Osumilite in the sapphirine–quartz terrane of Enderby Land, Antarctica: implications for osumilite petrogenesis in the granulite facies

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Abstract

Osumilite is an important rock-forming mineral at 10 exposures in the Late Archean Napier complex of coastal Enderby Land (50–52°E), Antarctica. Mineral assemblages of low compositional variance (all with quartz±plagioclase) are osumilite–garnet–sillimanite ±spinel, osumilite–sapphirine–sillimanite, osumilite–garnet–orthopyroxene–sillimanite, osumilite–orthopyroxene–biotite, and in a rock nearly free of iron (atomic $Mg/(Mg+Fe)$ = 0.89) from Reference Peak (67°15'S, 50°29'E), osumilite–enstatite ($En_{91}Fs_9$)–sillimanite. The Antarctic osumilite departs from the theoretical composition $(K, Na) (Mg,Fe,Mn)_2Al_3 (Si_{10}Al_2)O_{30}$ in that $(Mg+Fe+Mn)$ ranges from 2.20 to 2.39; Al, 4.45 to 4.67; and Si, 10.05 to 10.30. These variations approximately obey the substitution $(Mg+Mn+Fe) + Si = 2Al$. Atomic $Mg/(Mg+Mn+Fe)$ ratio (X_{Mg}) of Antarctic osumilite ranges from 0.79 to 0.97, and is greater than X_{Mg} of associated cordierite and biotite. Antarctic osumilite-bearing rocks have moderate to high Mg/Fe ratios (atomic $Mg/(Mg+Fe)$ = 0.43 to 0.89) and are peraluminous.

A petrogenetic grid in the system $K_2O–FeO–MgO–Al_2O_3–SiO_2$ for quartz (present in all assemblages), osumilite, K-feldspar, garnet, sapphirine, cordierite, orthopyroxene, and sillimanite can be constructed given the constraint that the univariant reaction osumilite = cordierite + enstatite + K-feldspar + quartz in the iron-free subsystem intersects the stable portion of the univariant reaction in the $MgO–Al_2O_3–SiO_2$ system: cordierite = enstatite + sillimanite + quartz. Pressure–temperature slopes of reactions involving osumilite but not cordierite are positive; slopes of reactions involving both osumilite and cordierite are negative, or in one case, parallel to the pressure axis. Osumilite is predicted to be stable in rocks of appropriate bulk composition at temperatures above 750°C and at total (lithostatic) pressures less than 8 or 9 kbar. M. Olesch and F. Seifert's recent experimental work and Olesch's calculations indicate that osumilite would be stable under hydrous conditions only when $P_{H_2O} = P_{Total} < 1$ kbar; at higher total pressures, H_2O partial pressure must be much less than total pressure. Minimum temperatures for the appearance of osumilite–garnet and osumilite–sillimanite assemblages are in general higher than those needed for osumilite–cordierite–orthopyroxene, and pressures of 5 or 6 kbar may be needed to stabilize osumilite–garnet or osumilite–sillimanite at geologically accessible temperatures. As the physical conditions proposed here for osumilite stability overlap the estimated conditions in many granulite-facies terranes, osumilite can be expected to occur at more localities than have been reported to date.

Introduction

Osumilite is a K–Al–Fe Mg tectosilicate crystallochemically similar to cordierite. Since Miyashiro (1956) originally described osumilite from Sakkabira, Kyushu, Japan, it has been reported in four geologic environments: in volcanic rocks at a number of localities (Rossi, 1963; Olsen and Bunch, 1970; Stankevich, 1974; Miyachi and Miyachi, 1978; Yokomizo and Miyachi, 1978); in rocks metamorphosed at high temperatures by basalt or dolerite (Chinner and Dixon, 1973; Hentschel *et al.*, 1977), in granulite-facies contact aureoles of anorthositic complexes (Berg and Wheeler, 1976; Maijer *et al.*, 1977; Bogdanova *et al.*, 1980); and in granulite-facies rocks of Enderby Land, Antarctica (D. J. Ellis, personal communication in Hensen, 1977a; Grew, 1978; Sheraton *et al.*, 1980; Ellis *et al.*, 1980).

Although osumilite remains a rare mineral, the increasing number of new localities may yet substantiate Miyashiro's (1956) statement that "probably osumilite is a relatively common mineral."

The emphasis of the present paper will be the osumilite-bearing rocks of Enderby Land, Antarctica (Fig. 1). Coastal portions of Enderby Land east of about 46°30'E are underlain by the pyroxene granulite-facies Napier zone or complex of Archean age (Ravich and Kamenev, 1975; Sheraton et al. 1980; Grew and Manton, 1979). Orthopyroxene–quartz–feldspar gneisses ("charnockite" and "enderbite"), pyroxene granulite, and garnetiferous rocks constitute most of the complex; probably no more than 5 percent are well-layered sequences of aluminous and siliceous rocks (Ravich and Kamenev, 1975; Kamenev, 1975; Sheraton et al., 1980; Grew, 1980a). These rocks first attracted the interest of petrologists when Dallwitz (1968) reported the sapphirine–quartz association from Spot Height 945 (Fig. 1).

Osumilite has been reported from 10 exposures in Enderby Land; at some of these, osumilite is an abundant rock-forming mineral (Sheraton et al. 1980; Ellis et al. 1980; Fig. 1, this paper). Ellis (1980) and Ellis et al. (1980) have published detailed discussions of the petrology of osumilite and associated minerals from Dallwitz' (1968) original sapphirine–quartz locality, Spot Height 945.

The present paper, an independent study (preliminary results were presented by Grew, 1978; 1980c,d), is based on field work I did as a member of the Australian National Antarctic Research Expeditions (ANARE) in Enderby Land during January, 1978, and January and February, 1980. The main aims of the paper are to review the chemistry of osumilite with an emphasis on osumilite from granulite-facies rocks, and to discuss the metamorphic conditions under which osumilite can form. The $P–T$ diagrams for osumilite petrogenesis in deep-seated metamorphic rocks proposed here differ in several important respects from the grids proposed by Hensen (1977a) and Ellis et al. (1980).

Antarctic osumilite-bearing rocks

Scope of study

I collected osumilite at six exposures in Enderby Land and samples from three of these exposures, Gage Ridge, Mount Hardy, and Reference Peak (Fig. 1, Table 1), have been studied in detail. J. W. Sheraton and his colleagues of the Australian Bureau of Mineral Resources in Canberra (BMR) discovered the osumilite locality on Gage Ridge and provided me with the information necessary to find this locality in the field. Moreover, J. W. Sheraton provided two samples from Mount Dungey and one from Brusilov Nunataks (Table 1). In addition to the 12 samples listed in Table 1, I have examined osumilite I collected at the above listed exposures, at Mount Riiser-Larsen, and at an unnamed mountain east of Mount Dungey, and an osumilite from Brusilov Nunataks collected and provided by Simon Harley (University of Tasmania). The garnet–osumilite assemblages shown on Figure 1 are based in part on the mineralogy of these other samples.

Physical appearance of Antarctic osumilite

In hand specimens, unaltered Antarctic osumilite is light gray, dark brown (with a purplish cast), or dark blue. It commonly has a distinctive moonstone appearance, which facilitates recognition in the field. This characteristic appears to be related to acicular or lath-shaped inclusions which are roughly oriented parallel to (001). Some Antarctic osumilite has a distinct prismatic cleavage; a few samples also show the indistinct basal cleavage reported by Ellis et al. (1980). These cleavages in blue osumilite lacking the moonstone appearance from Mount Riiser-Larsen serve to distinguish it in the field from cordierite.

Osumilite grains range from a few tenths of a millimeter to over 1 cm across. The osumilite grains are subequant in some cases, or are stubby tablets flattened in the (001) plane. The tablets commonly have a preferred orientation parallel to the foliation of the rock. In thin section, Antarctic osumilite is colorless or dichroic in pale yellow-brown. Absorption is $O > E$ and distribution of the color is patchy in some grains. A few samples are biaxial positive.

Petrography of osumilite-bearing granulites

Osumilite-bearing granulites are medium-grained (0.1 to 4 mm) to coarse-grained (≥ 1 cm) texturally complex rocks. Commonly osumilite is among the coarser-grained minerals in the rock. Foliation defined by concentration of minerals into distinct laminae or preferred orientation are characteristic of some of the samples. The osumilite-bearing rocks are largely free of alteration such as sericite replacing feldspars or pinite replacing cordierite. In all sections studied, osumilite is to some extent replaced by a very fine-grained plumose symplectitic intergrowth of cordierite, K-feldspar, and quartz in
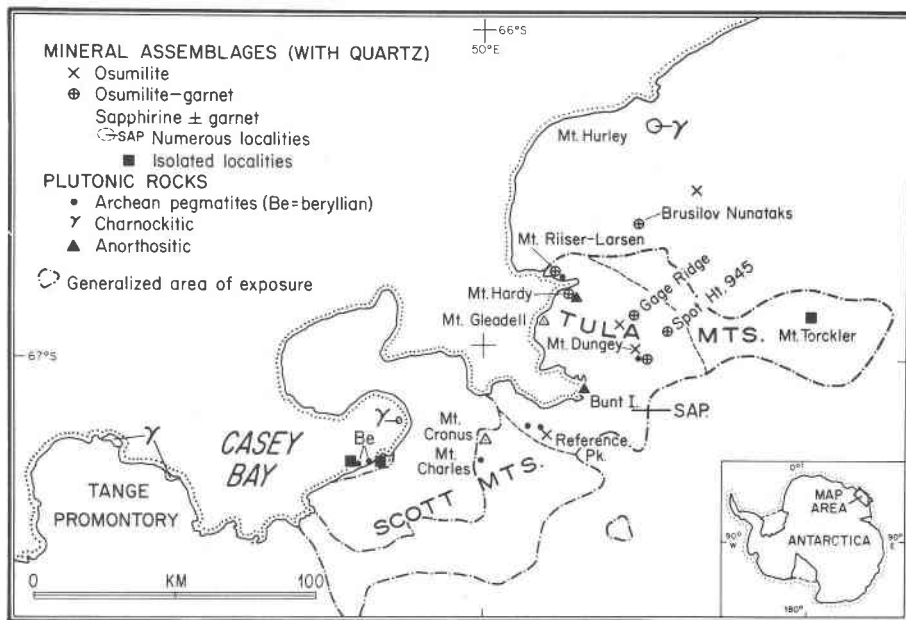
Fig. 1. Map of Enderby Land. Sources of data: Intrusive charnockitic rocks on Tange Promontory (Ravich and Kamenev, 1975, p. 494; Sheraton et al., 1980), east end Casey Bay (J. W. Sheraton, pers. comm., 1980), and north of the Tula Mountains (Ravich and Kamenev, 1975, p. 494). Specific localities of Soviet samples provided by Ye. N. Kamenev (pers. comm., 1978), and of Sheraton et al.'s (1980) two anorthosites by J. W. Sheraton (pers. comm., 1980). Distribution of sapphirine + quartz±garnet and osumilite±garnet assemblages from Grew (1979, 1980 a,c), Sheraton (1980), Sheraton et al. (1980), Ellis et al. (1980), and petrographic notes on rocks I collected during the 1979–80 season. Sapphirine occurs in quartzose granulites at Mount Cronus, but only as inclusions in sillimanite. Identification of pegmatites of possible plutonic origin is based on the presence of beryllium minerals in two pegmatites and of contact aureoles around several others (Grew, 1981).

which are aggregates or dendritic trails of orthopyroxene granules (Figs. 2 and 3, cf. Ellis et al., 1980, Fig. 2C, 2D). The proportion of alteration in the 12 samples studied in detail ranges from a few percent (sample 2045C) to nearly 100%, in which only osumilite relicts are preserved (samples 2046D and 2050A). Biotite occurs sparingly in the symplectite, and sillimanite appears to be a component only in sample 4124 (Table 1). The symplectite develops around the margin of osumilite grains or along cracks, forming microveinlets in the osumilite. The symplectite commonly coarsens into cordierite with vermicular quartz and discrete K-feldspar grains.

Other symplectic intergrowths are vermicular K-feldspar blebs in plagioclase in a zone adjacent to osumilite (sample 2045C) and sillimanite–orthopyroxene intergrowths (sample 2306B). Examples of simple overgrowths are sillimanite around spinel and K-feldspar around sillimanite (2045C). The significance of these intergrowths is not clear. The vermicular K-feldspar and simple overgrowths may be related to retrograde metamorphism, but the sillimanite–orthopyroxene intergrowths in 2306B possibly formed at the peak metamorphic conditions.

As a consequence of these textural complexities, definition of mineral assemblages involving osumilite is not straightforward. In some cases grains of two or three minerals are in direct physical contact and are clearly in textural equilibrium. However, in many osumilite-bearing rocks, cordierite, quartz, or K-feldspar formed from the breakdown of osumilite intervene between osumilite and associated minerals (Fig. 2). In deducing mineral assemblages in the 12 analyzed samples and in associated rocks, I have assumed that the minerals present in an apparently homogeneous layer of rock constituted a stable mineral assemblage under the maximum pressure-temperature (peak) conditions of the pyroxene-granulite-facies event, whether or not mutual grain contacts are preserved. There is no evidence that cordierite was a stable phase (with quartz) at the peak metamorphic conditions, a conclusion also reached by Sheraton et al. (1980) and Ellis et al. (1980). In the rocks studied in detail (Table 1), cordierite grains independent of the symplectitic intergrowths are interpreted to be derived from recrystallization of the intergrowths or from reactions involving other minerals such as sapphirine after the peak of metamorphism. Some K-feldspar

Table 1. Mineralogy of the Antarctic osumilite-bearing rocks and compositions of garnet, feldspar, and sillimanite

	Gage Ridge						Mt. Hardy		Mt. Dungey		Brusilov Nk's.	Reference Peak
	2045C	2045G[1]	2045E*	2045J[1]	2046D*	2050A*	2083C*	2083D*	4249*	4250[2]	4124	2306B*
Quartz[3]	X	X	X	–	X	X	X	X	X	X	X	X
K-feldspar[3]	tr[4]	–	–	–	X[4]	–	X[4]	(X)[4]	–	X[4]	–	–
Plagioclase	(X)[5]	–	–	–	–	–	(X)[4,5]	–	(X)[4]	(X)[4,7]	(X)[6,7]	(X)
Osumilite	(X)	(X)	(X)	(X)	(tr)	(tr)	(X)	(X)	(X)	(X)	(X)	(X)
Sillimanite	(X)	–	(X)	–	X	(X)	(X)	(X)	–	–	(tr)[8]	X
Garnet	(X)[5]	–	(X)	X	(X)	–	(X)[5]	(X)	–	–	–	–
Cordierite[3]	–	–	–	–	–	(X)	(tr)	(tr)	–	tr	–	–
Sapphirine	–	–	–	tr	–	(X)	–	–	–	(X)[9]	–	–
Orthopyroxene[3]	–	(X)	–	X	–	tr	–	–	(X)	(X)	tr	(X)
Biotite	–	tr	–	tr	–	tr	tr	tr	(X)	(X)	–	tr
Spinel	tr	–	(X)	–	(X)	(X)	(tr)	(tr)	–	–	–	–
Rutile	–	tr	X	X	X	X	X	X	tr	tr	–	X
Zircon/Monazite	–	–	–	tr	tr	tr	tr	tr	tr	tr	tr	tr
Ilmenite	–	–	(tr)	–	(tr)	(tr)	–	–	–	–	–	–
Corundum	–	–	–	–	(tr)	tr	–	–	–	–	–	–
Sulfides	–	–	tr	–	–	–	tr[10]	tr[10]	–	–	–	–
Graphite	–	–	–	–	–	–	–	–	–	tr	–	–
COMPOSITION OF GARNET (mole %)												
Almandine	51	–	50	–	54	–	44	44	–	–	–	–
Pyrope	45	–	47	–	44	–	51	51	–	–	–	–
Spessartine	1	–	1	–	1	–	1	1	–	–	–	–
Grossular	3	–	2	–	1	–	4	4	–	–	–	–
COMPOSITION OF FELDSPAR (mole %)												
Orthoclase	2	–	–	–	–	–	1	85	1	3	1	2
Albite	61	–	–	–	–	–	44	11	89	82	78	69
Anorthite	37	–	–	–	–	–	55	3	10	15	21	29
Celsian	0	–	–	–	–	–	–	1	0	–	–	–
COMPOSITION OF SILLIMANITE (weight %)												
Fe_2O_3	0.40	–	0.28	–	–	0.46	0.19	0.24	–	–	1.02	–
Cr_2O_3	–	–	0.21	–	–	0.24	0.33	0.26	–	–	0	–

X = present; tr = trace amounts only; – = absent; ( ) = probe analysis; * = XRF analysis
[1] Mineralogy only of osumilite-bearing part of rock.
[2] Thin section includes several layers of markedly differing mineralogy.
[3] Exclusive of fine-grained components in symplectitic intergrowths replacing osumilite.
[4] Commonly antiperthitic (plagioclase) or perthitic (K-feldspar).
[5] Microprobe data from Grew (1980a).
[6] Abundant antiperthitic lamellae, passing into a mesoperthitic feldspar.
[7] Analysis of lamellae-poor feldspar only.
[8] Confined to symplectic intergrowths replacing osumilite.
[9] No contact with quartz.
[10] Largely altered.

independent of the symplectites may have been part of the peak metamorphic assemblage, for K-feldspar in grains comparable in size with osumilite grains is in contact with osumilite.

Mineral assemblages that have a relatively large number of mineral phases and, consequently, low compositional variance, have been deduced for the rocks listed in Table 1 (problematical phases indicated in parentheses):

quartz–(K-feldspar)±plagioclase–osumilite–sillimanite–garnet–rutile±spinel (e.g. 2083C,D,2046D)

quartz–osumilite–sillimanite–sapphirine–(ilmenite–spinel)–rutile (2050A)

quartz–plagioclase–osumilite–orthopyroxene–biotite–rutile (4249)

quartz–plagioclase–osumilite–sillimanite–orthopyroxene–rutile (2306B).

A rock containing the assemblage quartz–osumilite–garnet–orthopyroxene was collected at Brusilov Nunataks by S. Harley, and I collected two rocks with the same assemblage, plus sillimanite, from Mount Riiser-Larsen.

Whole rock chemistry

X-ray fluorescence analyses (G. Stummer, analyst) of osumilite-bearing rocks and of some associated rocks lacking osumilite are listed in Table 2; these analyses and analyses reported by Ellis et al. (1980) are plotted in Figure 4. The osumilite-bearing rocks are characterized by moderate to high Mg/Fe ratios and by Al in excess of K + Na + 2Ca
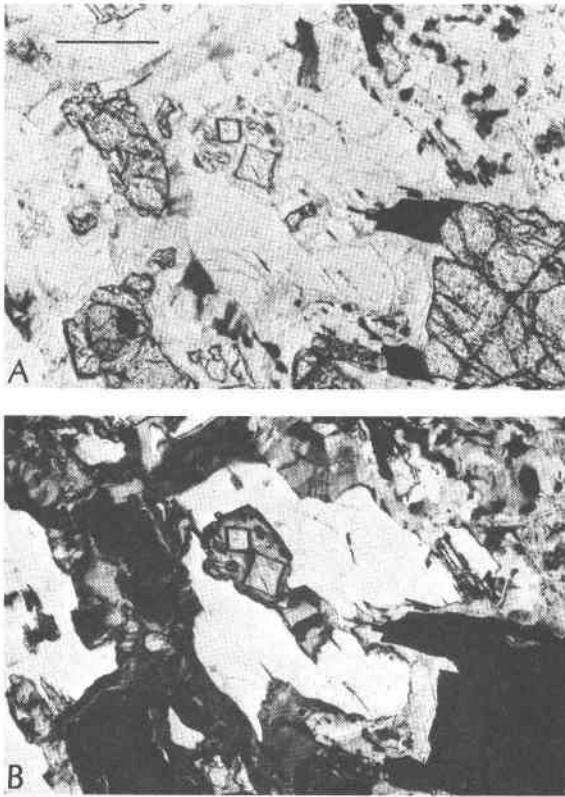
Fig. 2. Photomicrographs of osumilite granulite (2045E) from Gage Ridge, Enderby Land. (A) Plane polarized light. (B) Crossed nicols. Osumilite (white in B) is separated from garnet (high relief) and sillimanite (rhombic outlines) by quartz-cordierite intergrowths, K-feldspar, and fine orthopyroxene (dark). Also present are spinel (rounded inclusion in garnet, lower left) and rutile (dark grains next to garnet). Length of bar 0.3 mm.

43 to 89. In less aluminous osumilite-bearing host rocks, in which orthopyroxene is commonly present, the 100 Mg/(Mg+Fe) ratio is not less than 80. Osumilite-bearing rocks, except sample 4249, have lower Na/K ratios than associated rocks in which antiperthitic plagioclase and mesoperthitic feldspars (samples 2052B and 2083A, Table 2) or sodic oligoclase (sample 459A, Grew, 1980a) but not osumilite, are present.

Several Enderby Land osumilite-bearing rocks also contain certain minor elements in relatively large amounts, *i.e.* Cr and Ni (Table 2 and Ellis *et al.*, 1980), while 4249 has a Ba content of 1880 ppm (Ellis *et al.*, 1980).

The rocks listed in Table 2 are on the average more magnesian and have a notably higher Mg/Fe ratio than average pelitic rocks, features characteristic of many Napier complex metapelites (Sheraton, 1980). Sheraton (1980) attributes the unusual Mg-rich bulk compositions of the Napier complex metapelites to the sedimentary protolith from which the metapelites are derived and found no evidence for significant changes in major element composition during granulite-facies metamorphism.

Chemistry of minerals associated with osumilite

Ferromagnesian minerals associated with osumilite are relatively magnesian and contain little Mn (Tables 1 and 3 to 7) (for analytical methods, see below). Other notable features are the high Al content of orthopyroxene (Table 3, also Ellis *et al.*, 1980), low Ca contents of garnet (Table 1, see also
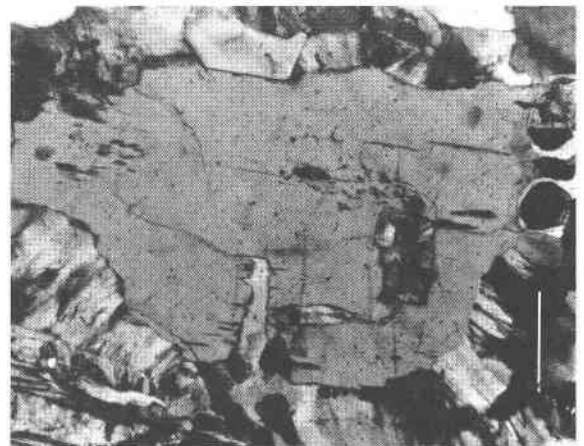
("Al$_2$O$_3$" in Figure 4), *i.e.* in excess of the amount required to form feldspar. Six of nine bulk compositions of osumilite-bearing rocks plot within the triangular areas bounded by the minerals constituting the rock. One rock (sample #2083D, labeled 68 in Fig. 4) plots on a boundary. Its composition is consistent with the mineralogy if K-feldspar is included in the mineral assemblage (see above), for the composition would then plot inside the tetrahedron K-feldspar–osumilite–garnet–sillimanite. The remaining two rock compositions (2050A and 2306B, labeled 74 and 89, respectively) plot in adjacent fields close to the boundary. This disagreement may be due to analytical errors compounded in the calculation of "Al$_2$O$_3$" or to the presence of minerals not included in the assemblage (*e.g.* secondary cordierite in 2050A and biotite in 2306B).

Atomic Mg/(Mg+Fe) ($\times$ 100) values range from

Fig. 3. Photomicrograph of an osumilite grain and of very fine symplectic intergrowth replacing it (sample 2045E). Indistinct basal cleavage is present in osumilite. Crossed nicols. Length of bar is 0.2 mm.

Table 2. Composition of osumilite-bearing and associated rocks (X-ray fluorescence analyses)

Oxide	Gage Ridge				Mount Hardy			Mount Dungey	Reference Peak
	2045E[1]	2046D[1]	2050A[1]	2052B[2]	2083A[3]	2083C[1]	2083D[1]	4249[1,4]	2306B[1,5]
					WEIGHT PERCENT				
SiO_2	54.06	77.67	52.50	63.80	63.86	71.83	58.42	73.4	74.68
TiO_2	0.89	0.80	1.36	0.03	0.06	0.49	0.78	0.72	0.52
Al_2O_3	24.76	10.67	29.83	18.83	15.08	14.80	29.14	11.42	13.80
Fe_2O_3[6]	12.00	6.10	5.70	3.95	6.43	4.60	3.45	2.07	0.82
MnO	0.126	0.101	0.056	0.067	0.038	0.081	0.128	0.01	0.0083
MgO	4.58	2.72	8.18	1.96	4.97	3.00	3.75	4.55	3.32
CaO	0.23	0.18	0.02	1.27	1.35	1.91	0.36	0.70	1.77
Na_2O	0.46	0.17	0.18	4.61	4.11	0.80	0.38	2.34	2.99
K_2O	2.65	1.27	1.29	4.69	3.15	1.66	2.85	2.20	1.11
P_2O_5	0.012	0.025	0.01	0.01	n.d.	0.01	0.015	0.02	0.03
L.O.I.[7]	0	0.15	0.44	0.32	0.33	0.62	0.40	0.78	0.50
TOTAL	99.92	99.93	99.71	99.56	99.41	99.84	99.74	98.22	99.58
					PARTS PER MILLION				
Cr	1060	540	1080	10	17	240	500	10	13
Ni	309	91	187	40	34	70	73	36	8
Sr	67	6	40	104	194	50	4	37	278
Rb	133	67	128	109	54	82	85	51	n.d.
X(Mg)[8]	0.430	0.469	0.740	0.496	0.605	0.564	0.683	0.813	0.889

n.d. - not detected

Analyst: G. Stummer, except for 4249. Totals are sums only of the 10 major elements, L.O.I. and of the 4 minor elements reported here.

[1] Mineralogy given in Table 1.

[2] Mineralogy: dominant plagioclase + mesoperthitic feldspars; subordinate orthopyroxene; minor spinel, sapphirine, garnet, and quartz; traces of corundum and opaque; secondary cordierite and biotite.

[3] Mineralogy: dominant plagioclase + mesoperthitic feldspars and orthopyroxene; minor sapphirine and quartz; secondary cordierite and biotite.

[4] From Ellis et al. (1980, Table 2). BMR No. 77284249. L.O.I. for this sample is 0.62 wt % H_2O^+, 0.13 wt % H_2O^-, and 0.03 wt % CO_2. 1880 ppm Ba also reported.

[5] Average of two analyses.

[6] All Fe as Fe_2O_3.

[7] Loss on ignition.

[8] Atomic ratio Mg/(Mg + Fe).

Grew, 1980a and Ellis *et al.,* 1980), high Ba and F contents of biotite (Table 4), and low alkali contents of secondary cordierite (Table 5). The Al content of biotite is insufficient to fill the tetrahedral site, indicating the absence of solid solution towards eastonite. In most samples, the Fe/Mg ratio in the cores of grains of the ferromagnesian silicates do not vary significantly from grain to grain in a given sample. However, in some samples, compositions do vary; for example, that of orthopyroxene in 2306B (Table 3).

Spinels contain significant Cr_2O_3 (up to 3.5 wt.% in 2083C; Ellis *et al.,* 1980, report 11%) and ZnO (up to 0.6 wt.%, Table 6). Some spinel grains in samples 2045E and 2050A were analyzed for nickel; these grains contain 0.2 wt.% NiO. Zn and Ni may also be present in spinels in the other samples, but these were not analyzed for Zn and Ni (significant Zn detected in element scan of spinel in 2083D). More-

over, Cr_2O_3 contents vary significantly from grain to grain in 2083C and 2083D as well as in 2045E and 2046D (Table 6).

Corundum in sample 2046D contains 0.7 wt. % Fe_2O_3 and 0.3 wt. % Cr_2O_3. Ilmenite in 2045E, 2046D, and 2050A contains little Fe_2O_3, *e.g.* its composition in 2050A is $(Fe_{0.90}^{2+}Mg_{0.09}Fe_{0.01}^{3+})$ $(Fe_{0.01}^{3+}Ti_{0.99})O_3$. Sillimanite, except that in sample 4124, is poor in Fe_2O_3 (Table 1) relative to other Antarctic sillimanites (0.25–1.36 wt.% Fe_2O_3, Grew, 1980b).

Characteristic of the Enderby Land sapphirine (Table 7), spinel, ilmenite, and sillimanite are their low Fe_2O_3 contents. This implies a low host rock Fe_2O_3 content, which is consistent with the presence of graphite in one sample (4250). An exception is sample 4124 from Brusilov Nunataks, which may contain significant Fe_2O_3, as sillimanite in this rock has a relatively high Fe_2O_3 content of 1 wt.%.
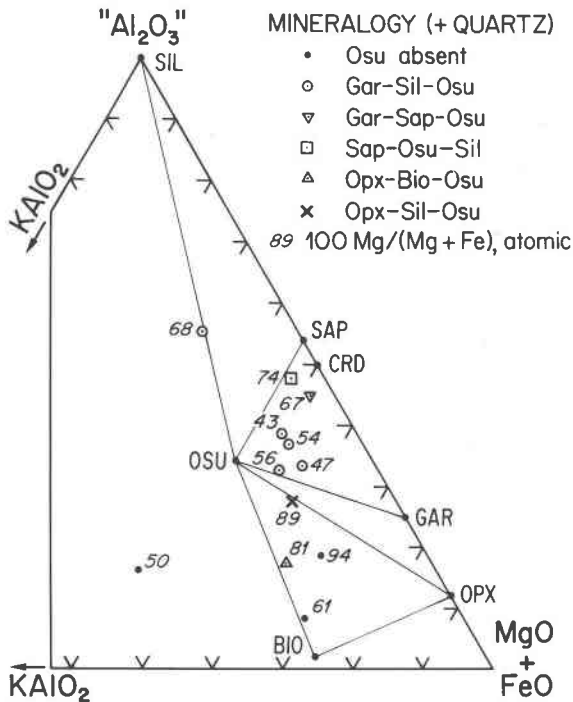
Fig. 4. Plot of host rock compositions in relative molecular proportions of "Al_2O_3," $KAlO_2$ and $MgO + FeO$. "Al_2O_3" is alumina content in excess of that required to make feldspar, *i.e.* 2"Al_2O_3" = $Al–K–Na–2Ca$. All Fe as FeO. Sources of data: Table 2, except samples marked 67 and 54, which are samples 76283350A and 77284348, respectively, from Ellis *et al.* (1980), and that marked 94, which is sample 459A, from Grew (1980a). Abbreviations: osu = osumilite; gar = garnet; sil = sillimanite; qtz = quartz; sap = sapphirine; opx = orthopyroxene, and bio = biotite.

By and large, Fe/Mg ratios of the associated minerals (core compositions) increase with increasing Fe/Mg of osumilite (Fig. 5), a relation consistent with attainment of chemical equilibrium. However, the distribution is not as regular as in sapphirine–quartz rocks from the same area (Grew, 1980a, Fig. 4). Osumilite has a lower Fe/Mg ratio than associated cordierite or biotite. Osumilite Na/K ratio is lower than K-feldspar Na/K ratio in sample 2083D, the same relation reported by Berg and Wheeler (1976), and is higher than biotite Na/K ratio in the Mount Dungey samples.

Conditions of formation

In an earlier paper (Grew, 1980a), I estimated that the temperatures and pressures of the pyroxene–granulite facies metamorphism in the Tula Mountains, where osumilite is most abundant (Fig. 1), were 900°C and 7 kbar, with water pressure much less than total pressure. These estimates were

based on Al content of orthopyroxene associated with sapphirine, compositions of coexisting clinopyroxene and orthopyroxene, compositions of garnet and plagioclase associated with sillimanite and quartz, and experimental data. Ellis (1980) and Sheraton *et al.* (1980) estimate metamorphic conditions to be 8–10 kbar and 900–980°C at Spot Height 945. Sheraton *et al.*, (1980) report calcic mesoperthites (some near $Or_{32}Ab_{50}An_{18}$) which they interpret to have crystallized at temperatures in excess of 920°C and water pressures less than 0.5 kbar.

Another indication of high metamorphic temperatures is complex intergrowths of orthopyroxene and clinopyroxene suggestive of inverted pigeonite. These intergrowths were found in a layer 1.7 m thick of quartz–pyroxene rock at Mount Riiser-Larsen, and in a layer of similar rock 1.5 m thick

Table 3. Composition of orthopyroxene

Oxide	Gage Ridge 2045G*	Mount Dungey 4249	Mount Dungey 4250	Reference Peak 2306B(1)[1]	Reference Peak 2306B(2)[2]
	WEIGHT PERCENT				
SiO_2	49.60	47.35	48.93	52.82	~54.59
TiO_2	0.27	0.30	0.35	0.09	0.06
Al_2O_3	9.58	11.58	11.40	8.75	6.78
Cr_2O_3	0.17	–	–	–	–
FeO	17.50	18.70	17.40	5.43	4.77
MnO	0.16	0.09	0.12	0.07	0.09
MgO	23.27	22.84	23.60	32.69	33.94
ZnO	0.08	–	–	<0.05	<0.05
BaO	0.01	0.02	0	–	–
CaO	0.12	0.06	0.13	0.04	0.03
Na_2O	0.02	0.01	0.02	0.01	0.01
K_2O	0.01	0.05	0.04	0.05	0.05
TOTAL	100.79	101.00	101.99	99.95	100.32
	CATIONS PER 6 OXYGENS				
Si	1.788	1.717	1.742	1.815	1.861
Ti	0.007	0.008	0.009	0.002	0.002
Al	0.407	0.495	0.479	0.354	0.272
Cr	0.005	–	–	–	–
Fe	0.528	0.567	0.518	0.156	0.136
Mn	0.005	0.003	0.004	0.002	0.003
Mg	1.251	1.234	1.252	1.674	1.725
Zn	0.002	–	–	–	–
Ca	0.004	0.002	0.005	0.001	0.001
Na	0.001	0.001	0.001	0.001	0.001
TOTAL	3.999	4.028	4.010	4.006	4.001
X_{Fe}	0.297	0.315	0.293	0.085	0.073

*UM analysis
All Fe as FeO
[1]Average of two grains in one part of section of sample 2306B.
[2]Third grain in another part of same section; a fourth grain contains SiO_2 - 53.69 wt %, Al_2O_3 - 7.73%, FeO - 5.10%, and MgO - 33.40%.

Table 4. Composition of biotite from Mount Dungey

Oxide	Sample Number 4249	4250
	WEIGHT PERCENT	
SiO_2	39.77	41.32
TiO_2	4.29	2.98
Al_2O_3	13.64	13.12
FeO	6.64	5.32
MnO	0.02	0
MgO	21.90	23.14
CaO	0.08	0.09
BaO	1.45	0.46
Na_2O	0.18	0.12
K_2O	9.85	10.38
F	4.83	5.29
Cl	0.34	0.25
TOTAL[1]	100.88	100.18
	CATIONS PER 44 ANIONIC CHARGES (Foster, 1960)	
Si	5.628	5.817
Al^{IV}	2.276	2.177
Total	7.904	7.994
Al^{VI}	0	0
Ti	0.457	0.315
Fe	0.786	0.626
Mn	0.002	0
Mg	4.621	4.857
Total	5.866	5.798
Ca	0.012	0.014
Ba	0.081	0.026
Na	0.049	0.032
K	1.778	1.864
Total	1.920	1.936

All Fe as FeO
[1]*Corrected for oxygen equivalents of halogens*

(containing magnetite as well) at Mount Gleadell (Figure 1). Field relations and mineralogy of these layers indicate that the protolith is a sedimentary rock and that the pigeonite, from which these intergrowths are believed to be derived, is metamorphic. At both localities, the intergrowths consist of coarse orthopyroxene with stout clinopyroxene lamellae visible to the naked eye as well as microscopic exsolution lamellae. The stout lamellae cut grain boundaries of the orthopyroxene (Fig. 6), a feature characteristic of pyroxenes formed from exsolution and subsequent inversion of pigeonite (Bonnichsen, 1969, Fig. 13-14). G. L. Nord, Jr., (personal communication, 1981) analyzed pyroxenes in sample 2322B from Mount Gleadell. Representative compositions of orthopyroxene are $En_{36}Fs_{63}Wo_1$ and $En_{36}Fs_{61}Wo_3$, and of clinopyroxene, $En_{29}Fs_{24}Wo_{47}$ and $En_{28}Fs_{24}Wo_{48}$; correspond-

ing MnO contents are 0.38 and 0.37, 0.01 and 0.17 wt.%, respectively. Recent experimental data suggest that pigeonite containing little MnO would crystallize in a metamorphic environment at temperatures of at least 825°C (Lindsley *et al.*, 1974; Lindsley and Grover, 1980; Podpora and Lindsley, 1979; Vocke *et al.*, 1981).

Osumilite: chemical and X-ray data

Methods

Compositions of osumilite and associated minerals in carbon-coated polished thin sections were measured with a fully automated (F and Cl on manual) ARL-EMX wavelength dispersive electron microprobe at UCLA and a fully automated JXA-5A wavelength dispersive electron microprobe at the University of Melbourne (UM) (Operating conditions: 15 kV on both microprobes; 20 nA sample current at UCLA; other details given in Grew, 1980a). The microprobe data are supplemented by semi-quantitative emission spectrographic analyses obtained (courtesy of R. V. Gaines) on two hand-picked samples of osumilite.

The compositions of the minerals analyzed at

Table 5. Composition of cordierite

Oxide	Locality/ Sample Number Gage Ridge 2050A	Mount Hardy 2083C	2083D
	WEIGHT PERCENT		
SiO_2	50.61	50.82	50.28
TiO_2	0.01	0.02	0.01
Al_2O_3	34.21	33.91	33.79
FeO	3.55	3.22[1]	3.22
MnO	0	0.02	0.02
MgO	11.89	12.07	11.87
CaO	0.04	0.08	0.11
Na_2O	0.03	0.03	0.01
K_2O	0.03	0.06	0.05
TOTAL	100.39[2]	100.23	99.36
	CATIONS PER 18 OXYGENS		
Si	4.991	5.013	5.004
Ti	0.001	0.001	0
Al	3.978	3.944	3.964
Fe	0.293	0.265	0.268
Mn	0	0.002	0.001
Mg	1.748	1.774	1.761
Ca	0.004	0.009	0.012
Na	0.005	0.006	0.002
K	0.004	0.007	0.006
TOTAL	11.024	11.020	11.018

All Fe as FeO.
[1]*Some variation in FeO content from grain to grain.*
[2]*Includes 0.02 wt % BaO.*

Table 6. Composition of spinel from Gage Ridge

Oxide	Sample Number/(Grain Number)			
	2045E		2046D	
	(1)	(2)	(1)	(2)
	WEIGHT PERCENT			
SiO_2	0.08	0.18	0.58	0.55
TiO_2	0	0.02	0	0
Al_2O_3	61.61	61.94	59.60	60.86
Cr_2O_3	2.36	1.37	2.90	1.75
FeO^1	21.76	22.32	24.47	24.46
MnO	0.04	0	0.07	0.07
MgO	13.07	12.42	11.01	11.19
ZnO	0.60	0.44	0.30	0.25
CaO	0	0.02	0	0.01
Na_2O	0.01	0.03	0.02	0.01
K_2O	0.01	0	0.03	0.01
TOTAL	99.54	98.74	98.98	99.16
	CATIONS PER 4 OXYGENS			
Si	0.002	0.005	0.016	0.015
Ti	0	0	0	0
Al	1.936	1.959	1.909	1.936
Cr	0.050	0.029	0.062	0.037
Fe^{3+2}	0.012	0.007	0.013	0.012
Total	2.000	2.000	2.000	2.000
Fe^{2+2}	0.473	0.494	0.543	0.540
Mn	0.001	0	0.002	0.002
Mg	0.519	0.497	0.446	0.450
Zn	0.012	0.009	0.006	0.005
Total	1.005	1.000	0.997	0.997
Na	0.001	0.002	0.001	0.001
K	0	0	0.001	0
Ca	0	0.001	0	0
TOTAL CATIONS	3.006	3.002	2.999	2.999

[1] *Total iron as FeO.*
[2] *From stoichiometry*

UCLA are in general an average of 3–4 grains per sample (rarely only 1–2 grains, except cordierite); each grain was analyzed at 2 to 5 points (generally 3 to 4 points). Compositions of minerals measured at UM are an average of 1–3 grains, each of which was analyzed at 1 or 2 points. Compositions of osumilite in samples 2045C and 2045G are averages of 10 and 7 analyses, respectively, with the UCLA microprobe (4 and 5 with the UM microprobe) of one grain at 3 spots close to one another. The UCLA analyses were repeated during each session on the electron microprobe. The maximum and minimum values reported in Table 8 for 2045C are a measure of the precision and reproducibility of the UCLA microprobe analyses. In general, the range of values for the UCLA analyses of 2045C are about ±2–3% of the average value for elements present in excess of 3 wt.% (Table 8). By comparison, the

range of values for 10 analyses on the UM probe (during several sessions) of three grains in sample 2045E are ±1 to 2.5% of the average values for SiO_2, MgO, and Al_2O_3, ±7% for K_2O and ±12% for FeO.

There are systematic differences between the two sets of analyses; UCLA analytic totals are higher and in the recast analyses, Al in $T1$ and total M cations are higher. For this reason, in the discussion on osumilite chemistry the two sets of analyses are treated separately. The UCLA analyses of a Nain osumilite (U.S. National Museum #143967) compare favorably to Goldman and Rossman's (1978) microprobe analyses and to Jarosewich et al.'s (1979) wet chemical analyses on material from the same bulk sample (J. H. Berg, personal communication, 1980, confirmed that all these analyses were done on the same bulk sample). The exception is
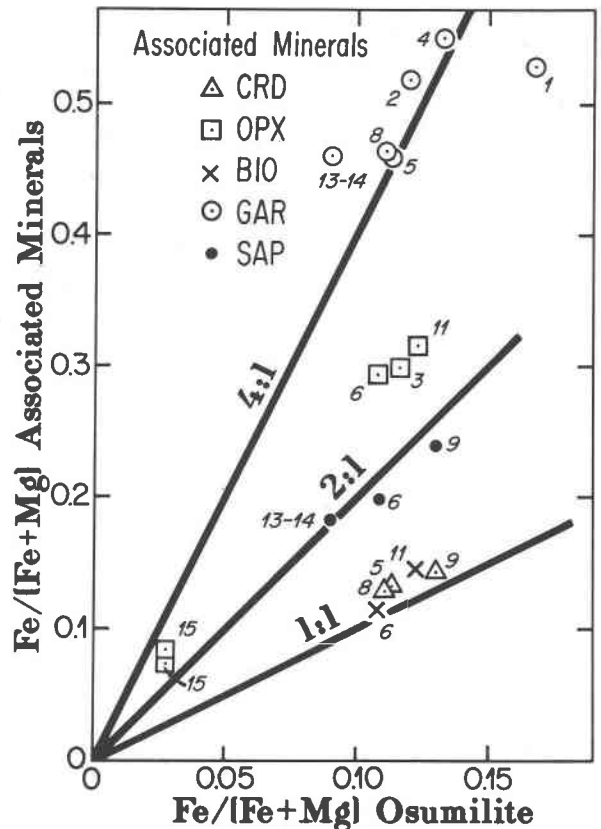

Fig. 5. Atomic Fe/(Fe+Mg) ratios of osumilite and associated sapphirine (sap), cordierite (crd), orthopyroxene (opx), biotite (bio), and garnet (gar). Number references are to samples as follows (same as Figures 7–9): 1 = 2045C, 2 = 2045E, 3 = 2045G, 4 = 2046D, 5 = 2083D; 6 = 4250; 8 = 2083C; 9 = 2050A; 11 = 4249; 13–14 = 76283358 of Ellis et al. (1980); and 15 = 2306B(1) and 2306B(2). Lines are for reference only.

Table 7. Composition of sapphirine

| Oxide | Locality / Sample Number | |
	Gage Ridge 2050A	Mount Dungey 4250
	WEIGHT PERCENT	
SiO_2	13.32	14.21
TiO_2	0.09	0.09
Al_2O_3	60.80	61.60
Cr_2O_3	0.52	0.01
FeO^{1}	9.79	8.12
MnO	0.06	0.05
MgO	15.53	16.86
NiO	0.03	-
CaO	0.08	0.09
K_2O	0.05	0.04
Na_2O	0.01	0.01
TOTAL	100.28	101.08
	CATIONS PER 20 OXYGENS	
Si	1.601	1.677
Ti	0.009	0.008
Al	8.618	8.571
Cr	0.050	0.001
Fe^{3+2}	0.112	0.058
Fe^{2+2}	0.873	0.743
Mn	0.006	0.005
Mg	2.785	2.966
Ni	0.003	-
TOTAL	14.056	14.029

[1] All Fe as FeO.
[2] Calculated from stoichiometry (Higgins et al., 1979; Grew, 1980a).

SiO_2 for which the UCLA probe value is significantly higher (Table 8). The systematic differences in the CaO content between the UCLA and UM analyses (Tables 8 and 9) may be due to differences in the way background was measured on the two microprobes. The wet chemical and spectrographic analyses (Table 10) suggest that CaO contents may not exceed 0.1 wt.%.

In some cases, cation subtotals and totals disagree by one significant figure with the sums of the cations as listed in the tables. This disagreement is due to rounding off of the values for each cation independently of rounding off cation subtotals and totals.

Cell dimensions of two osumilites were determined on powdered osumilite mixed with a silicon internal standard dried from a slurry on a glass slide. The 2θ values were recorded with graphite-crystal monochromated $CuK\alpha$ radiation at a scanning speed of 0.25° 2θ per minute; all reflections were scanned twice (once with increasing 2θ, once with decreasing 2θ). Twenty-five and 29 reflections (for samples 2045C and 2306B, respectively) between 17° and 81° 2θ were used to calculate cell parameters with a least squares computer program. The powder patterns were indexed assuming systematic extinctions of $l = 2n + 1$ for $h0l$ and hhl reflections consistent with space group $P6/mcc$ (Schreyer and Seifert, 1967; Brown and Gibbs, 1969).

Chemical composition

Electron microprobe and emission spectrographic analyses of osumilite from granulite-facies rocks of Enderby Land, Norway, and Labrador indicate that SiO_2, Al_2O_3, TiO_2, FeO, MgO, BaO, Na_2O, K_2O and possibly CaO may be present in amounts ≥ 0.1 wt.% (Tables 8, 9 and 10; also Berg and Wheeler, 1976; Maijer *et al.*, 1977; Ellis *et al.*, 1980). Other components, such as Cr_2O_3 and ZnO, appear to be present in trace amounts only; these elements have been included in the calculation of the structural formulae for completeness. Although beryllium and lithium-bearing minerals structurally similar to osumilite exist, *e.g.* milarite and brannockite, Be and Li were not detected in two samples of Antarctic osumilite (Table 10).

Compositions of osumilite from granulite-facies rocks approach the theoretical end-member $KMg_2Al_3(Si_{10}Al_2)O_{30}$ synthesized by Schreyer and Seifert (1967). Osumilite compositions can be best understood in terms of the 5 major crystallographic sites for cations in the osumilite group of minerals (Brown and Gibbs, 1969). (1) Tetrahedral positions ($T1$) forming double rings occupied by Si and minor Al; (2) tetrahedral positions ($T2$) linking the ring tetrahedra, occupied by Al and possibly Mg; (3) octahedral positions (M) occupied by the divalent ions, Cr, and Ti; and (4) two large sites in the channels, one of which ($C1$) is occupied by K, Na, Ca and Ba.

A Mössbauer spectrum of osumilite from sample 2045C may be fit with three doublets (Table 11). The isomer shifts and quadrupole splittings of the two Fe^{2+} doublets are similar to those for Nain osumilite reported by Goldman and Rossman (1978). According to these authors' assignments, the outer doublet corresponds to Fe^{2+} in sixfold coordination, and the inner one, to Fe^{2+} in a channel site. The parameters for the third doublet are similar to those for the EE' doublet in Goldman and Rossman's (1978) spectrum for Obsidian Cliffs osumilite, but are poorly constrained because of low Fe^{3+} concentration in sample 2045C. The iso-

Table 8. Composition of osumilite: comparison of microprobe data and wet chemical data

	GAGE RIDGE, ANTARCTICA								NAIN, LABRADOR		
	2045C[1]				2045G[1]		2045E[2]		USNM #143967[3]		
	MICROPROBE										W.C.
	UCLA			UM	UCLA	UM	UCLA	UM	UCLA	C.I.T.[4]	Smith'n[5]
	AVE	MAX[6]	MIN[6]	AVE	AVE	AVE	AVE	AVE	AVE	AVE	AVE
WEIGHT PERCENT											
SiO_2	61.50	62.80	59.69	61.82	62.38	61.77	61.98	61.35	61.79	60.8	60.20
TiO_2	0.10	0.17	0.06	0.13	0.12	0.12	0.13	0.08	0.14	0.2	0.18
Al_2O_3	23.84	24.35	23.33	23.44	23.06	22.27	24.31	23.31	22.52	22.2	22.60
Cr_2O_3	0	0.01	0	0.02	0.01	0.03	–	0.02	–	–	–
FeO	2.76	2.87	2.67	2.69	1.95	1.88	2.10	1.89	6.65	6.4	6.38
MnO	0.02	0.04	0.01	0.06	0.02	0.04	0.01	0.04	0.11	0.1	–
MgO	7.67	7.77	7.59	7.22	8.47	8.04	8.06	7.78	5.83	5.7	5.83
ZnO	–	–	–	0.07	–	0.06	–	0.05	–	–	–
BaO	0.15	0.18	0.06	0.14	0.25	0.25	0.17	0.12	0.28	–	–
CaO	0.10	0.14	0.04	0	0.12	0	0.11	0	0.12	–	<0.03
Na_2O	0.28	0.30	0.27	0.28	0.31	0.34	0.28	0.28	0.41	0.3	0.39
K_2O	4.41	4.50	4.26	4.53	4.23	4.28	4.40	4.46	4.12	4.2	4.00
TOTAL	100.83	102.14[7]	98.86[7]	100.40	100.92	99.08	101.55	99.38[8]	101.97	99.9	99.60
CATIONS PER 30 OXYGENS											
T1 Si	10.120			10.211	10.215	10.299	10.096	10.199	10.230	10.30	10.169
T1 Al	1.880			1.789	1.785	1.701	1.904	1.801	1.770	1.70	1.831
TOTAL	12.000			12.000	12.000	12.000	12.000	12.000	12.000	12.00	12.000
T2 Al	2.743			2.774	2.666	2.675	2.763	2.766	2.624	2.66	2.670
M Cr	0			0.003	0.001	0.004	–	0.002	–	–	–
M Ti	0.013			0.016	0.014	0.015	0.015	0.009	0.017	–	0.023
M Fe	0.380			0.372	0.267	0.262	0.285	0.263	0.920	0.90	0.901
M Mn	0.003			0.008	0.003	0.006	0.001	0.006	0.015	0.01	–
M Mg	1.882			1.778	2.067	1.999	1.958	1.929	1.439	1.43	1.468
M Zn	–			0.008	–	0.007	–	0.006	–	–	–
TOTAL	2.278			2.184	2.352	2.294	2.260	2.216	2.392	2.34	2.392
C1 Ba	0.010			0.009	0.016	0.016	0.011	0.007	0.018	–	–
C1 Ca	0.018			0	0.021	0	0.020	0	0.021	–	0
C1 Na	0.090			0.091	0.098	0.111	0.090	0.091	0.132	0.10	0.128
C1 K	0.925			0.955	0.884	0.910	0.914	0.945	0.870	0.89	0.862
TOTAL	1.043			1.054	1.018	1.037	1.034	1.043	1.041	0.99	0.990
TOTAL CATIONS	18.064			18.012	18.036	18.006	18.057	18.025	18.057	17.99	18.052

All Fe as FeO; 0 = <0.005%; – = not analyzed; Ave = Average; W.C. = wet chemical analysis
[1] University of California, Los Angeles (UCLA) and University of Melbourne (UM) analyses made at same spots on osumilite grains.
[2] UCLA and UM analyses made on different sections cut from same sample.
[3] UCLA and California Institute of Technology (C.I.T.) microprobe analyses and Smithsonian Institution (Smith'n) wet chemical analyses made on samples from same larger specimen collected by J. H. Berg (Berg, pers. comm., 1980).
[4] Goldman and Rossman (1978). Ti not included in calculation of formula.
[5] Jarosewich et al. (1979). Includes 0.02% H_2O.
[6] Maximum and minimum values from ten analyses, except BaO (six analyses) and Cr_2O_3 (4 analyses).
[7] These values are maximum (or minimum) totals and are not sums of the maximum (or minimum) values of the oxides.
[8] NiO less than 0.01%.

mer shift and quadrupole splitting suggest Fe^{3+} in octrahedral coordination. The proportion of Fe^{2+} the octahedral site to total Fe in the sample, 61%, is close to the 68% in the Nain osumilite (two doublet fit) studied by Goldman and Rossman (1978), although the total iron content of their sample is more than twice as great (6.5% vs. 2.7%). No ferric iron was reported in the Nain osumilite, and only 5% is present in sample 2045C. It is unlikely that any of the other Antarctic osumilites, except possibly sample 4124 from the Brusilov Nunataks, contains more Fe^{3+} than 2045C, as the Fe_2O_3 contents of associated sillimanite are as low, or lower, than that of sillimanite in sample 2045C (Table 1). The relatively

Table 9. Composition of osumilite

| | ANTARCTICA | | | | | | | | | NORWAY | |
| | Gage Ridge | | | Mt. Hardy | | Mt. Dungey | | Brusilov Nks. | Reference Peak | Rogaland | |
Oxide	2045J*	2046D*	2050A	2083C	2083D	4249[1]	4250[1]	4124[1]	2306B	OS-17[2]	OS-99[2]
	WEIGHT PERCENT										
SiO_2	62.19	60.43	62.01	62.60	62.87	60.86	62.90	61.15	62.20	62.59	61.99
TiO_2	0.08	0.14	0.12	0.09	0.11	0.14	0.14	0.06	0.03	0.09	0.07
Al_2O_3	22.65	23.29	23.61	23.36	23.71	23.88	24.01	23.77	23.06	23.13	23.05
Cr_2O_3	0.01	0.01	–	0	0	–	–	–	0.01	–	–
FeO	2.02	2.05	2.16	1.85	1.84	2.07	1.80	3.69	0.45	4.30	4.14
MnO	0.03	0.04	0.05	0.02	0.01	0.01	0	0	0	0.07	0.07
MgO	7.97	7.50	8.11	8.30	8.13	8.31	8.31	7.62	9.29	7.13	7.20
ZnO	0.13	0.03	–	–	–	–	–	–	–	–	–
BaO	0.11	0.05	0.24	0.09	0.06	0.24	0.03	0.11	–	0	0.06
CaO	0	0	0.09	0.10	0.09	0.12	0.11	0.15	0.01	0.12	0.12
Na_2O	0.34	0.33	0.30	0.21	0.18	0.42	0.29	0.33	0.29	0.51	0.51
K_2O	4.36	4.42	4.42	4.39	4.37	4.26	4.51	4.29	4.34	4.08	4.26
TOTAL	99.93[3]	98.29	101.11	101.01	101.37	100.31	102.10	101.17	99.68	102.02	101.47
	CATIONS PER 30 OXYGENS										
T1 Si	10.284	10.163	10.157	10.222	10.218	10.053	10.165	10.073	10.223	10.220	10.191
T1 Al	1.716	1.837	1.843	1.778	1.782	1.947	1.835	1.927	1.777	1.780	1.809
TOTAL	12.000	12.000	12.000	12.000	12.000	12.000	12.000	12.000	12.000	12.000	12.000
T2 Al	2.697	2.779	2.714	2.718	2.761	2.702	2.737	2.688	2.691	2.672	2.658
M Cr	0.002	0.002	–	0	0	–	–	–	0.002	–	–
M Ti	0.010	0.018	0.015	0.011	0.013	0.017	0.017	0.007	0.004	0.011	0.009
M Fe	0.279	0.289	0.296	0.252	0.250	0.286	0.243	0.508	0.062	0.588	0.569
M Mn	0.005	0.006	0.007	0.002	0.001	0.001	0	0	0	0.010	0.010
M Mg	1.965	1.881	1.980	2.021	1.971	2.047	2.003	1.872	2.276	1.735	1.765
M Zn	0.016	0.004	–	–	–	–	–	–	–	–	–
TOTAL	2.281[4]	2.199	2.297	2.287	2.235	2.351	2.263	2.388	2.344	2.344	2.353
C1 Ba	0.007	0.003	0.016	0.006	0.004	0.016	0.002	0.007	–	0	0.004
C1 Ca	0	0	0.015	0.018	0.016	0.022	0.019	0.026	0.002	0.020	0.021
C1 Na	0.108	0.109	0.095	0.067	0.056	0.134	0.092	0.106	0.093	0.162	0.163
C1 K	0.920	0.948	0.922	0.914	0.906	0.897	0.930	0.901	0.911	0.850	0.895
TOTAL	1.035	1.060	1.048	1.005	0.982	1.068	1.042	1.041	1.006	1.032	1.083
TOTAL CATIONS	18.013	18.039	18.059	18.009	17.978	18.121	18.043	18.116	18.040	18.049	18.095

All Fe as FeO; 0 = <0.005% present; – = not analyzed.
* analyses done at UM; unmarked analyses done at UCLA.
[1] *Collected by Australian Bureau of Mineral Resources (BMR) geologists and provided by J. W. Sheraton. Complete BMR registered sample numbers are 77284249, 77284250, and 77284124.*
[2] *Provided by C. Maijer.*
[3] *Includes 0.04% NiO.*
[4] *Includes 0.005 Ni.*

high Fe_2O_3 content of 1% in the Brusilov sillimanite suggests that this osumilite may contain a significant proportion of Fe^{3+}.

Osumilites from granulite-facies rocks contain close to 18 cations per formula unit of 30 oxygens. However, occupancies of individual sites vary considerably from sample to sample and deviate from expected values. Most notably, the total number of cations expected to have sixfold coordination, *i.e.* Mg, Fe, Mn, Ti, and Cr, exceeds 2 per formula unit,

the number of sixfold sites available. If 60% of the iron is in sixfold coordination, the total cations in sixfold coordination would not significantly exceed 2 in osumilite more iron-rich than 2045C. However, the majority of Antarctic osumilites are more magnesian than 2045C, including osumilites from Spot Height 945 (Ellis *et al.*, 1980). The sum of K + Na + Ba + Ca exceeds one per formula unit (Tables 8, 9), a feature also reported by Ellis *et al.* (1980). This may reflect a systematic error in my (and Ellis')

Table 10. Minor element composition of two Antarctic osumilites (semiquantitative emission spectrographic analyses, F. Savino, analyst; provided by R. V. Gaines). In weight prercent of the element.

Element	Sample Number	
	2045C	2306B
Ba	0.1-0.5	0.5-1.0
Be	<0.01	<0.01
B	<0.01	<0.01
Ca	0.01-0.05	0.01-0.05
Co	<0.01	<0.01
Cu	<0.01	0.01-0.05
Ga	0.01-0.05	0.01-0.05
Pb	<0.01	<0.01
Ni	<0.01	<0.01
Ta	<0.01	<0.01
Sn	<0.01	<0.01
V	0.01-0.05	0.01-0.05
Zn	<0.01	<0.01
Mn	0.01-0.05	<0.01
Li	←------------not detected----------→	

Table 11. Mössbauer parameters for an Antarctic and a Nain osumilite

Sample Number	% Area	Isomer Shift[1] (mm/sec)	Quadrupole Splitting (mm/sec)	Half-Width (mm/sec)	χ^2
2045C[2]	61(2)	1.19(3)	2.33(3)	0.45(2)	
	34(2)	1.13(3)	1.70(3)	"	297[3]
	5(2)	0.42(4)	0.73(4)	"	
USNM #[4]	68	1.20	2.35	0.42	392[5]
143967	32	1.14	1.86	0.42	

[1]*Relative to metallic Fe.*
[2]*Estimated error of least significant figure given in parentheses. Source of data: W. A. Dollase. Concentration in sample was 0.9 mg Fe/cm^{2}. Half-widths of the 3 peaks constrained to be the same. Dips of the 3 peaks are 1.3%, 0.7%, and 0.1%.*
[3]*250 channels fitted*
[4]*Goldman and Rossman (1978). Two doublet fit.*
[5]*200 channels fitted.*

microprobe analyses. For example, in formulae calculated from wet chemical analyses and Goldman and Rossman's (1978) microprobe analysis of a Nain osumilite (U.S. National Museum #143967, Table 8), this total does not exceed unity. Another possibility, suggested by W. Schreyer (personal communication, 1981) is that Na may enter a sixfold site. Na in sixfold coordination is present in a new mineral (eifelite) related to osumilite (Abraham *et al.*, 1980).

In an attempt to rationalize the variations in site occupanices, I have plotted analyses recast to 30 oxygens with the assumption that the tetrahedral sites in the ring (T1) are fully occupied, *i.e.* that Si + $Al^{IV(1)}$ = 12. The remaining Al is assigned to the tetrahedra linking the rings, T2. Values of Al in T2 and of the sum of cations that could occupy the M position, Mg, Fe, Ti, Mn, Cr and Zn, plot close to a straight line such that Al in T2 + M = 5, the value in Schreyer's and Seifert's (1967) synthetic end-member (Fig. 7). Systematic differences exist between the microprobe data from UCLA and UM, but both sets of data obey the trend. However, a plot of Al in T1 *vs.* the sum of the cations in M displays considerable scatter (Fig. 8), and the differences among different sets of microprobe data are more pronounced. The data are consistent with a relation: Al in T1 + M = 4, the value in Schreyer and Seifert's (1967) synthetic end-member. The relations indicated by the trends in Figures 7 and 8 are consistent with an overall substitution $(Mg+Fe^{2+}+Mn)$ + Si = 2Al proposed by Berg and Wheeler (1976) and confirmed by Ellis *et al.* (1980) and Olesch and Seifert (1981). Undoubtedly, much of the scatter in Figure 8 is due to errors in the analyses. Other possible contributions to the scatter are substitutions such as Ba + Al = K + Si or Ti + 2Al = Mg + 2Si, which could also account for the average excess of Al over the theoretical values.

In the Antarctic samples, Al content is related to associated minerals; osumilites in sillimanite-bearing rocks are on the average more aluminous than osumilites in rocks lacking sillimanite. The present analyses confirm Ellis *et al.*'s (1980) data that the Antarctic osumilites are more aluminous than the Nain and Rogaland osumilites.

Osumilite compositions from volcanic rocks and low-pressure contact metamorphic rocks also plot close to the 2Al = Si + Mg curve in Figures 7 and 8.
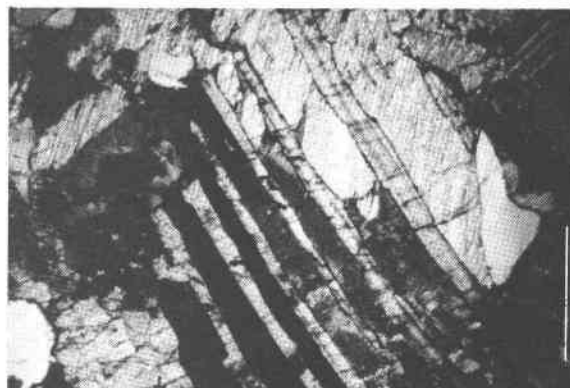
Fig. 6. Photomicrograph of sample 2345D from Mount Riiser-Larsen, Enderby Land, containing inverted pigeonite. Coarse lamellae in diagonal position, which extinguish simultaneously, are clinopyroxene. These lamellae cut grain boundaries of orthopyroxene, which are present in different optical orientations (light, dark gray, and black). White, featureless grains are quartz. Crossed nicols. Length of bar 1 mm.
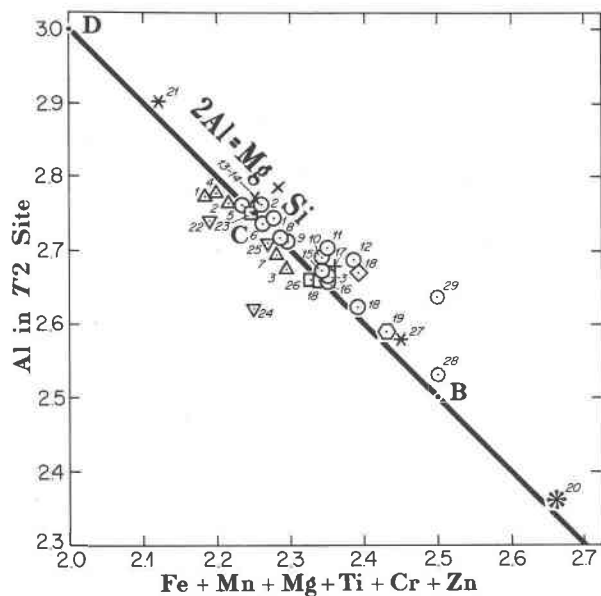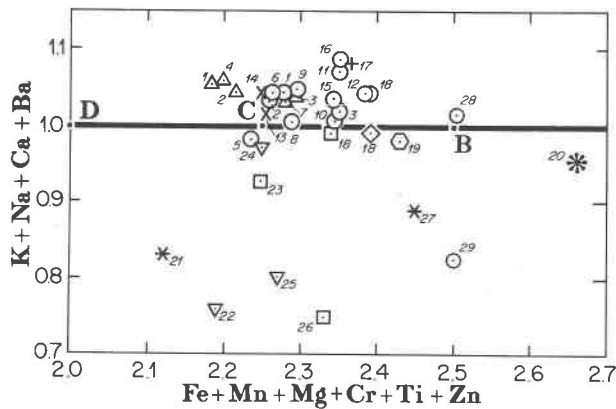
Fig. 7. Al in T2 site plotted versus all cations that could occupy M site in osumilite (per 30 oxygens). Line refers to ideal substitution Al in T2 + Al in T1 = Mg + Si. Type and source of analyses indicated by symbols. Electron microprobe: ○ = UCLA and △ = University of Melbourne analyses done as part of this study. + = Maijer et al. (1977); □ = Goldman and Rossman (1978); ✱ = Hentschel et al. (1977); ◐ = Berg and Wheeler, 1976; ✵ = Chinner and Dixon (1973); ▽ = Olsen and Bunch (1970; formulae recalculated by Berg and Wheeler, 1976); and X = Ellis et al. (1980); and ⊙ = Miyachi and Miyachi (1978). Wet chemical analysis: ◇ = Jarosewich et al. (1979). Synthetic: · − D = Schreyer and Seifert (1967), B, C = Olesch and Seifert (1981). Numbers refer to samples or localities. 1–14 are Enderby Land: 1 = 2045C, 2 = 2045E, 3 = 2045G, 4 = 2046D, 5 = 2083D, 6 = 4250, 7 = 2045J, 8 = 2083C, 9 = 2050A, 10 = 2306B, 11 = 4259, 12 = 4124, 13 and 14 = 76283358. 15–17 are Rogaland, Norway: 15 = OS-17, 16 = OS-99. 18–19 are Nain, Labrador, Canada: 18 = USNM #143967, 19 = 2-1817. 20 is Tieveragh, Ireland. 21 and 27 are Eifel District, Germany. 22 and 26 are Obsidian Cliffs, Oregon. 23–24 are Sakkabira, Japan. 28–29 are Iriki, Japan. 25 is Monte Arci, Italy.

In marked contrast to the osumilites from granulite-facies rocks, some volcanic samples are depleted in the large cations; sum of K + Na + Ca + Ba totals only 0.75 to 0.95 per 30 oxygens (Fig. 9). One explanation for this deficiency is substitution of Fe^{3+} for Fe^{2+} and concomitant loss in alkalis: K + Fe^{2+} = Fe^{3+} + □. Hentschel et al. (1980) proposed an analogous substitution in roedderite, a mineral structurally related to osumilite. In Obsidian Cliffs osumilite, Goldman and Rossman (1978) estimate an Fe^{3+} content of 0.23 per 30 oxygens (from Mössbauer data), which nearly balances the 0.25 deficit in K + Na in the C site.

In summary, natural and synthetic osumilites are remarkably stoichiometric in that cations (exclusive of K, Na, Ca, and Ba) total 17. Deviations from total stoichiometry (i.e. 18 cations per 30 oxygens) are observed only in some osumilites formed at low pressures in which significant amounts of ferric iron may be present. However, there appear to be more cations in six-fold coordination than there are octahedral sites, and fewer cations in four-fold coordination than there are tetrahedral sites.

Cell parameters

The most important compositional variations in osumilite are the substitution of Mg for Al (and concomitant substitution of Si for Al) and of Fe^{2+} for Mg. Both substitutions have an effect on the cell parameters (Table 12). By comparison of two iron-free synthetic osumilites, Olesch and Seifert (1981) show that both a and c increase as Mg increases and Al decreases. The effect of iron is less clear; a increases with increasing iron content, even if allowance is made for increasing total Fe^{2+} + Mg + Mn (Table 12); c apparently does not increase with increasing Fe^{2+}; it may decrease. Osumilite may thus behave like cordierite, in which a and b increase, but c decreases with increasing Fe + Mn (Selkregg and Bloss, 1980).

The significant increase in cell volume with decreasing Al content suggests that osumilite composition is a potential indicator of pressure. Osumilites formed at higher pressures would be expected to
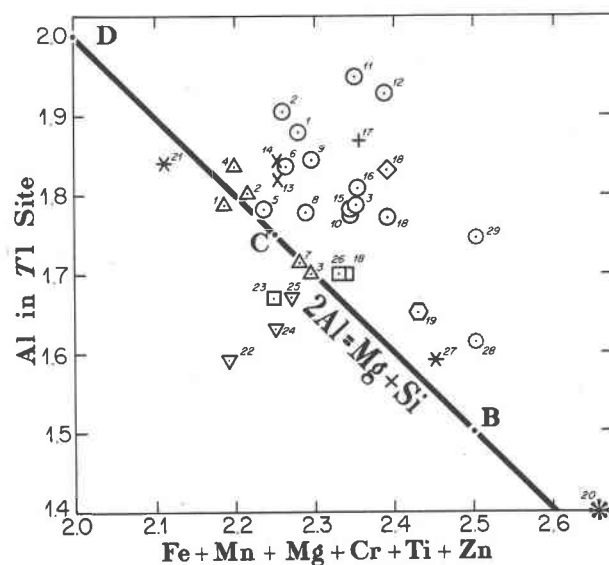
Fig. 8. Al in the T1 site plotted versus all cations that could occupy the M site in osumilite (per 30 oxygens). Explanation of symbols and numbers given in Figure 7.

Fig. 9. Total cations in the C1 site plotted *versus* all cations that could occupy the M site in osumilite (per 30 oxygens). Line is for ideal osumilite with full occupancy of the C1 site. Explanation of symbols and numbers given in Figure 7.

approach the end-member composition synthesized by Schreyer and Seifert (1967). However, the Reference Peak osumilite (2306B) formed at 7 kbar is similar in composition to Olesch and Seifert's (1981) osumilite of composition C synthesized at 0.5 to 1 kbar, suggesting that such compositional variation with pressure may be too small to be measured. Moreover, in a given area osumilite composition varies from one rock to another, suggesting that Al–Si–Mg ratios are influenced by compositional variables that would mask its variation with pressure. Maximum Al content in silica saturated systems can be attained only in osumilite associated with sillimanite and K-feldspar.

Petrogenetic grids for osumilite in the granulite facies

General considerations

On the basis of experimental studies of synthetic and natural osumilites, Schreyer and Seifert (1967) suggested that osumilite may be a metastable phase. Hensen (1977a) synthesized osumilite, apparently stably, in a model pelitic composition at 1000° and 1100° C, 3.6 to 6.3 kbar, and at low water and oxygen fugacities. Hensen (1977a) proposed a petrogenetic grid (based on Schreinemakers constructions) involving the phases osumilite, K-feldspar, quartz, cordierite, sapphirine, spinel, and orthopyroxene in the system $K_2O–MgO–FeO–Al_2O_3–SiO_2$. Ellis *et al.* (1980) proposed more extensive grids in this system (and in the iron-free subsystem), which, unlike Hensen's (1977a) grid, allow for the stable coexistence of osumilite with garnet or sillimanite.

Olesch and Seifert (1981) report that osumilite of composition $KMg_2(Al_{2.75}Mg_{0.25})(Al_{1.75}Si_{10.25})O_{30}$ is table at water pressures up to 0.8 kbar at temperatures of 780° to 790° C (where $P_{H_2O} = P_{Total}$). Towards lower temperatures, Olesch and Seifert (1981) report that this osumilite breaks down to cordierite + K-feldspar + phlogopite + quartz, and towards higher temperatures to the same phases (less quartz) + liquid. Olesch (1981) calculated that Olesch's and Seifert's (1981) stability field for osumilite should expand to higher pressures and to higher and lower temperatures where P_{H_2O} is less than P_{Total}.

Four grids based on Schreinemakers constructions are discussed here: one in the subsystems $MgO±FeO–Al_2O_3–SiO_2$, two in anhydrous systems, $K_2O–MgO–Al_2O_3–SiO_2$ (abbreviated KMAS) and $K_2O–FeO–MgO–Al_2O_3–SiO_2$ (KFMAS) and the fourth in the iron-free hydrous system (KMASH). I will consider only parageneses with excess quartz. The idealized K_2O-bearing systems are fair approximations of osumilite-bearing metamorphic rocks (Table 2); the most serious omission is Na_2O. The phases in this system (and abbreviations) are quartz (Qtz), osumilite (Osu), sillimanite (Sil), sapphirine (Sap), garnet (Gar), cordierite (Crd), biotite (Bio, Phl for phlogopite), orthopyroxene (Opx, En for enstatite), and K-feldspar (Kf). These minerals consist largely of the components under consideration and all of the minerals except biotite are anhydrous. The water content of cordierite is expected to be low at the high temperatures and low P_{H_2O} of the pyroxene–granulite facies (Newton and Wood,

Table 12. Cell parameters of osumilites of known composition

Sample Number or Reference	$\underline{M}$[1]	$\underline{x}$[2]	$\frac{a}{\overset{\circ}{A}}$	$\frac{c}{\overset{\circ}{A}}$	$\frac{V}{\overset{\circ}{A}^3}$
Synthetic[3]	2.00	0	10.078(2)	14.317(6)	1259.4(5)
Synthetic[4]	2.25	0	10.087(2)	14.333(5)	1263.0[8]
2306B[5]	2.34	0.027	10.0912(5)	14.336(1)	1264.3(1)
2045C[5]	2.28	0.17	10.1037(8)	14.329(2)	1266.8(2)
Rogaland[6]	2.36	0.25	10.114(1)	14.334(2)	1269.8[8]
Nain, #2-1817[7]	2.43	0.23	10.117(2)	14.255(6)	1263.6[8]

Standard errors of least significant figure in parentheses.
[1]*Number of cations per 30 oxygens of Fe + Mn + Mg + Ti + Cr + Zn.*
[2]*Molecular ratio (Fe + Mn)/(Fe + Mn + Mg).*
[3]*Schreyer and Seifert (1967).*
[4]*Olesch and Seifert (1981).*
[5]*This study. Compositions from UCLA analyses.*
[6]*Maijer et al. (1977)*
[7]*Berg and Wheeler (1976). N.B. Osumilite #2-1817 is much more magnesian than analyzed osumilite in sample 143967, Table 8, also from Nain.*
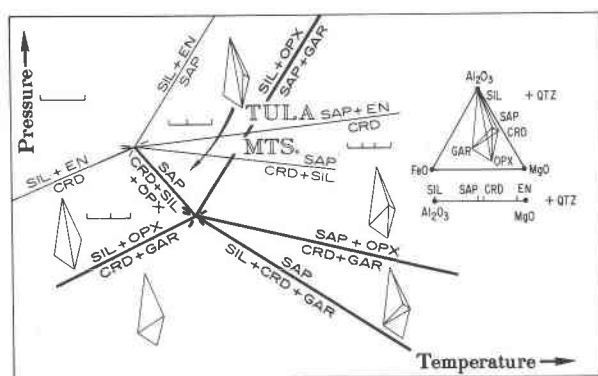[8]*Calculated from published values of a and c.*

Fig. 10. Pressure–temperature diagram based on Schreinemakers construction of phase relations in the systems MgO–Al$_2$O$_3$–SiO$_2$ (light lines) and FeO–MgO–Al$_2$O$_3$–SiO$_2$ (heavy lines). Compositional space for each system is shown on the right. Quartz is present in all assemblages. P–T conditions of metamorphism of rocks exposed in the Tula Mountains are indicated. Derivation of diagram is explained in text. Abbreviations: Crd = cordierite, Qtz = quartz, Sil = sillimanite, Sap = sapphirine, Opx = Orthopyroxene, En = enstatite, Gar = garnet. Curves for reactions Crd + Sil = Sap and Crd = Sap + En are shown divergent for illustration; in actuality, these curves are most likely almost coincident (Newton et al., 1974).

1979); Goldman et al. (1977) report that Nain cordierites are essentially anhydrous. Significant impurities in the minerals are Na$_2$O in K-feldspar and F, Ti and Ba in biotite. Reactions involving spinel have not been considered. Spinel commonly contains significant Cr, Zn, or Fe^{3+}, and its presence in quartz-bearing Enderby Land rocks, and possibly elsewhere as well, may be controlled as much by these components as by pressure and temperature conditions.

In writing reactions for the iron-free systems (KMAS, KMASH), the mineral compositions will be assumed to be constant. Compositions are cordierite (Mg$_2$Al$_4$Si$_5$O$_{18}$), osumilite KMg$_{2.33}$ Al$_{4.46}$Si$_{10.21}$O$_{30}$) (simplified from 2306B, Table 9); sapphirine (Mg$_{3.7}$Al$_{8.6}$Si$_{1.7}$O$_{20}$, average from Grew, 1980a) and enstatite (Mg$_{1.9}$Si$_{1.9}$Al$_{0.2}$O$_6$, or En$_{0.9}$ MgTs$_{0.1}$ in terminology of Danckwerth and Newton, 1978, from the hypothetical iron-free orthopyroxene coexisting with sapphirine, Grew, 1980a; this composition is less aluminous than the orthopyroxene in 2306B; which is (Mg,Fe)$_{1.83}$ Al$_{0.35}$Si$_{1.82}$O$_6$, see Table 3). Phlogopite is taken to have the theoretical composition (KMg$_3$Al Si$_3$O$_{10}$(OH)$_2$, which is close to the compositions of biotite associated with osumilite (Table 4), sillimanite (Grew, 1980a), or garnet (Ellis et al., 1980) in Enderby Land. There is no evidence for a shift to

more aluminous compositions which Olesch and Seifert (1981) reported in phlogopite formed from the breakdown of synthetic osumilite.

I have assumed that changes in the P–T slopes of the reaction curves due to changes in mineral compositions (e.g. Al in orthopyroxene) are not sufficient to affect the topology of the grids in the range of P–T under consideration.

Subsystems MgO–Al$_2$O$_3$–SiO$_2$ and FeO–MgO–Al$_2$O$_3$–SiO$_2$

Hensen and Essene (1971; also Hensen, 1972) and Newton et al. (1974) presented P–T diagrams for the MgO–Al$_2$O$_3$–SiO$_2$ system. The diagrams differ in the slopes of the cordierite breakdown reactions. Newton et al. (1974) calculated a positive slope for the reaction Crd = En + Sil + Qtz from thermochemical data, whereas Hensen and Essene (1971) report a negative slope on the basis of experimental data. Moreover, Newton et al. (1974) calculated a flat slope for the reaction Crd = Sap + Qtz + En, which is consistent with Newton's (1972) experimental data. Figure 10 is based on Newton et al.'s (1974, Fig. 2) diagram.

The slopes of the reactions around the invariant point Qtz–Gar–Sap–Crd–Opx–Sil (spinel-absent invariant point of Hensen, 1971, Fig. 1) in the FeO–MgO–Al$_2$O$_3$–SiO$_2$ system are taken from Hensen except for the slope of the reaction Crd + Gar = Sil + Opx + Qtz. Hensen and Green (1971, 1972, 1973; also Hensen, 1977b) report a negative slope for this reaction on the basis of experimental data. However, Marakushev and Kudryavtsev (1966) noted that the sizable negative $\Delta \overline{V}$ of forming Sil + Opx + Qtz from Gar + Crd implies that $\Delta \overline{S}$ is also negative, and the slope of the reaction, positive. In any case, the slope of this reaction should not differ markedly from that of Crd = Sil + En + Qtz in the iron-free system. Consequently, in Figure 10 the slope of the cordierite-breakdown reaction in the FeO-bearing system is shown as positive.

Assemblages with sapphirine–garnet–quartz, sillimanite–orthopyroxene–quartz, or with all five of these minerals (e.g. Mount Torckler) are found in the Tula Mountains, while primary cordierite appears to be absent (Grew, 1979; Sheraton et al., 1980; Ellis et al., 1980). The P–T conditions of metamorphism in the Tula Mountains are thus constrained to be close to those of the cordierite-breakdown curves and the curve for Sil + Opx = Sap + Gar + Qtz (Fig. 10).
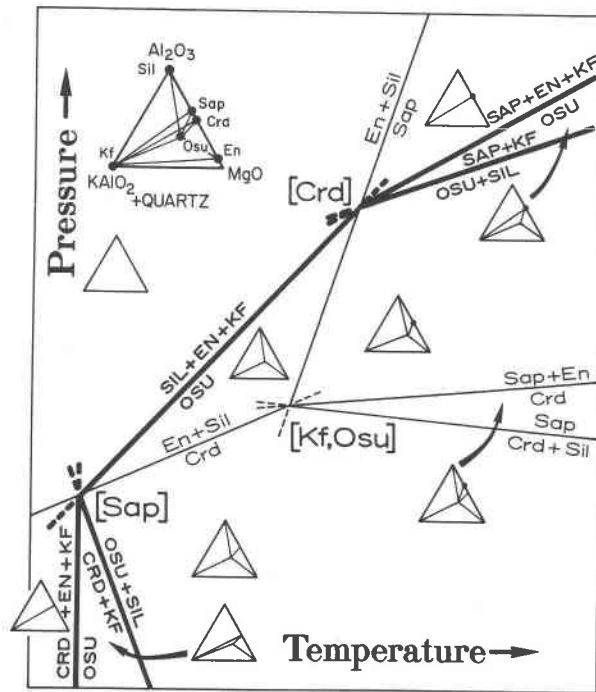
Fig. 11. Pressure–temperature diagram based on Schreinemakers construction of phase relations in the MgO–Al$_2$O$_3$–SiO$_2$ system (light lines) and in the K$_2$O–MgO–Al$_2$O$_3$–SiO$_2$ system (heavy lines), quartz present in all assemblages. Abbreviations for minerals given in text and in caption of Fig. 10, except Kf = K-feldspar, Osu = osumilite. Invariant points are identified by the absent phases in brackets. Derivation of diagram explained in text. To distinguish from cordierite, sapphirine is indicated by solid circles on the composition triangles.

System K$_2$O–MgO–Al$_2$O$_3$–SiO$_2$ (KMAS)

Slopes in the KMAS system are estimated qualitatively from calculated $\Delta\overline{V}$'s of reactions and constrained by the assumption that reactions forming osumilite or expanding its stability field involve an increase in entropy. The molar entropy of osumilite is probably higher than combinations of the other minerals reacting with it, $i.e.$ cordierite, quartz, K-feldspar, sapphirine, orthopyroxene, or sillimanite. Al and Si in the ring tetrahedra of osumilite are disordered (Brown and Gibbs, 1969), while in cordierite at temperatures near 1000° C or less, Al and Si are probably ordered (Putnis, 1980). Moreover, osumilite is a relatively low density mineral by comparison with sillimanite and the other ferromagnesian silicates except cordierite and, consequently, a higher molar entropy is expected (Fyfe $et\ al.$, 1958).

Molar volumes (room temperature values) of quartz (alpha-quartz, 22.688 cm^3), phlogopite

(149.91 cm^3), K-feldspar (high sanidine, 109.050 cm^3), and cordierite (233.220 cm^3) are from Robie $et\ al.$ (1978). Molar volumes for the other minerals are aluminous enstatite (62.230 cm^3, Danckwerth and Newton, 1978), osumilite (380.78 cm^3, from 2306B, Table 12), sillimanite (49.855 cm^3, Grew, 1980b), and sapphirine (197.945 cm^3, Newton $et\ al.$, 1974).

Volume changes of some osumilite-forming reactions or reactions expanding the stability field of osumilite ($\Delta\overline{V}$ for right hand side) are:

$$3Kf + En + 2.545Crd + 7.005Qtz = 3Osu$$
$$\Delta\overline{V} = +0.49 \text{ cm}^3 \tag{1}$$

$$2.33Crd + 2Kf + 3.97Qtz = 2Osu + 1.2Sil$$
$$\Delta\overline{V} = -30.19 \text{ cm}^3 \tag{2}$$

$$2.33Sap + 3.7Kf + 26.334Qtz = 3.7Osu + 3.618Sil$$
$$\Delta\overline{V} = +119.3 \text{ cm}^3 \tag{3}$$

$$Sap + 2.554Kf + 1.1847En + 14.4636Qtz$$
$$= 2.554Osu$$
$$\Delta\overline{V} = 94.2 \text{ cm}^3 \tag{4}$$

$$2.33En + 3.054Sil + 1.9Kf + 6.218Qtz = 1.9Osu$$
$$\Delta\overline{V} = 77.96 \text{ cm}^3 \tag{5}$$
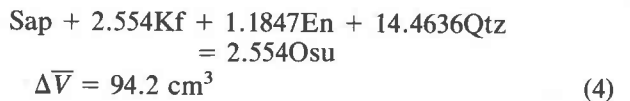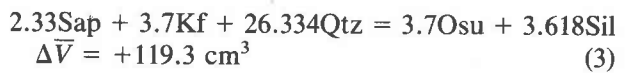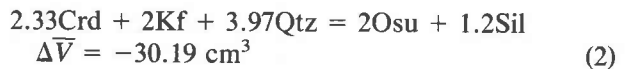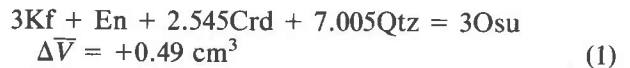
These calculations demonstrate that formation of osumilite from assemblages of sapphirine, sillimanite, and orthopyroxene is favored by decreasing pressure, but in cordierite-bearing assemblages, reactions forming osumilite or extending its stability field are either insensitive to pressure or are favored by a pressure increase.

An important step in constructing the grid shown in Figure 11 is fixing the position of the reactions involving osumilite relative to the invariant point [Kf, Osu] in the MgO–Al$_2$O$_3$–SiO$_2$ system (absent phases in brackets, a convention used in identifying invariant points, Zen, 1966). Osumilite could form by the reaction of cordierite, K-feldspar, enstatite and quartz (equation (1) from Ellis $et\ al.$, 1980, eq. 16). According to Ellis $et\ al.$, (1980, Fig. 19), reaction 16 intersects the reaction Crd = Sap + Qtz + En at a temperature above the invariant point Sap–Qtz–Crd–Sil–En or [Kf, Osu], $i.e.$ at an invariant point [Sil], which is equivalent to Hensen's (1977a) invariant point [Phl]. Hensen's (1977a) and Ellis $et\ al.$'s (1980) interpretation imply that osumilite is not stable at the P–T conditions of the invariant point [Kf, Osu]. However, Olesch and Seifert (1981) report a stability field for osumilite at 780–790° C and $P_{H_2O} = P_{Total} = 0.5$ kbar. At $P_{H_2O} \ll P_{Total}$, osumilite should be stable to total pressures above 0.5 kbar and at temperatures above and

below 780–790° C (Olesch and Seifert, 1981; Olesch, 1981). Given Newton *et al.*'s (1974) *P–T* estimate of 825° C and 6.7 kbar for the invariant point [Kf, Osu] in natural systems, it appears likely that osumilite will be stable at the *P–T* conditions of [Kf, Osu] if $P_{H_2O} \ll P_{Total}$. Consequently, the curve for reaction (1) should intersect the reaction curve Crd = Sil + En + Qtz at a temperature less than that of the invariant point [Kf, Osu]; this intersection is the stable invariant point [Sap] in Figure 11. Like the diagram proposed by Ellis *et al.* (1980), the diagram presented here predicts a stability field for the Sil–En–Osu assemblage in the iron-free system. However, in Figure 11 this assemblage is stable at temperatures below, as well as above, that of the [Kf, Osu] invariant point. Consequently, the highly magnesian Osu–Opx–Sil assemblage from Reference Peak (Sample 2306B), in which the orthopyroxene contains 91 to 93 mole % enstatite (Table 3), does not imply temperatures in excess of that of [Kf, Osu] as Ellis *et al.*'s (1980) diagram requires.

The presence of a stable [Crd] invariant point in Figure 11 depends on whether the curve for reaction (5) intersects the En + Sil = Sap + Qtz reaction. I predict that these two reactions do intersect, though in an inverse manner from that illustrated by Ellis *et al.* (1980). The slope of (5) in *P–T* space is probably less than that for En + Sil = Sap + Qtz, as the $\Delta \overline{V}$ (for a given number of oxygens) of (5) is considerably greater (78 cm^3 for 57 oxygens *vs.* 20 cm^3 for 61 oxygens).

The positions and slopes of other univariant curves in KMAS are constrained by the rules for constructing Schreinemakers diagrams (Zen, 1966) and by the negative $\Delta \overline{V}$ calculated for equation (2), and the positive $\Delta \overline{V}$'s for equations (3) and (4).

System K_2O–FeO–MgO–Al_2O_3–SiO_2 (KFMAS)

A logical approach to building a grid for the iron-bearing system (KFMAS) is to assume that two of the univariant reactions in this system correspond to the invariant points [Sap] and [Crd] in KMAS, and curves for these two reactions terminate in the corresponding invariant points. An important constraint on the slopes of these curves is the fractionation of iron relative to magnesium, which increases as follows: Osu < Crd < Bio < Sap < Opx < Gar (Fig. 5; Berg and Wheeler, 1976; Ellis *et al.*, 1980). As a result of this fractionation, the reactions stabilizing the KFMAS assemblage Osu + Sil + Opx relative to Kf + Crd and Sap + Kf must lie in the stability field of the KMAS assemblage Osu +

Sil (Fig. 12). This decreases the stability range of osumilite in iron-bearing rocks relative to that in iron-free rocks (Fig. 12).

By analogy with the iron-free system, reactions involving osumilite and not cordierite have positive slopes and the osumilite-bearing assemblages will lie on the low-pressure, high-temperature side of the curve. This assumption also dictates the position of the invariant point [Crd] in KFMAS relative to [Crd] in KMAS (Fig. 12). The remainder of the *P–T* diagram is then determined. The reaction Opx + Sil + Kf + Qtz = Osu + Gar must lie within the stability field of Opx + Sil, and thus on the low temperature side of the reaction Opx + Sil = Sap + Gar + Qtz (Fig. 12). The reaction stabilizing the Osu + Gar assemblage intersects the reaction Crd + Kf = Osu + Sil + Opx at the invariant point [Sap], which is located at a lower temperature than [Kf, Osu]. Thus the stability field of Qtz–Opx–Sil overlaps with that of Qtz–Osu–Gar. Ellis *et al.*'s (1980, Fig. 20) grid does not allow for osumilite to be stable within the stability field of Qtz–Sil–Opx in the iron-bearing system.

A critical assemblage to determine which diagram is consistent with the mineralogy of the Tula Mountains is Qtz–Gar–Osu–Sil–Opx. Ellis *et al.* (1980, Table 1) report this assemblage from Spot Height 945, and two samples I collected each at a different locality on Mount Riiser-Larsen also appear to contain his assemblage. One rock consists largely of quartz and osumilite (in part altered), minor sillimanite, and traces of orthopyroxene and garnet. Garnet and osumilite are in direct physical contact near quartz, but sillimanite and orthopyroxene do not happen to touch. In a thin section of the second rock, quartz, garnet, sillimanite, orthopyroxene, and symplectite (from osumilite breakdown) are in close proximity. Orthopyroxene touches sillimanite and garnet touches symplectite; unaltered osumilite is present nearby in the section. Elsewhere in this section, sapphirine is in direct contact with quartz. Whether the assemblage sapphirine–garnet–quartz (theoretically incompatible with Qtz–Sil–Opx) is also stable in this rock or in other rocks from Mount Riiser-Larsen is not clear from the textures. Pressure–temperature conditions may have been close to those of univariant reaction Sil + Opx = Sap + Gar + Qtz (Grew, 1980c). In any case, the assemblages from both Spot Height 945 and Mount Riiser-Larsen imply that the stability fields of the Osu–Gar–Qtz assemblage and the Qtz–Sil–Opx assemblage overlap, as indicated in Figure 12.
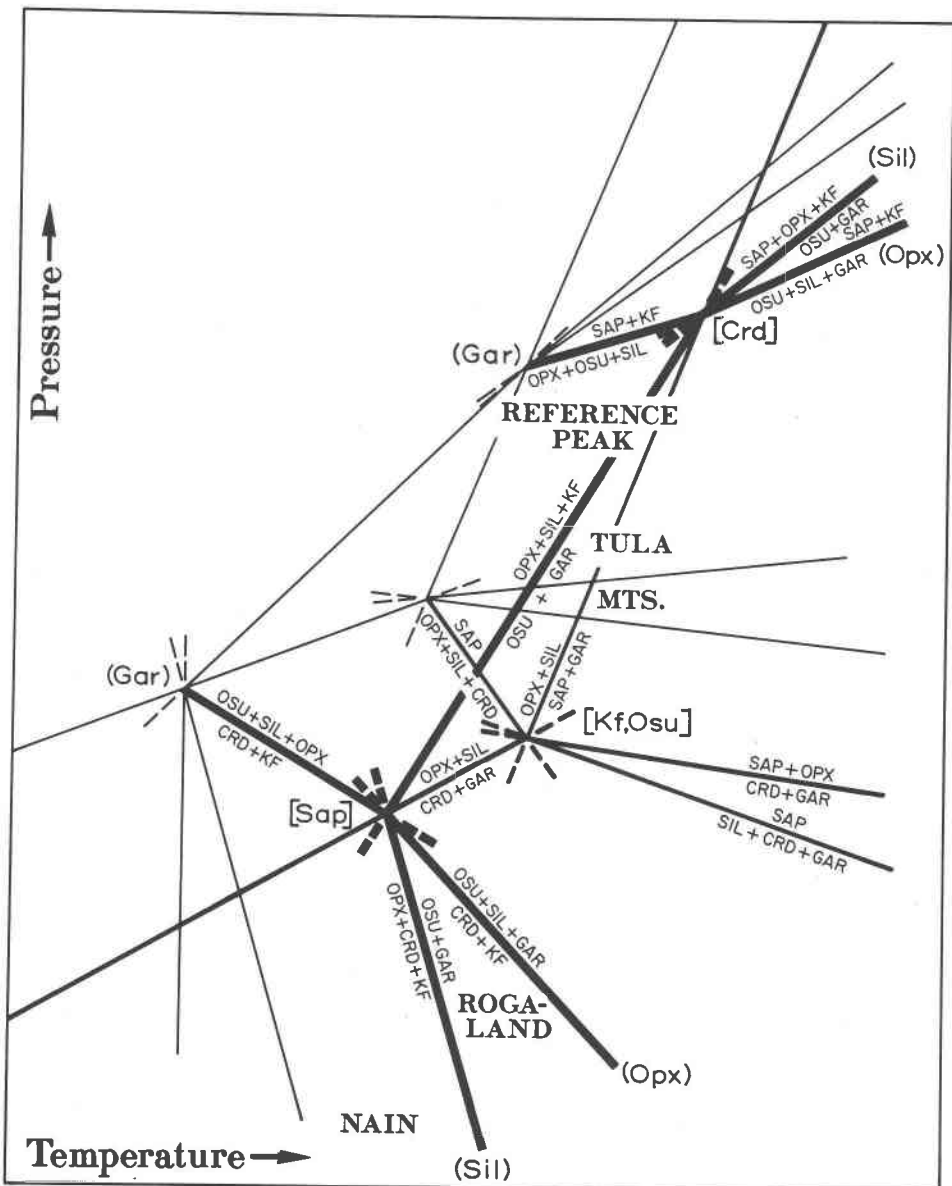
Fig. 12. Pressure–temperature diagram based on Schreinemakers construction of phase relations in the iron-free system from Figure 11 (lightest line), FeO–MgO–Al$_2$O$_3$–SiO$_2$ system in intermediate lines (marked by reactions), and K$_2$O–FeO–MgO–Al$_2$O$_3$–SiO$_2$ system in heaviest lines; quartz present in all assemblages. Relative P–T conditions are indicated for granulite-facies metamorphism at Reference Peak and in the Tula Mountains, Enderby Land, and for the contact aureoles of anorthositic complexes at Rogaland, Norway, and Nain, Labrador, Canada. Explanation of abbreviations is given in Figures 10 and 11 and derivation of diagram is explained in text.

The construction of the P–T grid in Figure 12 requires that the slopes of two of three reactions issuing from [Sap] and involvng both osumilite and cordierite be negative. The slope of the sillimanite-absent reaction (abbreviated as (Sil), see Zen, 1966) is not constrained. Ellis *et al.* (1980) suggest a positive slope for (Sil). To test their suggestion, I have estimated the volume change for (Sil). Iron/magnesium ratios of the reacting minerals (except

orthopyroxene) are taken from the compositions measured in samples 2083C and 2083D. The Fe/Mg ratio of orthopyroxene, a mineral not present in these samples, is estimated from Figure 5. The ratios of (Fe+Mg) to Al and Si are the same as those used for calculating $\Delta \overline{V}$ in the iron-free system (see above). Molar volumes of garnet (114.19 cm^3) and orthopyroxene (63.21 cm^3) were calculated from the volumes of pyrope, almandine, and orthoferrosilite

tabulated in Robie *et al.* (1967) and of aluminous enstatite (see above) and assuming ideal solid solution. The molar volume of cordierite is estimated to be 233 cm^3, for cordierite with 10% iron end member has a somewhat smaller molar volume than the magnesium end member (Newton and Wood, 1980). The molar volume of osumilite with 11% iron end member is estimated to be 381.14 cm^3 from data on osumilites in samples 2306B and 2045C (Table 12). The reaction (Sil) can be written as follows (atomic Fe/(Fe+Mg) ratio in parentheses):

$$0.58Opx\ (0.29) + 0.95Crd\ (0.13) + Kf + 2.04Qtz$$
$$= Osu\ (0.11) + 0.22Gar\ (0.46)$$
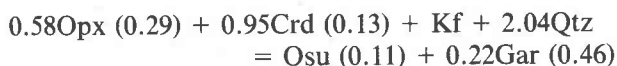
for which $\Delta \overline{V} = -6$ cm^3. Assuming that the $\Delta \overline{S}$ of this reaction is positive, its slope is negative though steep. Slopes of the other two reactions issuing from [Sap] and involving both cordierite and osumilite are negative and less steep than the slope of (Sil) (Figure 12). Consequently, the osumilite–garnet and osumilite–sillimanite assemblages may not be stable at low pressures at geologically accessible temperatures. They may require moderate pressures (possibly in excess of 5 kbar, see below) to appear at reasonable metamorphic temperatures.

At pressures above those of the metamorphism in the Tula Mountains, the grid predicts that osumilite-bearing assemblages will be replaced by assemblages such as Qtz–Kf–Sap or Qtz–Kf–Sil–Opx. These assemblages are relatively rare in the Tula Mountains (examples are described by Ellis *et al.,* 1980; Ravich and Kamenev, 1975, p. 147). The Qtz–Sap–Kf assemblage has been reported from other areas, such as Wilson Lake area, Labrador (Bourne, 1978). Unaltered osumilite is not known to occur in the Wilson Lake area, but osumilite breakdown products (Crd–Kf–Opx±Sil) have been found (R. K. Herd, personal communication, 1981). Strictly speaking, Qtz–Kf–Sap or Qtz–Kf–Sil–Opx are incompatible with osumilite only if the K-feldspar contains little Na, which is rarely the case under these high grade conditions. The presence of Qtz–Sap–Kf instead of osumilite may be due to high Na/K ratio of the host rock and it does not imply that pressures were too high for osumilite to be stable unless the K-feldspar is close to $KAlSi_3O_8$ in composition.

Mineral assemblages can be used to locate the position in *P–T* space of the Tula Mountain assemblages relative to the osumilite-bearing assemblages at Reference Peak, in the contact aureoles in Rogaland, Norway (Maijer *et al.,* 1977, 1981), Nain,

Labrador (Berg and Wheeler, 1976), and Dzhugdzhur, Soviet Far East (Bogdanova *et al.,* 1980). The presence of Sil–Opx–Osu rocks with high Mg/Fe ratios (*e.g.* 2306B, Table 2) and absence of Osu–Gar asemblages at Reference Peak are consistent with an increase in metamorphic pressure southward from the Tula to the Scott Mountains (Fig. 1), a trend suggested by Sheraton *et al.* (1980) and Grew (1980a). Moreover, the high Mg/Fe ratios in the Reference Peak rocks suggest metamorphic conditions approached those of reactions forming osumilite from Kf + En + Sil in the iron-free system.

In the Rogaland aureole, osumilite occurs with quartz, perthitic K-feldspar, plagioclase, biotite, magnetite, spinel, garnet, and orthopyroxene (Maijer *et al.,* 1977, 1981). Kars *et al.* (1980) report the same assemblage with cordierite. Kars *et al.'*s (1980) Qtz–Kf–Gar–Opx–Crd–Osu assemblage is theoretically univariant, *i.e.* (Sil) in Figure 12. However, as the perthitic K-feldspar is sodic, the assemblage becomes divariant, and the *P–T* conditions of its formation would lie within the stability field of garnet–osumilite. Moreover, Maijer *et al.* (1981) and Kars *et al.* (1980) suggest that garnet formation is not coeval with osumilite formation, but occurred during an earlier metamorphic event, and again during a later event, which straddled the osumilite-forming event. In thin sections I have examined of Rogaland osumilite-bearing rocks provided by Dr. Maijer, garnet grains are typically embayed and are commonly overgrown by plagioclase. A few garnet grains enclosed in osumilite are in direct contact with osumilite, but isolated from quartz. Thus there is no textural evidence for an equilibrium Qtz–Gar–Osu assemblage. Nonetheless, it is still possible that the quartz–garnet–osumilite assemblage was stable at some time during the metamorphism at Rogaland. It would have crystallized at pressures below and temperatures above the invariant point [Sap] and close to the univariant reaction (Sil) (Figure 12). Moreover, pressures and temperatures may have been close to those of [Sap]. C. Maijer and his collegues have also found osumilite in contact with sillimanite, although this sillimanite, like the garnet, may be a relic of the early metamorphic event (C. Maijer, personal communication, 1981). In contrast, osumilite at Nain and Dzhugdzhur has not been reported with garnet or sillimanite, and consequently metamorphism in these areas occurred at *P–T* conditions below (Sil) (Figure 12).

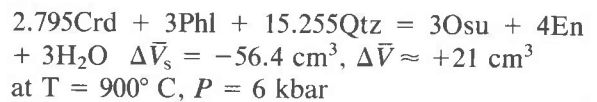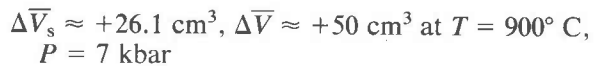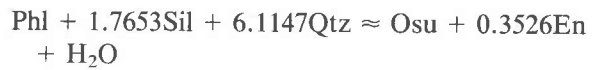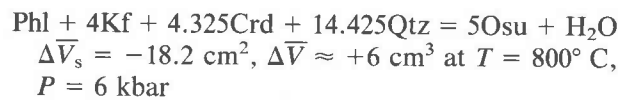Temperatures of 800° to 1000° C are inferred for

the osumilite-bearing rocks in the Rogaland aureole (Kars *et al.*, 1980). Pressures during the second garnet-forming event in the Rogaland aureole are estimated to have been about 6.3 kbar (Henry, 1974), pressures during the osumilite-forming event may have been the same. Berg and Wheeler (1976) estimate that the Nain osumilite crystallized at temperatures "near the high end" of 700–900° C and at a total pressure of 5.2 kbar. More recently Bohlen and Boettcher (1981) calculated pressures of 3.2 ± 0.2 kbar using orthopyroxene geobarometry and Perkins and Newton (1981), 3.3 kbar using an orthopyroxene–plagioclase geobarometer for the contact aureoles of the Nain complex. Berg (1977) estimated that the pressures in these contact aureoles ranged from 3.7 to 6.6 kbar. These pressure estimates, notably the more recent estimates for the Nain aureoles, are consistent with the relative positions for Nain, Rogaland, and the Tula Mountains indicated in Figure 12.

System K_2O–MgO–Al_2O_3–SiO_2–H_2O (KMASH)

Hensen (1977a) suggested that osumilite most probably formed from phlogopite-bearing assemblages under reduced water pressure. Olesch and Seifert (1981) reported the formation of osumilite by the univariant reaction Osu = Phl + Crd + Kf + Qtz + H_2O under water saturated conditions and Olesch (1981) calculated the effect of variable water partial pressure on the formation of osumilite by this reaction. In this section, I will consider some possible osumilite-forming reactions involving phlogopite in the iron-free system K_2O–MgO–Al_2O_3–SiO_2–H_2O (KMASH).

The petrogenetic grid for KMASH (Fig. 13) is less well constrained than that for KMAS. The position of the reactions involving phlogopite relative to those in the anhydrous system may vary as a function of fluid composition. The grid in Figure 13 has been constructed so that the invariant point [Sap, En] is stable and so that the assemblage Osu–Sil–En–Qtz has a stability field. The univariant reaction investigated experimentally by Olesch and Seifert (1981) is one of the reactions around [Sap, En]. Several reactions in the KMAS system illustrated in Figure 11 are metastable relative to phlogopite-bearing assemblages and are shown in dashed lines in Figure 13. As a result, the stability range of osumilite has shrunk in the hydrous system.

Slopes of reactions involving phlogopite and osumilite are positive. For example ($\Delta \overline{V}_s$ is volume change for solid phases only; volume of water estimated from Burnham *et al.*, 1969).

$$Phl + 4Kf + 4.325Crd + 14.425Qtz = 5Osu + H_2O$$
$$\Delta \overline{V}_s = -18.2 \text{ cm}^2, \ \Delta \overline{V} \approx +6 \text{ cm}^3 \text{ at } T = 800° \text{ C},$$
$$P = 6 \text{ kbar}$$

$$Phl + 1.7653Sil + 6.1147Qtz \approx Osu + 0.3526En + H_2O$$

$$\Delta \overline{V}_s \approx +26.1 \text{ cm}^3, \ \Delta \overline{V} \approx +50 \text{ cm}^3 \text{ at } T = 900° \text{ C},$$
$$P = 7 \text{ kbar}$$

$$2.795Crd + 3Phl + 15.255Qtz = 3Osu + 4En + 3H_2O \quad \Delta \overline{V}_s = -56.4 \text{ cm}^3, \ \Delta \overline{V} \approx +21 \text{ cm}^3$$
$$\text{at } T = 900° \text{ C}, P = 6 \text{ kbar}$$

No attempt will be made here to construct a grid for the hydrous iron-bearing system. In any case, biotite may not be involved in reactions at temperatures above the first appearance of osumilite. In some osumilite-bearing rocks, biotite may be stabilized by components not considered in constructing the grids, Ba, F, and Ti, which are important constituents of some Enderby Land biotites associates with osumilite (Table 4). In other osumilite-bearing rocks, biotite appears to be secondary (Ellis *et al.*, 1980; Berg and Wheeler, 1976). Of interest is the relative scarcity of biotite in the breakdown products of osumilite. In Enderby Land, these generally are cordierite, K-feldspar, quartz, orthopyroxene, and possibly, in rare cases, sillimanite (sample 4124, Table 1). Berg and Wheeler (1976) describe symplectites in the Nain rock similar to those in the Antarctic rocks, and biotite is rare in these symplectites. Maijer *et al.* (1981) report no secondary biotite in symplectites from Rogaland (also Schreyer and Seifert, 1967, Fig. 6); the Rogaland symplectites are also similar to the Antarctic ones (Maijer *et al.*, 1981).

Chemical constraints on osumilite stability

Berg and Wheeler (1976) and Ellis *et al.* (1980) note that at pressures and temperatures of the Nain aureole (*i.e.* $P \approx 5$ kbar, $T = 700–900°$ C), osumilite is excluded from Fe^{2+}-rich compositions by the join cordierite–orthopyroxene. At higher pressures and temperatures, the cordierite–orthopyroxene join is broken and osumilite becomes stable with garnet, thereby extending the stability field of osumilite to more Fe^{2+}-rich compositions (Ellis *et al.*, 1980).

However, available data suggest that osumilite associated with garnet is not always more iron rich than osumilite associated with cordierite and orthopyroxene, nor is there any evidence that rocks containing osumilite and garnet are always more iron rich than rocks containing osumilite, cordierite, and orthopyroxene. The atomic Fe/(Fe+Mg)
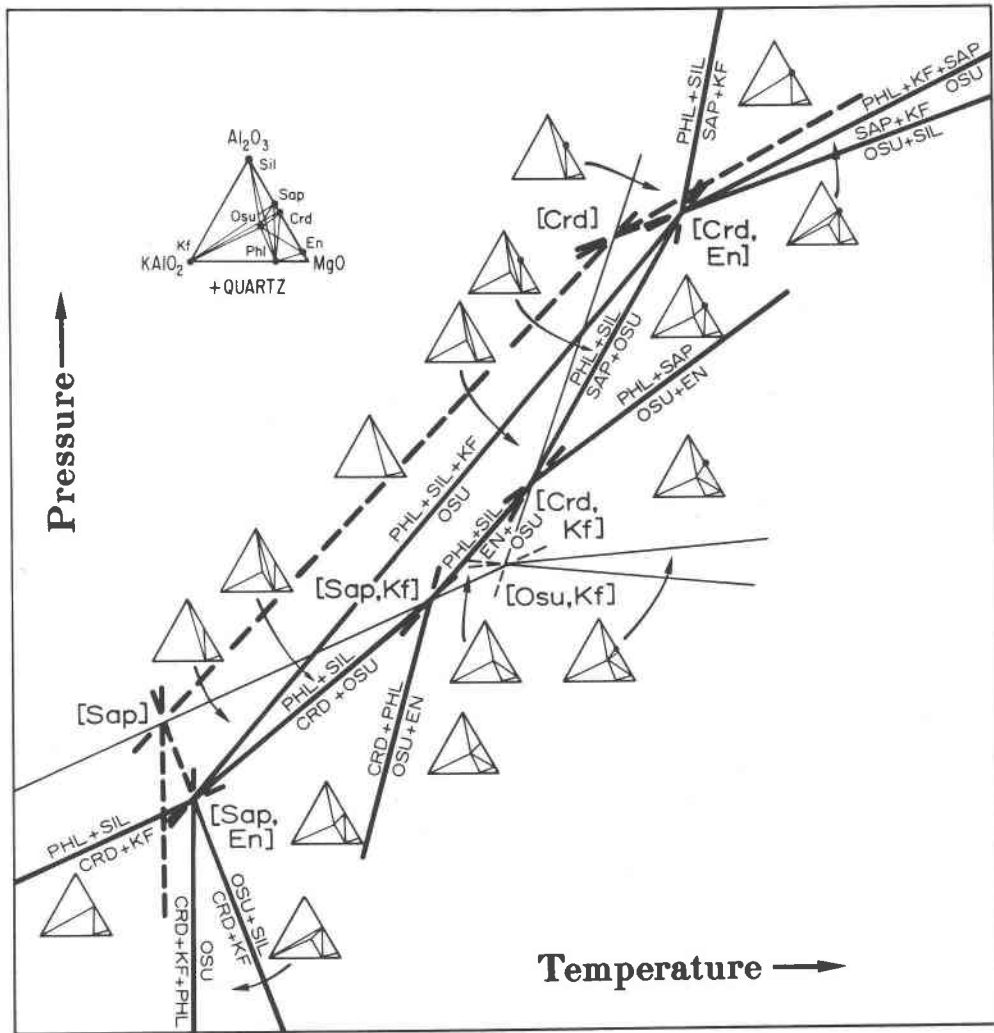
Fig. 13. Pressure–temperature diagram based on Schreinemakers construction of phase relations in the K_2O–MgO–Al_2O_3–SiO_2–H_2O system, in which phlogopite is involved. Metastable portions of K_2O–MgO–Al_2O_3–SiO_2 system indicated by dashed lines. Lighter lines are for the MgO–Al_2O_3–SiO_2 system. Reaction $Osu = Crd + Kf + Phl + Qtz + H_2O$ was studied experimentally by Olesch and Seifert (1981). Explanation of abbreviations and symbols given in Figures 10 and 11; except Phl = Phlogopite, and derivation explained in text.

ratio or X_{Fe} of associated osumilite and garnet range from 0.09 to 0.17 and 0.46 to 0.53, respectively, in Enderby Land rocks (Tables 1, 8, 9 and Ellis et al., 1980). The Brusilov osumilite, which is not associated with garnet, has an X_{Fe} of 0.21, the highest value obtained on Enderby Land osumilite. However, some of the iron in this sample may be ferric (see above). The X_{Fe} of osumilite associated with garnet from Rogaland ranges from 0.24 to 0.25 (Table 9; Maijer et al., 1977). The X_{Fe} values of osumilite, cordierite, and orthopyroxene from Nain are 0.23, 0.23, and 0.41 respectively in sample 2-1817 (Berg and Wheeler, 1976) and 0.39, 0.41, and 0.58 in sample #143967 (Table 8 and Grew, unpub. data),

and from Dzhugdzhur, 0.43–0.45, 0.42–0.43, and 0.61, respectively (Bogdanova et al., 1980). Clearly the most iron-rich osumilites are found in the Nain and Dzhugdzhur aureoles, and the X_{Fe} values of associated minerals suggest that these rocks may have a bulk X_{Fe} greater than those of the Enderby Land osumilite-bearing rocks. Thus the maximum possible X_{Fe} of the rock in which osumilite is stable is lower in rocks formed at higher pressures, in which osumilite occurs with garnet, than in rocks formed at lower pressures, where osumilite occurs with cordierite and orthopyroxene.

This apparent paradox can be explained by the marked shift in X_{Fe} of associated garnet, cordierite,

and orthopyroxene (or of associated garnet, cordierite, and sillimanite) with pressure: as P increases, X_{Fe} of the minerals in these assemblages decreases (Hensen and Green, 1971, 1972, 1973). Moreover, at pressures above the stability limit of cordierite, the X_{Fe} of garnet associated with sillimanite and orthopyroxene will continue to decrease with increasing pressure (Hensen and Green, 1971). Consequently, as pressure increases, osumilite associated with cordierite or garnet will become increasingly magnesian. As osumilite associated with garnet is more iron-rich (*i.e.* Fe^{2+}-rich) than osumilite associated with sapphirine or orthopyroxene, the maximum possible iron content in osumilite will decrease with increasing pressure. The shift of osumilite–garnet tie lines to more magnesian compositions means that osumilite becomes increasingly restricted to bulk compositions with only the highest Mg/Fe ratios.

Conclusion

The petrogenetic grids presented here, Olesch and Seifert's (1981) experimental data, Olesch's (1981) calculations, and the 900° C, 7 kbar estimate for metamorphic conditions of osumilite formation in the Napier complex imply that osumilite would be stable in rocks of appropriate bulk composition at temperatures above about 750° C, lithostatic pressures not exceeding 8 or 9 kbar, and at water pressures not exceeding 1 kbar. Moreover, a minimum pressure of about 5 kbar is indicated for the assemblages osumilite–garnet or osumilite–sillimanite at temperatures reasonable for deep-seated metamorphic rocks, that is, 900–1000° C. The appropriate bulk compositions at pressures of 5 kbar or less probably fall within the range of average metapelites. However, with increasing pressure, particularly above 7 kbar, the appropriate compositions are undoubtedly more magnesian than the average metapelite. Such magnesian metapelites, while nowhere abundant, are found in numerous granulite-facies terranes.

The metamorphic conditions indicated for osumilite formation overlap the estimated physical conditions for metamorphism in the granulite facies. Estimates of temperatures and pressures in granulite-facies terranes range mostly from 700° to 850° C and from 5 to 12 kbar. Recent estimates are, for example: 780±60° C, 9.5±0.5 kbar, Madras, India, (Weaver *et al.*, 1978); 790±50° C, 8±1 kbar, Adirondack Highlands, U.S.A. (Bohlen and Essene, 1977 and Bohlen *et al.*, 1980); 810±30° C, 10±1.5 kbar, West Greenland, (Wells, 1979); 725±25° C, 5 kbar, Finnish Lapland (Hörmann *et al.*, 1980); and 850±50° C, 12±2 kbar, NW Scotland, (Savage and Sills, 1980) (*cf.* 900±30° C, 7±1 kbar, Grew (1980a) and 940±40° C, 9±1 kbar, Ellis (1980) and Sheraton *et al.* (1980) for the Napier complex). Examples of P_{H_2O} estimates are 1–3 kbar in West Greenland (P_{H_2O} = 0.1–0.3 P_{Total}, Wells, 1979); 1–2 kbar in Finnish Lapland (P_{H_2O} = 0.25–0.35 P_{Total}, Hörmann, *et al.*, 1980), and ≤2 kbar in the Adirondacks (P_{H_2O} ≤ 0.25 P_{Total}, Bohlen and Essene, 1978) (*cf.* <0.5 kbar for the Napier complex, Sheraton *et al.*, 1980).

According to the above information, osumilite should occur in a larger number of granulite-facies terranes than has been reported in the literature. Possibly it has been overlooked, for it is similar in appearance to quartz and cordierite. Moreover, it commonly is replaced by a cordierite–quartz–K-feldspar symplectite that could be confused with other cordierite-rich symplectites. Schreyer and Seifert (1967) suggested that such a symplectite from Rogaland, Norway, was derived from osumilite 10 years before the discovery of unaltered osumilite by Maijer *et al.* (1977) in this area. Similar symplectites are also found in the Wilson Lake area, Labrador, Canada (see above) and at Paderu, Andhra Pradesh, in southern India (Grew, in prep.), localities from which no unaltered osumilite has been reported to date. More intense search may yet show that osumilite is an important constituent of some metapelitic rocks metamorphosed under conditions of relatively high temperatures and very low water pressures in the granulite facies.

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