

Högbomite in sapphirine-bearing rocks from the Bamble Sector, south Norway

D. VISSER, P. H. M. THIJSEN*

Department of Geochemistry, Institute of Earth Sciences, University of Utrecht, P.O. Box 80.021, 3508 TA Utrecht, The Netherlands

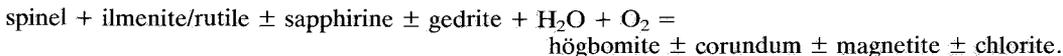
AND

J. C. SCHUMACHER**

Mineralogisches Institut, Olshausenstrasse 40, W-2300 Kiel, Germany

Abstract

Högbomite is reported from two upper-amphibolite and granulite-facies, sapphirine-bearing, Al-Mg-Fe-rich and silica-poor lens-shaped layers within the Bamble Sector, south Norway. Primary assemblages, indicating peak metamorphic conditions of 773–844 °C at 7 kbar (Mg-Fe exchange thermometry), are spinel-sapphirine-biotite-gedrite, spinel-corundum-sapphirine-cordierite and orthopyroxene-biotite-cordierite-plagioclase. Högbomite formed by hydrous alteration and oxidation of primary spinel and rutile and/or ilmenite according to the generalised reaction:



Suggested conditions of högbomite formation are 550–620 °C and 6–7 kbar. The högbomites contain 10.2–14.7 wt.% MgO, 0–0.3 wt.% ZnO, 58.9–62.1 wt.% Al₂O₃ and 15.6–17.6 wt.% Fe as FeO. The two högbomites may belong to different polytypes, as suggested by their differing TiO₂ (9.9–10.1 versus 5.7–5.8 wt.%) and calculated Fe³⁺- and H₂O-contents. The partitioning of Zn between spinel and högbomite is not uniform and is considered to depend upon prevailing f_{O_2} and $a_{\text{H}_2\text{O}}$.

KEYWORDS: högbomite, spinel, secondary formation, oxidation, sapphirine-bearing rock, microprobe analysis, Bamble Sector, south Norway.

Introduction

HÖGBOMITE, a complex Fe-Mg-(Co-Zn)-Al-Ti oxide, occurs as a minor constituent of sapphirine-bearing rocks at Snaresund and in the vicinity of Rangleåsen, Bamble Sector, south Norway (Fig. 1). The Snaresund sampling site is situated in the granulite facies zone of the Arendal-Tvedestrand area, whereas the Rangleåsen sampling site is located in the adjacent upper-amphibolite facies zone. Peak metamorphic conditions for this Kongsbergian (Gothian) (1600–1500 Ma) granulite facies event are estimated to

be 800 ± 60 °C and 7.3 ± 0.5 kbar (Lamb *et al.*, 1986). *P-T* conditions for the associated upper-amphibolite facies metamorphism are 740 ± 60 °C and 7 kbar (Visser and Senior, 1990). Uplift of the area occurred during the Sveconorwegian Orogeny (*ca.* 1200–1000 Ma), which manifested itself as a retrograde greenschist- to amphibolite-facies event with extensive fluid activity. *P-T* conditions for the onset and completion of this uplift and rehydration event are constrained at 625–700 °C and 6–7 kbar and 420–530 °C at 3–4 kbar respectively (Nijland *et al.*, in prep.; Visser and Senior, 1990). The sapphirine-bearing rocks of the Snaresund locality were first described by Touret and de la Roche (1971). Subsequently Lamb (1981) presented a detailed study of several sapphirine-bearing occurrences from the Bamble Sector including the Snaresund and Rangleåsen localities. Högbomite was not reported in either study.

* Present address: Forschergruppe Hochdruck Metamorphose, Institut für Mineralogie, Ruhr-Universität Bochum, Postfach 102148, W-4630 Bochum 1, Germany.

** Present address: Mineralogisch-Petrographisches Institut, Albertstrasse 236, W-7800 Freiburg, Germany.

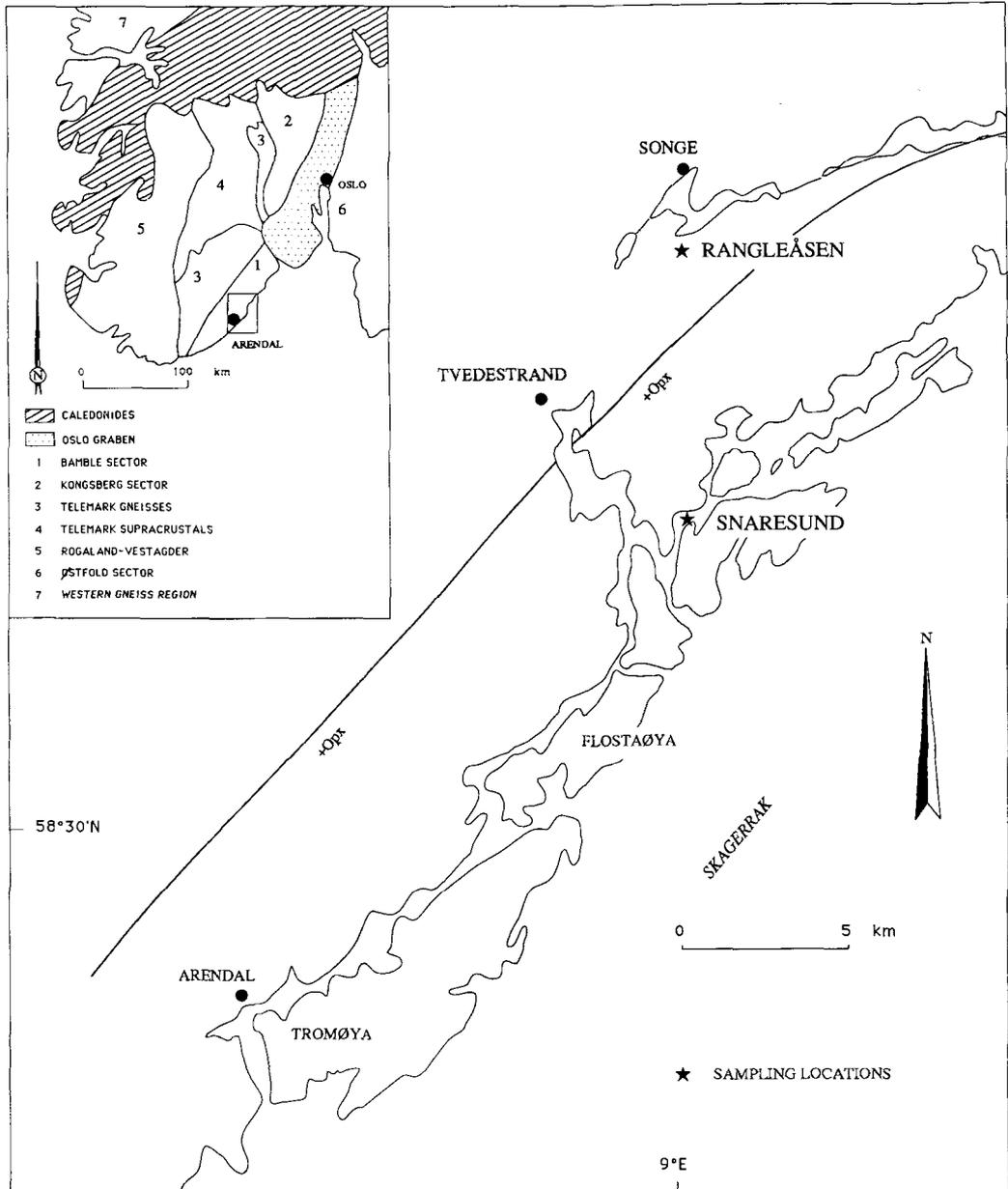


FIG. 1. Location map of the Arendal-Tvedestrand area, Bamble Sector, south Norway. Opx-isograd is after Field and Clough (1976). Geological division of southern Norway (inset) has been adapted from Verschure (1985).

The occurrence of högbomite in such high-grade aluminous and silica-poor rock types would not be surprising, since it has been reported from similar aluminous high-grade, regional metamorphic rocks from Australia (Woodford and Wilson, 1976), W. Greenland (Ackermund *et al.*, 1983),

India (Grew *et al.*, 1987) and Antarctica (Grew *et al.*, 1990).

Högbomite was first reported from Norway by Schetelig (1917) in iron-ores from the Fe-Ti-V Rödsand deposit, near Kristiansund. Kolderup (1928) described högbomite associated with

ilmenite-ores from the Sogn district and from the Bergen district associated with iron-ores in anorthositic rocks (Kolderup, 1936).

The present paper, which was largely motivated by the extensive review on högbomite occurrences by Petersen *et al.* (1989) and the numerous reports by Grew and co-workers (1987, 1989, 1990), describes the detailed petrography and mineral chemistry of the Bamble högbomites and associated minerals and discusses their conditions of formation. As far as the authors are aware, this is only the fourth report of högbomite from Norway, and the first in which compositional data are given.

Petrography

At Snaresund the only högbomite-bearing sample (MA 818) was collected from the core of a 0.4 m thick sapphirine-bearing lens-shaped outcrop within a layered sequence of acid to intermediate gneisses, quartzites and metabasites. The outer zone of the lens is dominantly phlogopite (60–80 vol. %) with minor amounts of orthopyroxene, gedrite, cordierite, sapphirine, spinel, corundum, rutile, and garnet. An intermediate

zone between the core and outer zone consists of orthopyroxene + gedrite + plagioclase (An% 50) with minor phlogopite, cordierite and rutile. The core consists of alternating layers of orthopyroxene poikiloblasts + phlogopite ± cordierite + plagioclase (An% 31–33) + quartz and cordierite + sapphirine + spinel + corundum. Minor rutile and trace amounts of ilmenite are present in the sample. Ilmenite shows variable amounts of hematite exsolution (Lamb, 1981). Embayed sapphirine aggregates and grains which are in optical continuity, separate anhedral spinel and/or corundum from a polygonal fabric of cordierite. Spinel is locally found in direct contact with corundum, but, generally, these minerals are separated by sapphirine. The green to black anhedral spinels show a moderate degree of magnetite exsolution. Some spinel grains are replaced by aggregates of magnetite, corundum and rarely chlorite. Högbomite is restricted to the alteration aggregates after spinel containing rutile and/or ilmenite inclusions or intergrowths. Högbomite forms euhedral to subhedral grains up to 100 µm partially rimming rutile and/or ilmenite (Fig. 2). Högbomite also occurs adjacent to spinel and sapphirine.

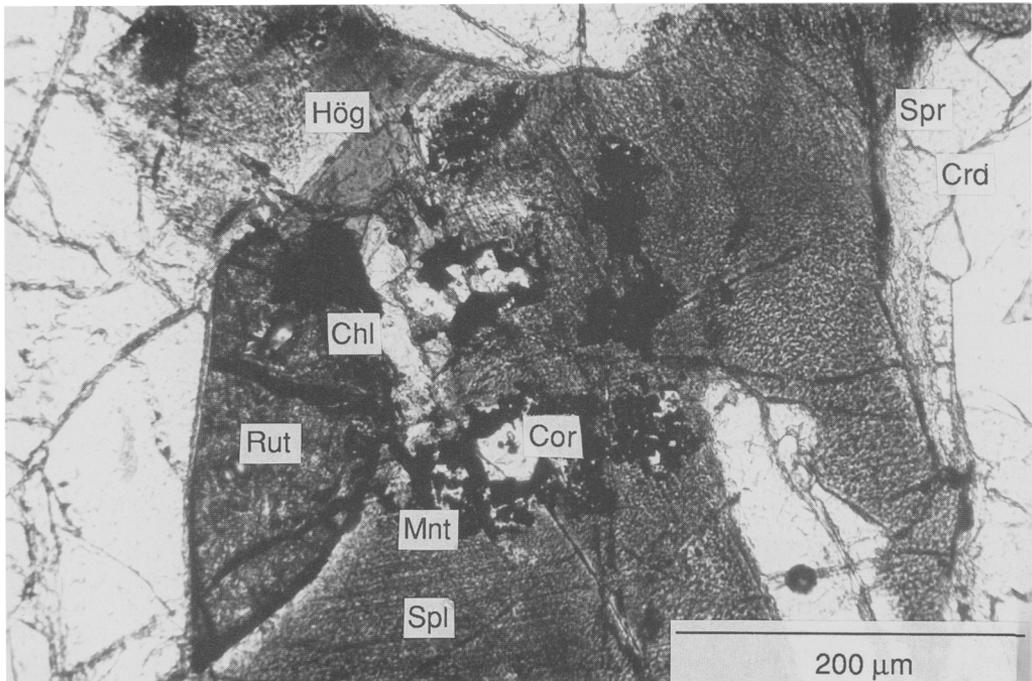


FIG. 2. Photomicrograph of sample MA 818 showing the replacement of spinel (Spl) adjacent to rutile (Rut) by högbomite (Hög), corundum (Cor), magnetite (Mnt) and chlorite (Chl). Spinel grains are dusted due to exsolved magnetite.

Sample PT 160 was collected west of Rangleåsen from a 0.5–2 m thick horizon in a layered sequence of cordierite–orthoamphibole gneisses and garnet–amphibolites. The layer consists of a decussate texture of euhedral orthoamphibole grains (75–90 vol. %) intergrown or interfingering with green to black spinel (1–15 vol. %). Sapphirine (1–3 vol. %) commonly forms a discontinuous rim between orthoamphibole and spinel but may also occur as solitary grains intergrown with orthoamphibole. Pale brown phlogopite, orthopyroxene, rutile, and ilmenite are present in minor amounts. Euhedral monazite is present in trace amounts in sample PT 160 but occurs as an important accessory mineral (up to 5 vol. %) in many other samples from this locality. Sample PT 160 is cut by linear retrograde zones or fractures along which orthoamphibole, sapphirine and biotite are altered to chlorite. Spinel grains cut by these zones show more extensive magnetite exsolution and, locally, are pseudomorphed by magnetite, corundum and chlorite. Spinel grains outside these zones exhibit only minor magnetite \pm corundum exsolution. Högbomite occurs exclusively within the alteration zones. Small euhedral to subhedral grains, often in the absence of another Ti-bearing phase, are located on or in the margins of spinel and pseudomorphs of spinel. Larger euhedral grains occur in fractures within the decomposed spinels and along spinel–rutile and spinel–ilmenite contacts. Additionally, högbomite can be found in grain contact with orthoamphibole and sapphirine, but has not been found with biotite. Breakdown of högbomite was not observed in either sample MA 818 or sample PT 160.

Mineral chemistry

Minerals in sample MA 818 were analysed with a JEOL JXA-8600 superprobe at 15 kV and 10 nA at the University of Utrecht; mineral and synthetic compounds were used as standards. Raw count data were corrected with a Tracor Northern PROZA correction program. Analyses of minerals in sample PT 160 were performed using the Cameca electron microprobe (Camebax Micro), at 15 kV, 10 nA with PAP (Pouchou and Pichoir, 1984) correction at the University of Kiel, Germany. Representative analyses of högbomite and associated minerals are listed in Tables 1 and 2.

The högbomites from both samples are unzoned and display no major compositional differences within the sample. Compared to those from sample PT 160 the högbomites from sample MA 818 are distinctly richer in TiO₂, lower in Al₂O₃ and MgO and slightly higher in Fe as FeO

(see Tables 1 and 2 for specific ranges). ZnO (below detection limit in MA 818), MnO and Cr₂O₃ values are low. SiO₂, NiO, CaO, K₂O, Na₂O, Cl and F are below detection limit. Mg/(Mg + Fe_{Tot}) range 0.51 (MA 818) and 0.60–0.63 (PT 160).

Oxide totals of the högbomite analyses with Fe_{tot} = FeO are well below 100%, 96.71–97.91% for MA 818 and 98.36–98.8% for PT 160. The deficit could be made up by Fe³⁺, H₂O and/or unanalysed light elements such as BeO and Li₂O. Calculation of Fe³⁺ on the basis of the general anhydrous formula for högbomite ($R^{2+}_{2-2x}Ti_xR^{3+}_4O_8$, with $R^{2+} = Mg, Fe^{2+}, Mn, Zn, Co, \dots Ni$ and $R^{3+} = Al, Fe^{3+}, Cr$ and V) as proposed by Zakrzewski (1977) yields appreciable Fe³⁺, $Fe^{3+}/(Fe^{3+} + Fe^{2+}) = 0.46–0.47$ and 0.35–0.38 respectively for MA 818 and PT 160, with recalculated oxide totals close to 100% for PT 160 but still far below 100% for sample MA 818 (see Tables 1 and 2). The correlation of the total cation content with Ti used by Zakrzewski (1977) to calculate Fe³⁺ is probably invalid (cf. Grew *et al.*, 1987; Petersen *et al.*, 1989) or perhaps only applicable to anhydrous polytypes. Schmetzer and Berger (1990) observed the general replacement scheme $(2a-b)Mg^{2+} + (2b)Al^{3+} \rightarrow (a + b)Ti^{4+}$ in iron-free högbomite-24R from Tanzania (högbomite can be described as various stacking sequences or polytypes of spinel-like ($R^{2+}Al_2O_4$) and nolanite-like (TiAl₄O₈) layers; Peacor, 1967), and on the basis of structure refinements, Gatehouse and Grey (1982) found one Ti per 22 cations and 32 anions of an 8H polytype högbomite instead of two Ti predicted by Zakrzewski's formula.

The refinements of Gatehouse and Grey (1982) further suggest the presence of 2 hydroxyl groups, on the basis of undersaturated oxygen positions, giving an anion composition of 30O + 2OH instead of 32O. However, the formula of Gatehouse and Grey (1982) gives more than 22 cations for complete analyses of 36R and 10H polytypes (see Rammlmair *et al.*, 1988) and appears, therefore, only applicable to 8H polytypes. Evidence for the presence of water in the högbomite structure is still very inconclusive. While Nel (1949) reported H₂O-values of 1.10 and 1.29 wt.% H₂O⁺ in iron-rich högbomites, infrared spectra of Fe–Ti-rich högbomites obtained by Zakrzewski (1977) and a wet-chemical analysis of a zincian högbomite (Wilson, 1977) did not yield any meaningful H₂O or hydroxyl content. Nonetheless, low oxide totals have been reported from an increasing number of (frequently 8H) högbomite occurrences (Čech *et al.*, 1976; Coolen, 1981; Mancktelow, 1981; Ackermann

Table 1. Chemical compositions of minerals in sample MA 818, Snaresund, Bamble Sector, south Norway.

	Högbomite [#]		range	Spl	Ilm	Mag	Crn	Spr	Crd
SiO ₂	0.00	0.00	0.00	0.00	0.38	0.00	0.00	12.70	48.91
TiO ₂	9.95	9.95	9.95-10.14	0.02	51.51	0.15	0.03	0.04	0.03
Al ₂ O ₃	60.06	60.06	58.90-60.06	63.55	0.00	0.36	99.82	62.29	33.26
Cr ₂ O ₃	0.00	0.00	0.00-0.12	0.24	0.00	0.00	0.11	0.00	0.00
Fe ₂ O ₃	3.80	8.51	-	2.08	3.55	67.37	0.79	2.17	-
FeO	14.15	9.91	17.37-17.57	19.12	42.30	30.67	-	6.96	3.16
ZnO	0.00	0.00	0.00	0.83	0.00	0.00	0.00	0.00	0.00
MnO	0.12	0.12	0.00-0.12	0.00	0.28	0.00	0.00	0.11	0.00
MgO	10.22	10.22	10.18-10.22	14.58	2.34	0.00	0.00	15.81	11.63
CaO	0.00	0.00	0.00	0.00	0.00	0.12	0.00	0.00	0.00
Na ₂ O	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.16
H ₂ O(calc)	1.48	0.00	-	-	-	-	-	-	-
Total	99.78	98.77		100.42	100.36	98.67	100.74	99.86	97.15
O-basis	31	32		4	3	4	3	20	18
Si	0.000	0.000		0.000	0.009	0.000	0.000	1.516	4.978
Ti	1.520	1.551		0.000	0.958	0.004	0.000	0.004	0.002
Al	14.380	14.673		1.954	0.000	0.017	1.991	8.766	3.991
Cr ₃	0.000	0.000		0.005	0.000	0.000	0.001	0.000	0.000
Fe ³⁺	0.580	1.327		0.041	0.066	1.975	0.010	0.195	-
Fe ²⁺	2.405	1.718		0.418	0.875	0.999	-	0.695	0.269
Zn	0.000	0.000		0.016	0.000	0.000	0.000	0.000	0.000
Mn	0.021	0.022		0.000	0.006	0.000	0.000	0.011	0.000
Mg	3.095	3.158		0.567	0.086	0.000	0.000	2.814	1.765
Ca	0.000	0.000		0.000	0.000	0.005	0.000	0.000	0.000
Na	0.000	0.000		0.000	0.000	0.000	0.000	0.000	0.032
Sum	22.000	22.449		3.001	2.000	3.000	2.002	14.000	11.030
X _{Mg}	0.563	0.648		0.576	0.089	-	-	0.802	0.868
X _{Fe³⁺}	0.194	0.436		0.089	0.070	0.664	-	0.219	-

K, F and Cl below detection limit. Mineral abbreviations after Kretz (1983). Fe³⁺ in spinel, ilmenite, magnetite and sapphirine were calculated assuming perfect stoichiometry. # - H₂O and Fe³⁺ in högbomite calculated respectively on basis of 32 O + 2 OH, 22 cations (Gatehouse and Grey, 1982) and 32 O (Zakrzewski, 1977). X_{Mg} = Mg/(Mg+Fe²⁺) and X_{Fe³⁺} = Fe³⁺/(Fe²⁺+Fe³⁺).

et al., 1983; Beukes *et al.*, 1986, sample 217-6; Gieré, 1986; Grew *et al.*, 1987; Grew *et al.*, 1990). This deficit appears not to be due to BeO or Li₂O, at least in cases where these constituents have been sought (Wilson, 1977; Coolen, 1981; Grew *et al.*, 1987).

Assuming that the deficit is due to OH and possibly Fe³⁺ counted as Fe²⁺ and that the högbomites are 8H polytypes, we also calculated our högbomite analyses on the basis of 30O + 2OH. With recalculated oxide totals close to 100% these assumptions could be valid for sample MA 818, but are not supported by the higher oxide totals in PT 160. Clearly, the assumption of Zakrzewski (1977) gives the preferable formula (10H-polytype?) for PT 160, while the formula for the 8H-polytype is favoured for MA 818, suggesting that the two högbomites belong to different polytypes. This is supported by the large difference in Ti content between the two samples, that is the proportion of spinel-like and nolanite-like layers differs for the two högbomites.

Spinel is principally hercynite-spinel solid

solutions with low gahnite (MA 818: 1.4-3.7 mol. %; PT 160: 0.8-1 mol. %) and magnetite contents (MA 818: 1.5-2.0 mol. %; PT 160: 2.0-2.8 mol. %). All analysed spinels from sample MA 818 show weak compositional zoning with X_{Mg} (0.58-0.61) and ZnO (0.75-1.95wt.%) increasing from rim to core. The unzoned spinels from sample PT 160 are more magnesian with X_{Mg} = 0.67-0.70.

Sapphirine grains of sample PT 160 lie on the join between the 2:2:1 and 7:9:3 compositions in terms of (Mg,Fe²⁺)O:(Al,Fe³⁺,Cr)₂O₃:SiO₂. The compositions of MA 818 sapphirines are slightly more aluminous than the 7:9:3 composition and are therefore termed peraluminous (Schreyer and Abraham, 1975). X_{Mg} values are 0.80-0.83 (MA 818) and 0.89-0.90 (PT 160). The calculated Fe³⁺/(Fe²⁺ + Fe³⁺) ratios are 0.21-0.28 (MA 818) and 0.39-0.44 (PT 160).

The orthoamphibole in sample PT 160 is a gedrite with Al^{IV}, Na (A) and X_{Mg} of 1.75-1.87, 0.40-0.43 and 0.79-0.81 respectively. Stoichiometry on the basis of 15 cations (excluding Na + K, which are assigned to the A-site) suggests

Table 2. Chemical compositions of minerals in sample PT 160, Rangleåsen, Bamble Sector, south Norway.

	Högbomite [#]		range	Spl	Ilm	Mag	Chl	Crn	Rt	Ged	Spr
SiO ₂	0.00	0.00	0.00	0.00	0.00	0.00	26.36	0.00	0.00	44.69	13.29
TiO ₂	5.66	5.66	5.66- 5.75	0.00	53.11	0.00	0.00	0.13	98.50	0.29	0.10
Al ₂ O ₃	62.11	62.11	61.21-62.11	64.71	0.00	0.00	25.21	99.04	0.04	17.87	61.81
Cr ₂ O ₃	0.00	0.00	0.00- 0.05	0.02	0.00	0.00	0.00	0.00	0.02	0.00	0.03
Fe ₂ O ₃	13.93	6.33	-	2.81	0.00	68.02	-	0.98	-	-	3.27
FeO	3.85	10.70	15.61-16.98	15.01	45.83	29.93	5.23	-	0.23	10.23	3.79
ZnO	0.32	0.32	0.25- 0.32	0.55	0.00	0.07	0.00	0.00	0.02	0.00	0.04
MnO	0.04	0.04	0.02- 0.10	0.07	0.00	0.00	0.00	0.00	0.04	0.16	0.08
MgO	14.26	14.26	14.09-14.75	17.55	0.04	0.34	27.82	0.00	0.00	21.60	18.16
CaO	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.56	0.00
Na ₂ O	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	1.60	0.00
H ₂ O(calc)	1.54	0.00	-	-	-	-	-	-	-	-	-
Total	101.71	99.41		100.77	98.98	98.36	84.62	100.15	98.85	97.00	100.57
O-basis	31	32		4	3	4	28	3	2	23	20
Si	0.000	0.000		0.000	0.000	0.000	5.156	0.000	0.000	6.250	1.563
Ti	0.831	0.873		0.000	1.013	0.000	0.000	0.002	0.998	0.031	0.009
Al	14.291	15.023		1.946	0.000	0.000	5.813	1.990	0.001	2.946	8.568
Cr	0.000	0.000		0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.003
Fe ³⁺	2.047	0.977		0.054	0.000	2.000	-	0.013	-	-	0.289
Fe ²⁺	0.629	1.836		0.321	0.973	0.978	0.856	-	0.003	1.197	0.372
Zn	0.046	0.048		0.010	0.000	0.002	0.000	0.000	0.000	0.000	0.003
Mn	0.007	0.007		0.002	0.000	0.000	0.000	0.000	0.000	0.019	0.008
Mg	4.149	4.363		0.667	0.002	0.020	8.112	0.000	0.000	4.503	3.184
Ca	0.000	0.000		0.000	0.000	0.000	0.000	0.000	0.000	0.084	0.000
Na	0.000	0.000		0.000	0.000	0.000	0.000	0.000	0.000	0.434	0.000
Sum	22.000	23.127		3.000	1.988	3.000	19.937	2.004	1.002	15.464	14.000
X _{Mg}	0.868	0.817		0.675	-	0.020	0.905	-	-	0.790	0.895
X _{Fe³⁺}	0.765	0.347		0.144	0.000	0.672	-	-	-	-	0.437

K, Ni, F and Cl below detection limit. See Table 1 for other notes.

0.09-0.24 Fe³⁺ per formula unit and Fe³⁺/(Fe³⁺ + Fe²⁺) of 0.08-0.21. TiO₂ values range 0.29-0.44 wt. %.

Chlorite was only analysed in sample PT 160 and is an Al-rich clinocllore with X_{Mg} and Al^{IV} in the range 0.90-0.91 and 2.45-2.84 respectively. Magnetite contains traces of Ti, Al, Cr, Mn, Mg, Ca and Zn (in PT 160, but below detection limit in MA 818).

Corundum contains 0-0.13 wt. % TiO₂, 0-0.11 wt. % Cr₂O₃ and 0.48-0.88 wt. % Fe₂O₃. Rutile is nearly pure TiO₂ with minor FeO (0.23 wt. %) and traces (<0.15 wt. %) of ZnO (PT 160), MnO, Cr₂O₃ and Al₂O₃. Ilmenite from sample MA 818 contains 2.34-3.00 wt. % MgO and minor MnO (<0.50 wt. %). Ilmenite from sample PT 160 is nearly pure FeTiO₃ with less than 0.15 wt. % MgO and Al₂O₃.

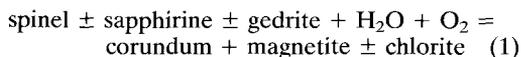
Högbomite genesis

The primary assemblages orthopyroxene-phlogopite-cordierite-spinel-corundum (MA 818), gedrite-spinel (PT 160) and the sapphirine coronas formed by the generalised reactions based on textural relations: spinel + corundum +

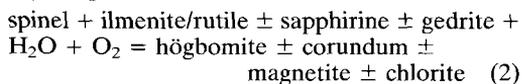
cordierite = sapphirine + vapour (MA 818) and gedrite + spinel + corundum = sapphirine + vapour (PT 160), developed coevally with the upper-amphibolite to granulite facies regional metamorphic event (M3b according to Visser and Senior, 1990) in the Arendal-Tvedestrand area of the Bamble Sector. The second reaction is more Al-rich on the product than on the reactant side suggesting that, although primary corundum is not observed in the sample, corundum must have participated in the reaction. Temperature estimates on the basis of sapphirine-spinel (Owen and Greenough, 1991; with estimated precision of ±100 °C) and orthopyroxene-phlogopite (Sengupta *et al.*, 1990; with estimated precision of ±50 °C) Mg-Fe exchange reactions for this event yield 773-815 °C (MA 818) and 811-844 °C (PT 160) for sapphirine-spinel (rim) pairs and 775-823 °C at 7 kbar for coexisting orthopyroxene-phlogopite pairs (MA 818).

The spatial relationship with spinel in the alteration zones in PT 160 and with the alteration aggregates after spinel in MA 818 suggests a secondary formation of högbomite from the primary high-grade spinel-bearing assemblages. The breakdown of spinel in the absence of Ti-

bearing phases took place according to the generalised reaction:



A prerequisite for the development of large högbomite grains in PT 160 and högbomite in MA 818 is the presence of ilmenite and/or rutile intergrowths with/or inclusions in spinel suggesting the generalised reaction:



The continued stability of spinel and ilmenite and/or rutile outside of the alteration zones in PT 160 implies that the introduction of a hydrous fluid was essential for the production of högbomite (reaction 2). The increased exsolution of magnetite within spinel grains which are cut by the alteration zones, and the high $\text{Fe}^{3+}/(\text{Fe}^{2+} + \text{Fe}^{3+})$ ratio which is calculated for högbomite, suggests increased oxygen fugacity levels associated with the hydrous fluids. Oxygen fugacity levels in sample MA 818 are inferred from the similar $\text{Fe}^{3+}/(\text{Fe}^{2+} + \text{Fe}^{3+})$ ratios of primary and secondary minerals and suggest that very little change accompanied högbomite formation.

Temperature conditions of högbomite-forming reaction 2 can not be determined from the mineralogy of the two samples. Biotite–garnet thermometry in other samples from the Snare-sund lens (Thijssen, 1990) in which cordierite is pinitised and sapphirine is chloritised have therefore been used to estimate the temperature of reaction 2. Adjacent biotite and garnet grains, which are believed to have re-equilibrated during this rehydration event, yield 550–620°C at 6–7 kbar. These P – T conditions are similar to those of the first stages of rehydration during the Sveconorwegian orogeny (M III according to Nijland *et al.*, in prep.; early M₄ according to Visser and Senior, 1990) before uplift occurred. Garnet–core and matrix biotite pairs in the samples analysed by Thijssen (1990) yield 760–825°C at 6–7 kbar, consistent with T –estimates of primary assemblages in MA 818 and PT 160.

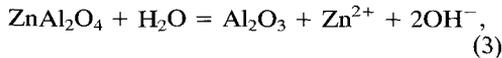
Gedrite could have been a source of Ti for small högbomite grains located on or in the margins of spinel and spinel pseudomorphs and not associated with ilmenite or rutile in sample PT 160 (TiO_2 in spinel is below detection limit). However, Ackerman *et al.* (1983) concluded that a comparable TiO_2 content (up to 0.43 wt.%) of gedrites associated with högbomites in the Fiske-naeset rocks was insufficient to enable högbomite to form in the absence of rutile. Alternatively Ti could have been mobilised along grain

boundaries from nearby (5 to 20 mm distance) alteration aggregates which involve ilmenite and/or rutile. Introduction of Ti from outside the local rock system by Cl^- , F^- and/or PO_4^{3-} -complexes analogous to those described by Gieré (1986) is considered unlikely since apatite and monazite are not associated with högbomite and Cl and F in högbomite and coexisting hydrous phases are below or close to detection limit.

The partitioning of elements between högbomite and spinel appears to be close to unity for Fe–Mg in both samples, with högbomite showing minor enrichment in iron, relative to spinel. The Fe–Mg partitioning is consistent with the partitioning data of Coolen (1981), Gieré (1986) and Rammlmair *et al.* (1988) and supports reaction 2 deduced from petrography. The partitioning data of Zn between högbomite and spinel, expressed in terms of the partitioning coefficient K_D $\text{Zn-Fe}(\text{Hög-Spl}) [K_D = (\text{Zn/Fe})^{\text{Hög}}/(\text{Zn/Fe})^{\text{Spl}}$, with Fe as total Fe], show a moderate depletion in högbomite relative to spinel in sample PT 160 ($K_D = 0.56$ – 0.63), while Zn was not detected in högbomite from sample MA 818 although coexisting spinel contains 0.75–1.95 wt.% ZnO ($K_D = 0.00$). The distribution of zinc between högbomite and spinel pairs from other localities is not uniform. The majority of the occurrences show K_D in the range 0.40–0.75 (e.g. Beukes *et al.*, 1986; Gieré, 1986; Spry and Petersen, 1989; Grew *et al.*, 1990). A very low K_D (0.017) was recorded by Rammlmair *et al.* (1988) and $K_D \geq 1$ was found in the studies by Beukes *et al.* (1986; sample 023 WDB), Grew *et al.* (1989) and Spry and Petersen (1989; sample PS84-132).

In the case of the Bamble högbomite, the difference in Zn distribution for the two samples could be due to their possible difference in polytype. Data available in the literature concerning this point are still scarce and inconclusive, but, the data seem to suggest that Zn-partitioning is independent of the polytype. $K_D < 1$ is recorded for 18R, 8H and most 10H högbomite–spinel pairs (Rammlmair *et al.*, 1988; Petersen *et al.*, 1989; Grew *et al.*, 1990). Some 10H (e.g. Beukes *et al.*, 1986; Grew *et al.*, 1987) and mixed 8H/10H (Petersen *et al.*, 1989) högbomite–spinel pairs are shown to have a $K_D \geq 1$.

The irregular distribution of Zn may depend on the geochemical behaviour of the gahnite component (ZnAl_2O_4) in spinel in assemblages without other Zn-bearing phases such as staurolite. Increase of the gahnite component will stabilise hercynite–spinel solid solutions with decreasing temperatures and increasing f_{O_2} (Spry and Scott, 1986). Increased $a_{\text{H}_2\text{O}}$ will involve decomposition of the gahnite component according to:



liberating zinc which then becomes available for incorporation in hōgbomite. In many localities, hōgbomite is reported to have formed during the cooling or uplift stages following peak metamorphic conditions in amphibolite- and granulite-facies terrains (e.g. Grew *et al.*, 1987; Rammilmair *et al.*, 1988; Grew *et al.*, 1990). The increase of f_{O_2} and/or decrease in temperature associated with this hōgbomite formation will tend to concentrate the gahnite component in the decomposing spinel at contacts with hōgbomite. This may explain the observed Zn-depletion in hōgbomite relative to spinel in many of the occurrences. A good example of this is reported by Rammilmair *et al.* (1988) who observed a discontinuously zoned spinel with extreme Zn-enrichment in a very thin rim adjacent to hōgbomite. Variations in $a_{\text{H}_2\text{O}}$ and f_{O_2} may account for the observed variation in Zn-partitioning. The difference in Zn-partitioning between samples PT 160 and MA 818 is believed to be caused by higher $a_{\text{H}_2\text{O}}$ in sample PT 160. This is supported by more extensive retrogression in PT 160.

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