe-Ti oxide, whereas in another sample (R195) as well as in the Hydra anorthosite, ilmenite is the dominant Fe-Ti oxide. Both ilmenite and titaniferous magnetite occur in the symplectites in all samples. Representative analyses are presented in Tables 3 and 4.

The complex exsolution phenomena described previously make analysis by microprobe techniques exceedingly difficult. This is especially true in the case of titaniferous magnetites, which show considerable compositional heterogeneity per sample. In R53, for example, TiO<sub>2</sub> contents vary from 0.20 to 7.5 wt%. This variation undoubtedly reflects extensive subsolidus re-equilibration, which also accounts for the low Al<sub>2</sub>O<sub>3</sub> contents of many analyses. Rarely, apparently homogeneous, small grains with high TiO<sub>2</sub> (up to 17%) and Al<sub>2</sub>O<sub>3</sub> (up to 2%) are found as discrete inclusions in subhedral pyroxene crystals (e.g., in R486). Attempts to reintegrate the original bulk compositions of exsolved grains using broad-beam techniques were successful only in the case of titaniferous magnetite in a symplectite in sample R25, the composition of which is similar to that of the discrete inclusions in pyroxene in R486. Comparison of the reintegrated titaniferous magnetite composition and the compositions of magnetite and

TABLE 3. Representative analyses of titaniferous magnetites

	GA248		R195	R53		R186		R486		R25	H1	L1
	1	2		1	2	1	2	1	2			
SiO <sub>2</sub>	0.50	0.05	n.d.	0.35	0.35	n.d.	n.d.	n.d.	0.39	n.d.	0.39	0.76
TiO <sub>2</sub>	1.30	0.35	0.40	7.40	0.66	3.10	0.25	17.0	0.28	0.95	1.08	0.27
Al <sub>2</sub> O <sub>3</sub>	0.80	0.65	0.65	3.93	n.d.	0.90	0.10	1.23	0.33	0.70	0.52	0.26
Cr <sub>2</sub> O <sub>3</sub>	0.80	0.90	1.45	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	1.02	n.d.
Fe <sub>2</sub> O <sub>3</sub> *	61.5	65.2	64.4	48.9	67.5	61.6	67.8	33.6	67.3	66.3	64.5	65.6
FeO*	31.8	31.2	31.4	38.9	32.4	33.9	30.9	45.8	31.9	31.9	32.8	31.7
MgO	0.40	0.15	0.10	n.d.	n.d.	n.d.	0.05	0.26	n.d.	n.d.	n.d.	0.16
MnO	0.05	0.05	n.d.	n.d.	n.d.	0.10	n.d.	0.30	n.d.	0.10	n.d.	n.d.
V <sub>2</sub> O <sub>3</sub>	0.80	0.85	0.70	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.
Total	98.0	99.4	99.1	99.5	100.9	99.6	99.1	98.12	100.2	100.0	100.3	98.8
Number of cations on the basis of four oxygens												
Si	0.019	0.002	n.d.	0.013	0.013	n.d.	n.d.	n.d.	0.013	n.d.	0.015	0.029
Ti	0.038	0.010	0.012	0.209	0.019	0.089	0.007	0.489	0.008	0.027	0.023	0.008
Al	0.037	0.029	0.030	0.174	n.d.	0.041	0.005	0.055	0.015	0.032	0.031	0.012
Cr	0.025	0.027	0.044	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	0.031	n.d.
Fe <sup>3+</sup> *	1.799	1.893	1.881	1.382	1.936	1.780	1.981	0.987	1.939	1.914	1.854	1.913
Fe <sup>2+</sup> *	1.033	1.002	1.006	1.222	1.032	1.086	1.004	1.464	1.023	1.024	1.046	1.028
Mg	0.023	0.002	0.006	n.d.	n.d.	n.d.	0.003	0.015	n.d.	n.d.	n.d.	0.009
Mn	0.002	0.002	n.d.	n.d.	n.d.	0.003	n.d.	0.010	n.d.	0.003	n.d.	n.d.
V	0.025	0.026	0.022	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.
Total	3.000	3.000	3.000	3.000	3.000	3.000	3.000	3.000	3.000	3.000	3.000	3.000
Usp**	0.041	0.011	0.013	0.254	0.019	0.093	0.007	0.513	0.008	0.028	0.033	0.008

Note: n.d. = not detected.

\* FeO and Fe<sub>2</sub>O<sub>3</sub> contents were calculated assuming  $\Sigma\text{Cat} = 3.000$  on a four-oxygen basis.

\*\* Ulvöspinel contents were calculated following Stormer (1983) and are expressed as mole fractions.

TABLE 4. Representative analyses of ilmenites

	GA248	R195	R186	R486		R25	H1	L1
				1	2			
SiO <sub>2</sub>	n.d.	n.d.	n.d.	0.21	0.22	n.d.	0.46	0.44
TiO <sub>2</sub>	50.9	46.1	52.4	49.7	51.8	53.3	49.4	49.9
Al <sub>2</sub> O <sub>3</sub>	n.d.	0.10	n.d.	5.65	n.d.	n.d.	n.d.	n.d.
Cr <sub>2</sub> O <sub>3</sub>	n.d.	0.20	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.
Fe <sub>2</sub> O <sub>3</sub> *	6.31	13.0	0.32	0.24	0.56	0.08	5.92	6.10
FeO*	39.7	40.6	46.1	44.4	46.3	47.3	41.5	41.0
MgO	3.20	0.15	0.25	n.d.	n.d.	n.d.	1.62	2.14
MnO	0.35	0.40	0.65	0.53	0.60	0.65	0.61	0.54
V <sub>2</sub> O <sub>5</sub>	n.d.	0.25	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.
Total	100.5	100.8	99.7	100.8	99.5	101.3	99.5	100.1
Number of cations on the basis of three oxygens								
Si	n.d.	n.d.	n.d.	0.005	0.006	n.d.	0.012	0.011
Ti	0.942	0.873	0.996	0.991	0.988	0.999	0.932	0.932
Al	n.d.	0.003	n.d.	0.162	n.d.	n.d.	n.d.	n.d.
Cr	n.d.	0.004	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.
Fe <sup>3+</sup>	0.117	0.243	0.008	0.005	0.012	0.002	0.113	0.113
Fe <sup>2+</sup>	0.817	0.859	0.973	0.906	0.981	0.985	0.870	0.853
Mg	0.117	0.006	0.009	n.d.	n.d.	n.d.	0.061	0.079
Mn	0.001	0.009	0.014	0.011	0.013	0.014	0.013	0.011
V	n.d.	0.005	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.
Total	2.000	2.000	2.000	2.000	2.000	2.000	2.000	2.000
R <sub>2</sub> O <sub>3</sub> **	0.058	0.121	0.004	0.002	0.006	0.001	0.056	0.057

Note: n.d. = not detected.

\* FeO and Fe<sub>2</sub>O<sub>3</sub> contents were calculated assuming ΣCat = 2.000 on a three-oxygen basis.

\*\* R<sub>2</sub>O<sub>3</sub> contents were calculated following Stormer (1983) and are expressed as mole fractions.

ilmenite in the same sample demonstrates that spinel (not analyzed) is a product of the exsolution process.

The ilmenites are much more homogeneous in composition, and there is little difference between analyses performed with a broad beam and those performed with a focused beam. The ilmenites in most samples contain low R<sub>2</sub>O<sub>3</sub> (0–3%) and low MgO (<2.0%), suggesting that they too have experienced extensive subsolidus re-equilibration, probably to the extent that discrete grains have formed from an initially homogeneous phase (external granule exsolution; Buddington and Lindsley, 1964). A similar conclusion was reached by Ashwal (1982) on the basis of his study of co-existing Fe-Ti oxides in the Marcy anorthosite massif and by Duchesne (1972a), who used mineral separates, for the oxides in the Bjerkreim-Sokndal lopolith. Ilmenites that contain relatively high R<sub>2</sub>O<sub>3</sub> (6–13%) occur as inclusions in subhedral pyroxenes in two samples (GA248 and R486) and as discrete grains in R195. MgO is relatively high in ilmenites in GA248 (~3%), suggesting that these approach solidus compositions, but is low in those in R486 and R195 (<2%), indicating some re-equilibration.

Temperatures and oxygen fugacities estimated from the experimental data of Buddington and Lindsley (1964) and Spencer and Lindsley (1981) are <630 °C and <10<sup>-18</sup> bar for most samples, confirming the conclusions reached above. The Fe-Ti oxides that occur as discrete inclusions in subhedral pyroxenes in R486 yield higher temperatures (~890 °C) and *f*<sub>O<sub>2</sub></sub> (~10<sup>-12.5</sup> bar), which may be compared with Duchesne's (1972a) estimates of 900–975 °C and 10<sup>-11</sup> atm for samples from a similar stratigraphic level. Temperatures estimated from Bishop's (1980) ilmenite-orthopyroxene geothermometer are likewise low

for most samples [≤680 °C, assuming equilibration at 5 kbar, which is consistent with estimates for the crystallization of the quartz monzonitic phase of the lopolith (Rietmeijer, 1979) and with estimates of the conditions of metamorphism in the aureole surrounding the lopolith (Hermans et al., 1976; Jansen and Majjer, 1980)]. Higher temperatures are calculated for the more magnesian ilmenite-pyroxene assemblages in GA248 (765 °C).

Spinel were analyzed in only two samples, one from the Lyngdal hyperite and one from the Hydra anorthosite (Table 5). The spinels are hercynites with negligible Cr<sub>2</sub>O<sub>3</sub> (<0.8%) and low Fe<sub>2</sub>O<sub>3</sub> (<1%, calculated assuming perfect analysis and stoichiometry), and there is no difference between the spinel that occurs as lamellae and the spinel that occurs as discrete inclusions in Fe-Ti oxides. Comparison of the co-existing magnetite and spinel compositions with the solvus data of Turnock and Eugster (1962) suggests temperatures of <600 °C, providing additional evidence for re-equilibration of the oxide phases in the intrusive bodies. Kars et al. (1980) have presented evidence that the spinel-magnetite pairs in metamorphic rocks from the contact aureole of the Bjerkreim-Sokndal lopolith re-equilibrated to relatively low temperatures (<800 °C).

All of the analyzed olivines (Table 5) are relatively magnesian (Fo<sub>70</sub> to Fo<sub>64</sub>) and within any one sample show a restricted range of composition (≤2 mol% Fo). The distribution of Mg and Fe between olivine and orthopyroxene in each sample conforms with the relationship derived by Medaris (1969) from his experimental study (see also Van Lamoen, 1979; Zeck et al., 1982).

Finally, the average bulk compositions of symplectites in one sample from the Bjerkreim-Sokndal lopolith and



TABLE 5. Analyses of spinels and olivines and broad-beam analyses of symplectites

	Spinel		Olivines			Symplectites				
	H1	L1	GA248	H1	L1	L1		R25		
SiO <sub>2</sub>	n.d.	n.d.	37.5	36.0	37.1	39.3		34.6		
TiO <sub>2</sub>	n.d.	n.d.	n.d.	n.d.	n.d.	1.12		2.19		
Al <sub>2</sub> O <sub>3</sub>	64.9	62.7	n.d.	n.d.	n.d.	2.57		1.36		
FeO*	21.5	25.2	26.8	30.9	32.3	35.9		52.5		
MgO	13.5	11.0	34.8	32.0	31.8	20.9		8.36		
MnO	0.31	0.42	0.66	0.21	0.29	0.18		0.54		
CaO	n.d.	n.d.	0.17	n.d.	0.10	0.29		0.55		
Total	100.2	99.3	99.9	99.1	101.6	100.3		100.0		
	Number of cations on the basis of four or six oxygens									
O	4	4	4	4	4	4	6	4	6	
Si	n.d.	n.d.	1.000	0.988	0.996	1.073	1.610	1.046	1.568	
Ti	n.d.	n.d.	n.d.	n.d.	n.d.	0.023	0.035	0.050	0.075	
Al	1.998	1.988	n.d.	n.d.	n.d.	0.083	0.124	0.049	0.073	
Fe <sup>2+</sup>	0.470	0.567	0.598	0.709	0.725	0.820	1.230	1.328	1.992	
Mg	0.526	0.441	1.383	1.309	1.273	0.851	1.276	0.377	0.566	
Mn	0.007	0.010	0.015	0.005	0.007	0.004	0.006	0.014	0.021	
Ca	n.d.	n.d.	0.005	n.d.	0.003	0.008	0.013	0.018	0.027	
Sum	3.001	3.006	3.000	3.012	3.004	2.862	4.294	2.880	4.321	
Mg/(Mg + ΣFe <sup>2+</sup> )	0.528	0.438	0.698	0.649	0.637	0.509	0.509	0.221	0.221	

Note: n.d. = not detected.

\* Total Fe as FeO or Fe<sup>2+</sup>.

in the Lyngdal hyperite are reported in Table 5. The symplectites in these samples are relatively fine grained and are, therefore, readily amenable to analysis using defocused-beam techniques. For each sample, at least nine spots were measured per symplectite, and at least three symplectites were analyzed. The analyses for all samples show a relatively small range in composition, and hence averages are reported in Table 5. Attention is directed to the low but significant Al<sub>2</sub>O<sub>3</sub>, TiO<sub>2</sub>, and CaO contents of the symplectites.

## DISCUSSION

### Origin of the symplectites: Magmatic or subsolidus?

The Fe-Ti oxides in the samples examined in the present study and also in many other intrusive bodies (e.g., McSween, 1980; Garrison and Taylor, 1981) have clearly experienced complex subsolidus re-equilibration, probably involving oxidation-reduction reactions (cf. Duchesne, 1972a; Ashwal, 1982) and external granule exsolution (Buddington and Lindsley, 1964). It is thus not possible to reintegrate the original (solidus) compositions of the Fe-Ti oxides with confidence, and measured compositions are of little help in constraining models for the origin of symplectites. The data presented in the preceding section demonstrate that the oxides in the symplectites in the Bjerkreim-Sokndal lopolith, Hydra anorthosite, and Lyngdal hyperite have compositions significantly different than those occurring as inclusions in pyroxenes in the same intrusions. Equilibration temperatures for the latter are substantially lower than those indicated by broad-beam analyses of the pyroxenes, suggesting some re-equilibration of even these oxides. Moreover, the relatively low Mg contents of the ilmenites that occur as inclusions in pyroxenes in sample R486 confirm the fact

that re-equilibration has occurred. Duchesne (1972a) proposed that re-equilibration of the Fe-Ti oxides was enhanced by the presence of a volatile-rich fluid phase trapped after solidification (see also Morse, 1980, p. 328), evidence for which is provided by the occurrence of autometamorphic biotite and amphibole. However, the possible role of convecting hydrothermal fluids (cf. Taylor and Forester, 1979; Norton and Taylor, 1979) cannot be discounted and must be evaluated through detailed stable-isotope studies. Inclusions of oxides in orthomagmatic pyroxenes were presumably relatively isolated from these fluids.

The exsolution textures of the feldspars and pyroxenes indicate that the silicates have also re-equilibrated during subsolidus cooling. External granule exsolution is inferred for most of the pyroxenes (and is also inferred for some of the plagioclase feldspars). The orthopyroxenes in the symplectites have low Ca contents and do not show exsolution. There are two possible explanations:

1. The symplectites crystallized from a melt, but at lower temperatures than the subhedral pyroxenes in the same sample or than pyroxenes in most other layered intrusions and anorthosites (Bohlen and Essene, 1978; Ashwal, 1982). Comparison of the temperatures estimated for the formation of the symplectitic pyroxenes with those estimated from bulk analyses of larger, exsolved pyroxenes leaves no doubt that the latter crystallized at substantially higher temperatures than the former (Barton, in prep.). Arculus and Wills (1980) and Wilson et al. (1981) ascribed the presence of magmatic low-Ca pyroxene (hypersthene) in plutonic blocks from the Lesser Antilles and in the Fongen-Hyllingen intrusive complex, Norway, to relatively low-temperature crystallization and suggested that solidus temperatures and pyroxene crystallization temperatures were depressed by the presence

of water. Equilibration temperatures (800–900 °C) estimated for the pyroxenes in the symplectite-bearing samples using the thermometers of Wood and Banno (1973) and of Wells (1977) are consistent with crystallization at the solidus of hydrous anorthosite at 5 kbar (800 °C, Luth and Simmons, 1968). However, the absence of abundant pegmatitic segregations in the Bjerkreim-Sokndal lopolith and the absence of abundant orthomagmatic amphibole indicates that H<sub>2</sub>O-saturated residual melts were not, in general, generated during crystallization of the anorthositic-leuconorit phase, and we thus consider it extremely unlikely that the symplectites formed by coprecipitation from a melt.

2. The symplectites formed under subsolidus conditions. This explanation is entirely consistent with the compositional similarity between the symplectitic orthopyroxene, that which occurs as rims around olivine and clinopyroxene and that which occurs as discrete granules (which are clearly of subsolidus or metamorphic origin). Furthermore, subsolidus crystallization explains the absence of any relationship between estimated equilibration temperatures and Mg/(Mg + ΣFe<sup>2+</sup>) of the pyroxenes and stratigraphic height in the intrusion. As noted earlier (see Fig. 5), the compositions of the pyroxenes in the symplectites conform to established subsolidus trends in the conventional pyroxene quadrilateral, whereas the bulk compositions of the exsolved pyroxenes approach solidus trends in other intrusions. Temperatures of formation of the symplectites based upon the Lindsley (1983) pyroxene thermometer are 700–800 °C and are probably more reliable for granulite-facies conditions than temperatures obtained using the methods of Wood and Banno (1973) and of Wells (1977) (Stephenson, 1984). These temperatures are consistent with formation under subsolidus conditions. Finally, it is noteworthy that similar equilibration temperatures are estimated for co-existing pyroxenes in the symplectite-bearing samples, in other samples from the anorthositic-leuconorit phase of the Bjerkreim-Sokndal lopolith (Barton, in prep.), in samples from the Lyngdal hyperite (Knibbeler and Van Zutphen, oral comm., 1983) and in the noritic Hunnedalen dyke system (Drent, 1982). These temperatures are lower than those estimated for the equilibrium mineral assemblages in the metamorphic aureole immediately surrounding the Bjerkreim-Sokndal lopolith (Jansen and Majjer, 1980), which indicates regional re-equilibration of pyroxenes in many of the intrusive bodies in Rogaland-Vest Agder during postintrusive cooling (i.e., during the M3 metamorphic phase; Jansen and Majjer, 1980; Kars et al., 1980).

The evidence cited above strongly suggests that the orthopyroxene-Fe-Ti oxide symplectites described in this study formed during subsolidus cooling. The low Ca contents of orthopyroxenes in symplectites in gabbroic intrusions in other parts of the world suggest that they are probably also of subsolidus origin, even if formation during late-magmatic crystallization (Ambler and Ashley, 1977; Garrison and Taylor, 1981) has previously been proposed.

### Formation of the symplectites

A number of models can be envisaged for the formation of orthopyroxene-Fe-Ti oxide symplectites: exsolution from a pre-existing homogeneous phase, eutectoidal breakdown of a pre-existing homogeneous phase, reaction between primary magmatic minerals during cooling, oxidation or replacement of olivine, and independent nucleation and growth of low-Ca pyroxene and Fe-Ti oxide.

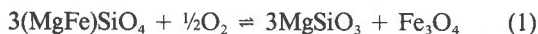
**Formation by exsolution.** The irregularity of the symplectites is difficult to explain in terms of an exsolution model, especially when it is recalled that even the apparently lamellar symplectites are irregular in three dimensions and that the Fe-Ti oxide “lamellae” are not continuous along their length. However, exsolution from a supersaturated solid solution by discontinuous or cellular precipitation (Yund and McCallister, 1970; Chadwick, 1972) can produce textures similar to those described here. This process has been invoked to explain the origin of the pyroxene-ilmenite symplectites that occur as discrete nodules in kimberlites (Dawson and Reid, 1970) as well as the pyroxene-plagioclase symplectites described by Boland and van Roermund (1983) and the symplectitic augites described by Boland and Otten (1985). In the case of the symplectites that are the subject of the present study, there is no logical precursor solid-solution phase. The molar proportions (converted from volume proportions) of orthopyroxene:Fe-Ti oxide (avg. 66:34) together with the occurrence of abundant ilmenite in some symplectites (e.g., in sample R195) rules out formation of symplectites from an original, nonstoichiometric pyroxene: orthopyroxene crystallized under a range of P-T conditions contains  $\leq 1.4$  wt% TiO<sub>2</sub> (Deer et al., 1978), and this is also true of hypersthene megacrysts in anorthosites (Emslie, 1978; Maquil, 1978). Direct evidence is provided by the bulk analyses of the symplectites in the Lyngdal hyperite and in the Bjerkreim-Sokndal lopolith that deviate strongly from pyroxene stoichiometry (Table 5). We thus conclude that the symplectites do not form by exsolution, although we believe that the fine ( $\sim 3$  μm), oriented (parallel to 010 and 001) platelets of ilmenite found in some nonsymplectitic pyroxenes in the samples that we have studied formed by exsolution, possibly from nonstoichiometric hosts (cf. Garrison and Taylor, 1981). As noted by Barink (1982), silicates crystallized at high temperatures may be nonstoichiometric, and the oxygen released upon cooling may combine with Fe<sup>2+</sup> and Ti to produce magnetite, hematite, or ilmenite inclusions in the host phase. This process can only yield relatively minor amounts of Fe-Ti oxide (see Barink, 1982).

**Formation by eutectoidal breakdown.** Eutectoidal reactions are known from studies in metallurgical systems (Chadwick, 1972) to produce textures similar to those of the symplectites. Ringwood and Lovering (1970) have proposed that the pyroxene-ilmenite intergrowths that occur as discrete nodules in kimberlite diatremes result from eutectoidal breakdown of garnet. This possibility can be discounted for the orthopyroxene-Fe-Ti oxide

symplectites because there is no evidence that garnet occurred as a primary magmatic, xenocrystal, or metamorphic phase in the rocks under consideration. Additional evidence against this hypothesis includes the relatively low Al content of the symplectites and the improbability that garnet transforms to pyroxene and Fe-Ti oxide at pressures that are geologically realistic for layered intrusions (contrast Ringwood and Lovering, 1970). As with the exsolution hypothesis, there is no other likely precursor phase that could form the symplectites by eutectoidal decomposition.

**Reaction between primary magmatic minerals during cooling.** The development of coronas of orthopyroxene and amphibole and amphibole-spinel symplectite is thought to result from subsolidus reaction between olivine and plagioclase (Van Lamoen, 1979; Mongkoltip and Ashworth, 1983). Such reactions clearly cannot be responsible for the formation of orthopyroxene-Fe-Ti oxide symplectites described here. There is absolutely no evidence that these symplectites formed *directly* by reaction between any pair of orthomagmatic minerals although, as shown in the following sections, formation of the symplectites must be related to reactions occurring elsewhere in the same rock.

**Oxidation or replacement of olivine.** Following the suggestion of Muir et al. (1957), many workers have proposed that orthopyroxene-Fe-Ti oxide symplectites form by the breakdown of olivine according to a reaction of the type



(see Haggerty and Baker, 1967; Goode, 1974). Petrographic data for the Hydra anorthosites and Lyngdal hyperites can be interpreted in terms of this reaction, which should not be confused with the eutectoidal reaction described above (note that the latter must, by definition, be invariant). However, the reaction mechanism remains obscure. Muir et al. (1957) and Van Lamoen (1979) favored discontinuous replacement reactions whereas Johnston and Stout (1984) proposed a continuous oxidation reaction for the formation of orthopyroxene-Fe-Mg ferrite symplectites in lherzolite xenoliths from Kauai, Hawaii. Johnston and Stout (1984) have discussed the different reaction mechanisms in detail, and we note here only that major points of difference are (1) which components are added to or subtracted from the olivine during replacement and (2) does olivine persist, though changing composition, throughout the reaction? In the following sections we discuss these and other points relevant to the formation of the symplectites via oxidation and/or replacement of olivine.

1. Olivine will become unstable during cooling if  $f_{\text{O}_2}$  does not decrease parallel to a  $T$ - $f_{\text{O}_2}$  buffer curve. This follows from the work of Nitsan (1974), which indicates that subsolidus decomposition of olivine according to Reaction 1 (or a similar reaction) does not require an absolute increase in  $f_{\text{O}_2}$ . The results of the present study, together with results reported by Duchesne (1972a), in-

dicate that during cooling from 975 to 840 °C,  $f_{\text{O}_2}$  values in the Bjerkreim-Sokndal lopolith paralleled those of the QFM buffer. During subsequent cooling, olivine of fixed composition will break down to orthopyroxene and Fe-Ti oxide if the rate at which  $f_{\text{O}_2}$  decreases with respect to temperature is less than that defined by the QFM buffer. We suggest that influx of water along fractures resulting from thermal contraction may alter the  $T$ - $f_{\text{O}_2}$  cooling path sufficiently to induce oxidation of orthomagmatic olivine.

2. Reaction 1 is divariant and thus cannot be uniquely balanced, even if minor components are ignored and the Fe-Ti oxide is assumed to be pure magnetite, unless the bulk composition is known (Hensen, 1980). The assumption that olivine composition remains constant during symplectite formation (Van Lamoen, 1979) is not necessarily valid (cf. Grieve and Gittins, 1975), and indeed Johnston and Stout (1984) have demonstrated that continuous oxidation of olivine occurs under at least some circumstances. However, the symplectites in the Bjerkreim-Sokndal lopolith, Lyngdal hyperite, and Hydra anorthosite cannot have formed in this way, since the bulk compositions of the symplectites differ considerably from those of the host olivine or, in the case of the symplectites in the Bjerkreim-Sokndal lopolith, from olivine stoichiometry. Moreover, summing the compositions of the orthopyroxene and Fe-Ti oxide in symplectites from the Bjerkreim-Sokndal lopolith and measured proportions of these phases yields unrealistic olivine compositions. For Ga248, the calculated olivine composition is  $\text{Fo}_{41}$ , whereas the measured olivine composition is  $\text{Fo}_{70}$ , and comparable compositions for the monzonorites are  $\text{Fo}_{19}$  (calculated) and  $\text{Fo}_{30}$  (measured—see Duchesne, 1972b). The occurrence of spinel and ilmenite in the symplectites together with the measured  $\text{Al}_2\text{O}_3$  and  $\text{TiO}_2$  contents of the symplectites demonstrates that the intergrowths cannot form by isochemical replacement of olivine. We conclude that formation of the symplectites in the Bjerkreim-Sokndal lopolith, Lyngdal hyperite, and Hydra anorthosite involves diffusion of components other than oxygen. Moreover, we note that our data indicate that there is no a priori reason to assume that  $\text{Al}_2\text{O}_3$  or  $\text{TiO}_2$  behave as immobile components (cf. Mongkoltip and Ashworth, 1983) during symplectite formation.

3. Replacement of olivine involves nucleation and growth of new phases. The occurrence of granular pyroxenes and low-Ti magnetites in the symplectite-bearing samples described here is evidence that these phases nucleated and grew during retrograde metamorphism. Nucleation most probably occurred at grain boundaries or imperfections (e.g., vacancy clusters and dislocations) within orthomagmatic minerals since this is energetically most favorable. Evidence from one of the samples from the Bjerkreim-Sokndal lopolith suggests that nucleation occurred on a crystal of apatite, though this interpretation is debatable since the apatite crystal could have been engulfed during growth (this can be inferred in the case of albite-clinopyroxene symplectites in the Lyngdal hyperite). Nevertheless, it is thermodynamically improbable

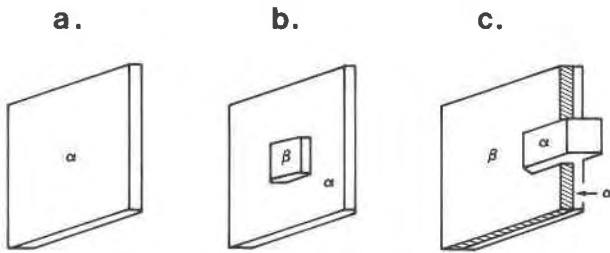


Fig. 6. Illustration of the bridging mechanism described in the text. (a) Idealized, isolated plate of  $\alpha$ -phase. (b) Nucleation of  $\beta$ -phase on the surface of existing plate of  $\alpha$ -phase. (c) Growth of a second lamella of  $\alpha$  is initiated after bridging at the phase edge. This mechanism allows the formation of intergrowths without requiring continuous nucleation of each phase, i.e., the formation of symplectites is a consequence of a specific growth mechanism. Bridging will result in discontinuities in lamellae along the direction of propagation.

that replacement of olivine is in most cases isochemical. The En content of orthopyroxene, for example, will be determined by the Fo content of the olivine according to a relationship such as that determined by Medaris (1969), as shown by the work of Van Lamoen (1979) and Zeck et al. (1982) and the work presented in this paper. Furthermore, the orthopyroxene will have Ca and Al contents commensurate with equilibrium with phases such as clinopyroxene and plagioclase occurring in the same rock (or forming at the same time). The same holds true for high-temperature metamorphic magnetite, the Ti and Al contents of which must be consistent with *at least* partial equilibrium with coexisting spinel and ilmenite. Numerous examples of nonisochemical replacement have been described from metamorphic rocks, e.g., sillimanite by grandierite (van Bergen, 1980) and osumilite by cordierite, alkali feldspar, quartz (Jansen, oral communication, 1983), and it is perhaps surprising that previous workers have sought to obtain balanced solutions for the formation of symplectites.

4. The formation of the symplectites is critically dependent upon the availability of Fe-Ti oxide; where grains of orthomagmatic Fe-Ti oxide are not available, olivine is partially replaced by orthopyroxene. During symplectite formation, nucleation is probably cooperative (Chadwick, 1972), nucleation of one phase inducing (via supersaturation) nucleation of the second phase. Either phase could nucleate first, depending upon a number of factors such as variations in the local chemical environment and the degree of lattice coherency between the original and the new phases.

5. Following nucleation, the two phases propagate via a sympathetic growth mechanism (Chadwick, 1972) governed by the requirements of minimal interfacial and strain energy. Chadwick (1972) and Chalmers (1964) have described a bridging mechanism to account for the sympathetic nucleation and multiplication of lamellae in a eutectic alloy (Fig. 6); such a mechanism is entirely consistent with our observations on orthopyroxene-Fe-Ti oxide symplectites. The form of the intergrowths suggests

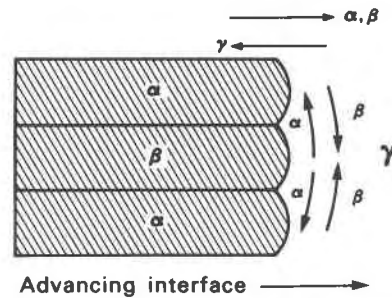


Fig. 7. Directions of diffusion required for growth of a symplectite. Components must diffuse both parallel and perpendicular to the direction of growth (from left to right in the above diagram). Diffusion probably occurs predominantly along grain boundaries.

nucleation and growth under supercooled conditions, perhaps induced by strong thermal and chemical gradients or, more probably, by the relatively high activation-energy barriers for solid-state nucleation. The precise form of the intergrowth is likely to be governed by a number of factors, including the degree of lattice coherency, surface- and strain-energy requirements, evolution of latent heat during crystallization, and diffusion rates. The last factor is important, since the growth of the symplectite requires diffusion both perpendicular and parallel to the advancing interface (Fig. 7). Mongkoltip and Ashworth (1983) have illustrated the possible diffusion processes involved in symplectite and corona formation. Since volume diffusion is relatively slow for solid-state reactions (Yund and McCallister, 1970; Freer, 1981; Boland, oral comm., 1983), grain-boundary diffusion is likely to have been dominant (note that the reacting interface must have been incoherent) and may have been enhanced by the presence of a thin film of intergranular fluid (note the occurrence of autometamorphic amphibole and biotite).

6. Our observations strongly suggest that not all symplectitic intergrowths of Fe-Ti oxide and orthopyroxene result from the replacement of olivine. It thus seems impossible to avoid the conclusion that such intergrowths can form in a recrystallizing rock at any point where nucleation and growth of Fe-Ti oxide and orthopyroxene occurs. This conclusion is consistent with Zeck et al.'s (1982) description of orthopyroxene-ilmenite symplectite that apparently replaces olivine and with the occurrence in the Bjerkreim-Sokndal lopolith of rims of symplectite around orthomagmatic clinopyroxene (from which orthopyroxene has exsolved) in the vicinity of orthomagmatic Fe-Ti oxide.

In summary, there is no reason to expect that the replacement of olivine by symplectites is isochemical (though it can be for components other than oxygen); furthermore, not all orthopyroxene-Fe-Ti oxide symplectites appear to form by replacement of olivine. We believe that the occurrence of symplectites reflects the overall re-equilibration of a magmatic rock to the *P-T* conditions prevailing during retrograde metamorphism and that their rarity results from the complex factors re-

quired for formation. It is not clear why the proportions of the phases in the symplectites are relatively constant, though the proportions and the size of the phases may be governed by diffusion rates and by growth rate. The interlamellar spacing of eutectoid alloys cannot be predicted simply from theoretical considerations, but a relationship of the type  $\lambda^{\nu}R \approx \text{constant}$  (where  $\lambda$  is interlamellar spacing,  $\nu$  is a constant and  $R$  is growth rate) is probably applicable to metallurgical systems (Chadwick, 1972). However, it is noteworthy that the symplectites in the Lyngdal hyperite are much finer grained than those in the Bjerkreim-Sokndal lopolith and Hidra anorthosite even though all of these intrusives have experienced a similar subsolidus cooling history. It is thus clear that before the formation of symplectites by solid-state processes can be fully understood, detailed studies involving X-ray, TEM, and, especially, experimental techniques must be undertaken.

### CONCLUSIONS

The symplectites in intrusive bodies from southwestern Norway formed during prolonged subsolidus cooling (equivalent to the M3 phase of retrograde metamorphism recognized extensively in the basement of Rogaland and Vest Agder). In some samples, the symplectites appear to replace olivine, but this replacement is not isochemical. The compositions of the component phases of the intergrowths reflect the overall re-equilibration of all phases in the rocks with falling temperature, and there seems to be no grounds to believe that any chemical components are inert or immobile during symplectite formation. In some samples, there is no evidence that the symplectites replace olivine, and it is concluded that such intergrowths can form during retrogression where the component phases are available. Certainly, the presence of discrete grains of orthomagmatic Fe-Ti oxide is critical for the formation of the symplectites. The mechanism of nucleation and growth can be described in a qualitative manner only. Nucleation is probably cooperative, nucleation of one phase inducing nucleation of the second phase, and growth occurs by a sympathetic mechanism governed by the requirements of minimal interfacial and strain energy. A bridging mechanism accounts for the multiplication of the lamellae once growth is initiated. Integrated mineral, chemical, TEM, experimental, and theoretical studies are required before symplectite formation can be fully understood. However, as stressed earlier, an understanding of their formation should provide much useful information about the cooling histories of plutonic rocks and hence about the thermal history of the crust.

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