

# Grandierite from aluminous metasedimentary xenoliths within acid volcanics, a first record in Italy

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**SUMMARY.** Grandierite has been identified in two aluminous metasedimentary xenoliths from quartz-latic volcanic rocks from Mt. Amiata and Mt. Cimino, central Italy. Physical and electron microprobe data for the grandierites and petrological data for the grandierite-bearing xenoliths are presented. The grandierites formed by a reaction involving pre-existing aluminium rich minerals, possibly at temperatures of at least 800 °C and at low pressures. The grandierite from Mt. Amiata replaces sillimanite. Several common characteristics can be demonstrated for magmatic and metamorphic grandierite-bearing rocks. It is suggested that metamorphic rocks in which grandierite occurs have often undergone partial melting.

THE rare Mg-, Fe-, Al-borosilicate grandierite has been identified in the course of a detailed study of metasedimentary xenoliths from acid volcanics of southern Tuscany and northern Latium, Italy. The volcanic rocks are part of two Pliocene-Pleistocene volcanic complexes: Mt. Amiata and Mt. Cimino (fig. 1). Both belong to the Tuscan-Latium Magmatic Province, which is characterized by extrusive and intrusive acidic rocks of crustal anatectic origin (Marinelli, 1967). The Mt. Cimino complex is surrounded and partly covered by the Quaternary potassic alkaline products of the Roman Magmatic Province. Chemical, petrological, and volcanological data for Mt. Amiata are

given by Mazzuoli and Pratesi (1963) and for Mt. Cimino by Puxeddu (1971) and Micheluccini *et al.* (1971). Most volcanic rocks of the complexes contain abundant, predominantly aluminous, xenoliths of metasedimentary origin. Grandierite was observed in only one xenolith from each complex. Hitherto grandierite has been described from aplitic and pegmatitic rocks by Lacroix (1922, cited by McKie, 1965) and Zav'yalova *et al.* (1975), from crystalline rocks associated with charnockites by McKie (1965) and von Knorring *et al.* (1969), from high-grade regional metamorphic rocks by de Villiers (1940),\* Krogh (1975), and Vrána (1979), from rocks in contact aureoles by Black (1970) and Brewer *et al.* (1979), and from xenoliths in intrusive bodies by de Roever and Kieft (1976) and Semroud *et al.* (1976). Grandierite has not been previously described from xenoliths in volcanic rocks, and the samples described in this paper represent the first recorded discovery of grandierite in Italy.

## Petrography

**Mt. Amiata sample.** The hornfels-like xenolith, MA 4.1 (about 15 × 5 cm), was collected from the Corno di Bellaria flow (Mazzuoli and Pratesi, 1963), which is a quartz-latic perlitic lava. The xenolith is fine grained and shows a concentric zonation. The central part, in which most of the grandierite is concentrated, consists mainly of sanidine, green spinel, and biotite, while plagioclase and biotite are the major phases in the ~ 0.5 cm wide rim. Hyperssthene is present in the outer part of this rim. The total amount of grandierite in the xenolith is less than 1%. Accessory minerals are sillimanite, ilmenite, graphite, apatite, and zircon. The spinel

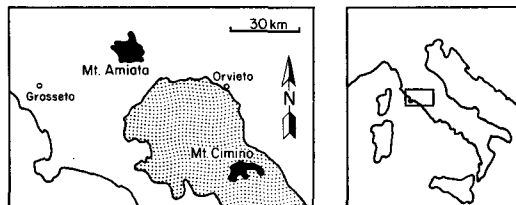


FIG. 1. Location of Mt. Amiata and Mt. Cimino. Black areas belong to the Tuscan-Latium Magmatic Province, dotted areas to the Roman Magmatic Province.

\* de Villiers considered his sample to be of pegmatitic origin. Following Vrána (1979) it is listed under regional metamorphic rocks, based on the similarities with the occurrence described by this author.

occurs both disseminated through the xenolith and in concentrations together with a turbid isotropic mass (presumably altered plagioclase) around grandidierite crystals (fig. 2*a*).

These concentrations have the shape of square prisms, up to 1 cm long, and are believed to be pseudomorphs after andalusite, which is often replaced by sillimanite during high-grade thermometamorphism before complete conversion into the spinel with or without plagioclase has taken

place, as observed in other xenoliths (van Bergen, in prep.). Parallel orientated ilmenite grains indicate a weak relict foliation.

Grandidierite occurs almost exclusively within the spinel pseudomorphs. It forms idiomorphic crystals with the shape of slender prisms, up to 1 cm long and less than 1 mm wide. A representative basal section is shown in fig. 2*b*. Locally a poor (100) cleavage is visible. The crystals have irregular cracks, parallel to the basal plane, and they often

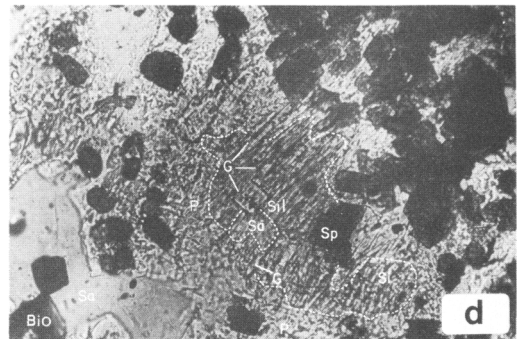
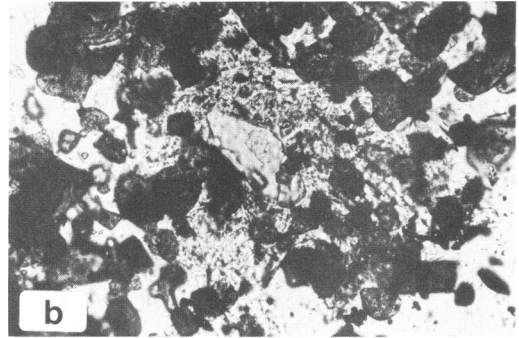
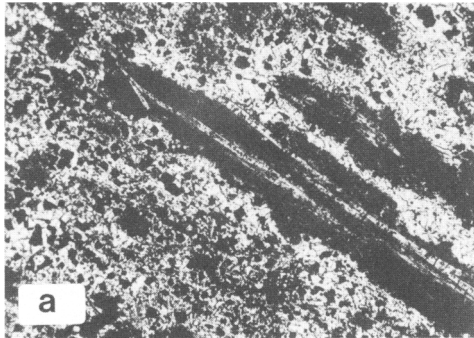
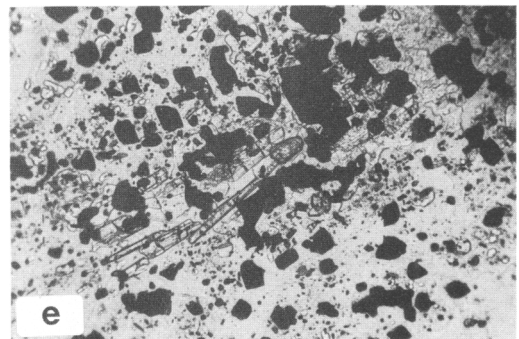


FIG. 2. (a) MA 4.1. Spinel pseudomorphs after andalusite/sillimanite porphyroblasts, containing slender grandidierite crystals. The matrix consists of sanidine, spinel, and biotite. Nicsols parallel. Length of photograph: 5.1 mm. (b) MA 4.1. Eight-sided basal section of grandidierite in an isotropic substance. The long axis of this section is parallel to  $\gamma$ . Nicsols parallel. Length of photograph: 0.55 mm. (c) MA 4.1. Grandidierite crystal in an aggregate of spinel and an isotropic substance (altered plagioclase?). Note the cracks parallel to the basal plane and the rounded inclusions of green spinel. Nicsols parallel. Length of photograph: 1.3 mm. (d) MA 4.1. Small grandidierite crystals in a relict of sillimanite enclosed by plagioclase. Nicsols parallel. Length of photograph: 0.55 mm. G = grandidierite; Sil = sillimanite; Sa = sanidine; Sp = spinel; P = plagioclase; Bio = biotite. (e) MC 6.9. Slender grandidierite crystals in association with two rounded grains of corundum (high relief). Dark grains: green spinel and minor biotite. Light areas: sanidine and corroded plagioclase. Nicsols parallel. Length of photograph: 1.3 mm.



contain small rounded spinel inclusions (fig. 2c). The *c* axes of the grandidierites are generally parallel to the long axes of the prism-shaped spinel pseudomorphs. Although closely associated with spinel, grandidierite also shows grain-to-grain contacts with biotite and feldspars. A sillimanite relict, containing several idiomorphic grandidierite crystals, was found in the core of one spinel pseudomorph (fig. 2d).

The grandidierite is strongly birefringent and displays a characteristic blue pleochroism. The extinction is straight and the elongation is variable. The optical axial plane is parallel to (001). Refractive indices and further optical data are given in Table I.

TABLE I. Optical data of grandidierite from Mt. Amiata and Mt. Cimino

Optical orientation:	$c \perp$ optical axial plane
Pleochroism:	$\alpha$ = bright blue $\beta$ = colourless $\gamma$ = light greenish blue
Absorption scheme:	$\alpha > \gamma > \beta$
Axial angle (U-stage):	$2V_{\alpha} = 32^{\circ} \pm 1^{\circ}$ (MA 4.1) $2V_{\alpha} = 27^{\circ} \pm 1^{\circ}$ (MC 6.9)
Dispersion:	$r < v$ clear
Birefringence:	$\Delta = 0.0364$
Refractive indices (Na light):	$\alpha = 1.5982 \pm 0.0002$ $\beta = 1.6290 \pm 0.0002$ $\gamma = 1.6346 \pm 0.0002$ (calculated)

Values for birefringence and refractive indices apply to MA 4.1.  $\alpha$  and  $\beta$  were determined by a double variation method;  $\gamma$  was calculated after determination of the birefringence with an Ehringhaus rotary compensator.

The X-ray pattern of a single grain from the Mt. Amiata sample is compared with data from Olesch and Seifert (1976, synthesized Mg end-member grandidierite) and McKie (1965, grandidierite from Sakatelo, Madagascar, with  $Fe/(Fe + Mg) = 0.118$ ) in Table II. This pattern identifies the mineral as grandidierite.

Sanidine is the major phase. It has a variable grain size. The smaller grains tend to form polygonal textures, especially where the sanidine is free of inclusions. Plagioclase grains are irregular in shape and are often separated from each other by a thin glass film. Glass, either alone or around small spinel grains, is also present as inclusions in the plagioclase. Some plagioclase grains show a slight normal zonation. Spinel, abundant and ubiquitous, is dark green and hercynitic. It forms subidiomorphic to idiomorphic crystals. Biotite is strongly pleochroic:  $\alpha$  yellow and  $\gamma$  dark brown. The randomly orientated crystals often contain ilmenite and spinel inclusions.

*Mt. Cimino sample.* The xenolith, MC 6.9 (about  $12 \times 6$  cm), has roughly the same characteristics as the Mt. Amiata sample. It was found in a quartz-latic ignimbrite near Bagnaia (see map by Micheluccini *et al.*, 1971). The main components are sanidine, green spinel, biotite, and plagioclase. Corundum is present as relatively large anidomorphic grains, mainly in spinel-rich parts (fig. 2e). The zonation within the xenolith and the accessory mineral content are similar to MA 4.1, except for the absence of sillimanite relicts. The spinel pseudomorphs, which occur in MA 4.1, are also lacking. Grandidierite is disseminated through the rock and is usually accompanied by spinel. It has essentially the same properties as in MA 4.1, but the crystals are considerably smaller and they are corroded along crystal boundaries and cracks. The largest dimensions rarely exceed 100 microns. Several crystals are enclosed in biotite. Apart from contact with biotite there are also grain-to-grain contacts with feldspars and corundum. Optical data are given in Table I and a single-crystal X-ray pattern in Table II. Plagioclase is not limited to a plagioclase-biotite rim. The characteristics of the other main phases do not differ from MA 4.1.

#### Mineral chemistry

Electron microprobe analyses of the grandidierites were carried out at two laboratories using different apparatus. The results and mean values are listed in Table III. The boron contents, which are expected to be about 10–12 wt %, could not be determined. The analyses are in good agreement with data in the literature (cf. Semroud *et al.*, 1976, Table I). The  $Fe/(Fe + Mg)$  ratios, calculated from the mean values, are relatively high and in fact the MA 4.1 grandidierite is the most Fe-rich natural specimen described to date. The combination of the high  $Fe/(Fe + Mg)$  ratio and the high refractive indices (Table I) in this sample is in accordance with the positive linear correlation between the mean refractive indices of grandidierite and its  $X_{Fe}$ , which was demonstrated by Olesch and Seifert (1976).

Differences in mineral chemistry between the two xenoliths are small. The Mt. Amiata grandidierite has a higher  $Fe/(Fe + Mg)$  ratio. The  $Fe/(Fe + Mg)$  ratios of seventeen co-existing spinel and biotite grains show varying values. In MA 4.1 the  $X_{Fe}^{sp}$  ranges from 0.75 to 0.79 and the  $X_{Fe}^{bt}$  from 0.44 to 0.53; in MC 6.9 the  $X_{Fe}^{sp}$  ranges from 0.67 to 0.75 and the  $X_{Fe}^{bt}$  from 0.50 to 0.57. The biotites in the plagioclase-biotite rims have the lowest Fe contents. The hypersthene has not been considered because of its isolated position in the outer part of the plagioclase-biotite rims, and because of its

probable formation by reaction with the surrounding magma.

As most analyses were carried out on minerals in close proximity to each other, the  $X_{Fe}$  values may be representative of at least local equilibrium, especially regarding the closely associated grandierite and spinel or part of the spinel. Although total rock equilibrium may not have been reached, the following general relation holds:

$$X_{Fe}^{spi} > X_{Fe}^{bi} > X_{Fe}^{grand}$$

Microprobe analyses of sanidines yielded  $Ab_{1.6}Or_{8.3}An_1$  for MA 4.1 and  $Ab_{1.8}Or_{8.1}An_1$  for MC 6.9. An percentages of unzoned plagioclases range from 55 to 58% and from 47 to 73% respectively. An contents increase from the core to the rim of the xenoliths, i.e. towards the volcanic rock.

#### Genesis of the grandierite

The following genetic sequence for the grandierite-bearing core of sample MA 4.1 is

TABLE II. X-ray patterns of the grandierites from Mt. Cimino and Mt. Amiata compared with the grandierite from Sakatelo ( $Fe/(Fe + Mg) = 0.118$ ) and a synthetic Mg end-member grandierite

Mt. Cimino (MC 6.9)		Mt. Amiata (MA 4.1)		Sakatelo (McKie, 1965)		Synthetic (Olesch and Seifert, 1976)	
$d_{obs}$	<i>I</i>	$d_{obs}$	<i>I</i>	$d_{obs}$	<i>I</i>	$d_{obs}$	<i>I</i>
5.53	80	5.53	90	5.482	vs	5.451	100
5.17	100	5.21	100	5.17	vvs	5.179	91
		5.05	90	5.04	vvs	5.019	87
4.86	40	4.86	20	4.84	w	4.816	12
				4.59	vw		
		3.97	10	3.97	vw	3.958	12
3.78	70			3.75	vw	3.752	15
		3.71	50	3.708	ms	3.693	55
3.46	50	3.46	20	3.449	w	3.431	18
3.28	10	3.30	10	3.29	vw	3.291	5
3.13	5	3.15	20	3.150	w	3.146	13
				2.99	vw		
2.95	15	2.96	30	2.956	m	2.957	23
2.92	5			2.92	vw		
				2.878	w	2.868	20
2.85	5	2.85	10	2.855	vw		
2.75	85	2.75	80	2.744	vs	2.732	77
		2.66	10	2.66B	vw		
2.60	15	2.60	40	2.602	m		
2.57	20	2.56B	40	2.584	ms	2.595	38
				2.551	mw	2.540	20
		2.51	20	2.515	w	2.512	20
		2.45	10	2.48B	vw	2.468	2
		2.40	5	2.41B	vw		
2.33	5	2.34	10	2.337	vw	2.342	4
		2.29	30	2.287	mw	2.281	21
		2.21	30	2.210	mw	2.202	21
2.15	60	2.16	60	2.166	s	2.165	34
				2.149	m	2.135	20
				2.072	vw		
		2.05	10	2.050	vw		
2.02	20	2.03	20	2.031	mw	2.039	11
		2.01	10	2.012B	vw		
1.948	20	1.949	30	1.946	mw	1.9501	13
numerous weak reflections		numerous weak reflections				numerous weak reflections	

MC 6.9: single crystal in a Debye-Scherrer camera, Co-K $\alpha$  radiation.

MA 4.1: single crystal in a Gandolfi camera containing a two-axes rotating sample-holder, Co-K $\alpha$  radiation.

Both determinations were made at the Z.W.O. Laboratory for Isotope Geology, Amsterdam, by R. H. Verschure.

TABLE III. Chemical compositions of the grandidierites from Mt. Amiata (MA 4.1) and Mt. Cimino (MC 6.9)

	MA 4.1			MC 6.9		
	1	2	mean	3	4	mean
SiO <sub>2</sub>	19.9	19.9	19.9	20.0	19.8	19.9
TiO <sub>2</sub>	n.d.	0.02	0.02	n.d.	0.04	0.04
Al <sub>2</sub> O <sub>3</sub>	50.8	49.9	50.35	50.9	50.3	50.6
FeO	7.8	7.7	7.75	6.6	6.9	6.75
MnO	0.14	0.1	0.12	0.08	0.1	0.09
MgO	9.1	9.3	9.2	9.8	9.7	9.75
CaO	n.d.	0.00	0.00	n.d.	0.01	0.01
Na <sub>2</sub> O	n.d.	0.01	0.01	n.d.	0.00	0.00
K <sub>2</sub> O	n.d.	0.04	0.04	n.d.	0.06	0.06
Total	87.74	86.97	87.39	87.38	86.91	87.20
Numbers of atoms on the basis of fifteen oxygens						
Si	1.99	2.01	2.00	2.00	1.99	2.00
Ti	—	0.00	0.00	—	0.00	0.00
Al	5.99	5.94	5.97	5.99	5.97	5.98
Fe	0.65	0.65	0.65	0.55	0.58	0.57
Mn	0.01	0.01	0.01	0.01	0.01	0.01
Mg	1.36	1.40	1.38	1.46	1.46	1.46
Ca	—	0.00	0.00	—	0.00	0.00
Na	—	0.00	0.00	—	0.00	0.00
K	—	0.00	0.00	—	0.01	0.01
Fe/(Fe + Mg)			0.320			0.281

Electron microprobe analyses on the same polished thin sections were performed both at the electron microprobe laboratory of the Instituut voor Aardwetenschappen, Vrije Universiteit Amsterdam (1 and 3) and at the Vening Meinesz Laboratorium voor Geochemie en Mineralogie, Rijks Universiteit Utrecht (2 and 4) by P. Maaskant and the author respectively. 1, 2, 3 and 4: mean values of at least four grains. n.d.: not determined.

based on textural relations, including those observed in similar xenoliths (van Bergen, in prep.). Sillimanite, which is probably a paramorph after andalusite, is clearly a relict phase. Green spinel partly replaced the sillimanite along crystal boundaries and it formed spinel pseudomorphs. In other xenoliths sillimanite and pre-existing andalusite are often replaced by spinel with or without plagioclase. Subsequently the remaining sillimanite was converted to grandidierite. Some of the spinel was used in this reaction. The often-observed preferential orientation in the pseudomorphs may be the result of some similarity in crystal structure between grandidierite and sillimanite. According to Stephenson and Moore (1968) the structural relationship with andalusite is closer, however. Biotite is generally a late phase, although the presence of an early generation cannot be excluded. Sanidine is clearly the last-formed mineral.

Because in MC 6.9 sillimanite is absent and the grandidierite is more or less randomly distributed,

there is no evidence that grandidierite formed by the same mechanism as in MA 4.1. The genetic sequence of spinel, grandidierite, biotite, and sanidine is similar. Plagioclase is a late phase but is prior to sanidine. Time relations between corundum and grandidierite and between corundum and spinel are not quite clear. Other xenoliths show corundum to be contemporaneous with or later than spinel. That grandidierite is found in spinel-rich parts of the sample and that it contains small rounded spinel inclusions emphasize a genetic relation with the spinel.

Two stages can be distinguished in the thermometamorphism which in general affected the Mt. Amiata and Mt. Cimino xenoliths: a contact-metamorphic stage and a subsequent pyrometamorphic stage (Rodolico, 1935). The grandidierite-bearing samples show aspects of both stages. Although a strict separation between these stages is difficult, the generation of grandidierite is most likely to have taken place during the latter.

Crystallization of grandidierite after engulfment of the xenoliths by the magma might point to a metasomatic supply of boron from the magma. However, the scarcity of grandidierite-bearing xenoliths, and the fact that in the same outcrops similar xenoliths do not contain grandidierite (nor tourmaline nor any other boron mineral) make it improbable that the magma was the source of boron. The small amounts of grandidierite may have derived boron from primary tourmaline, either directly by a high-temperature transition of tourmaline into grandidierite (MC 6.9 (?)), or after a small-scale redistribution of boron (MA 4.1, MC 6.9 (?)) possibly by means of an interstitial melt. Regarding other grandidierite occurrences in metamorphic environments, most authors are also inclined to attribute its presence to a boron enrichment of the original sediments (Krogh, 1975; Semroud *et al.*, 1976; de Roever and Kieft, 1976) although elsewhere boron metasomatism may have taken place (e.g. Black, 1970).

#### Comparison with other occurrences

Grandidierite has been described from pegmatitic and aplitic rocks, crystalline rocks associated with charnockites, high-grade regional metamorphic rocks, and thermometamorphic rocks (contact aureoles and xenoliths). Although petrological data on some grandidierite occurrences, especially those in Madagascar, are inadequate, it seems possible to recognize some common characteristics in all of these rocks and to discuss the implications with respect to the formation of grandidierite. All occurrences known to the author are listed in Table IV, together with mineral associations and references.

TABLE IV. Mineral contents of grandierite-bearing rocks (7a excluded)

	quartz	K-feldspar	plagioclase	spinel	corundum	sillimanite	andalusite	cordierite	biotite	garnet	hypersthene	tourmaline	kornerupine
1. Mt. Amiata (Italy)		x	x <sup>1</sup>	x		r			x		x <sup>1</sup>		
2. Mt. Cimino (Italy)		x	x	x	x				x		x <sup>1</sup>		
3. Maratakka (Surinam)		x	x	x	x			x	x				
4. Tizi-Ouchen (Algeria)				x	x				x				
5. Blanket Bay (Cuvier Island, New Zealand)													
6. Landing Bay (Cuvier Island, New Zealand)			x					x	x	SP	x	s	
7. Transbaikaliya (U.S.S.R.)	(a)	x	x	x					x		x	s	
	(b)	x <sup>2</sup>	x <sup>2</sup>	x				x	x		x		
8. Andrahomana (Madagascar)		x	x						x			x	
9. Sakatelo (Madagascar)									x	ALM			
10. Sahakondro (Madagascar)									x				
11. Fort Dauphin area (Madagascar) <sup>3</sup>				x					PHL				
				x					PHL				
12. Vestpolltind (Norway)		x	x						x	x			
13. Lukusuzi river (Zambia) <sup>4</sup>		x			m	x				x		x	x
14. Natal (S. Africa)		x							x	m		x	x

1-4: thermometamorphic xenoliths; 5-6: hornfelses from contact aureole; 7: hornfels from contact aureole; without grandierite (a), containing aplitic and granitic grandierite-bearing veinlets (b); 8: pegmatitic and aplitic rocks; 9-11: rocks in association with charnockites; 12-14: high-grade regional (poly-) metamorphic rocks.

<sup>1</sup> Present only in outer rims of the xenoliths. <sup>2</sup> Present only in interstitial segregations. <sup>3</sup> With diopside. <sup>4</sup> With dumortierite.

r = relict, m = in minor amounts, s = secondary, SP = spessartine, ALM = almandine, PHL = phlogopite.

References: 1 and 2—this work; 3—de Roever and Kieft (1976); 4—Semroud *et al.* (1976); 5 and 6—Black (1970); 7—Zav'yalova *et al.* (1975); 8—Lacroix (1922, cited by McKie, 1965) and Semroud *et al.* (1976); 9—McKie (1965); 10—Christophe-Michel-Lévy *et al.* (1959) and de la Roche (1963); 11—von Knorring *et al.* (1969); 12—Krogh (1975); 13—Vrána (1979); 14—de Villiers (1940).

In addition, a recently reported occurrence in the aureole of a hypersthene granodiorite in the Mchinji area, Malawi, may be mentioned (Brewer *et al.*, 1979). Here grandierite and tourmaline occur sporadically in an originally low-grade quartz-rich metasedimentary rock, the former only in association with sillimanite that probably originated during thermometamorphism. The uncertain occurrence recorded by Hlawatsch (1918, cited by McKie, 1965) has not been included.

Apart from the common presence of biotite/phlogopite, the dominant presence of aluminium-rich minerals is striking. The genesis of grandierite is related to aluminous rocks, which is in accordance with its mineral chemistry (Semroud *et al.*, 1976). These authors pay attention to the fact that in magmatic and metamorphic rocks grandierite usually crystallizes after the aluminium-rich minerals and before feldspars and quartz (if present). The Mt. Amiata and Mt. Cimino occurrences are good examples of this genetic sequence.

Direct crystallization of grandierite instead of an  $Al_2SiO_5$  phase in boron-enriched environments is suggested by McKie (1965) for contaminated magmas, and by Krogh (1975) for metamorphic rocks. In experiments Rosenberg and Foit (1975, cited by Olesch and Seifert, 1976) found grandierite as a high-temperature breakdown product of tourmaline.

For the formation of grandierite, reaction of aluminium-rich minerals with boron or boron minerals must also be taken into account, especially with regard to the sillimanite-bearing Mt. Amiata rock and the above-mentioned genetic sequence. Evidence for reaction with other boron minerals is scarce. The presence of tourmaline next to grandierite is reported only from Transbaikaliya (Zav'yalova *et al.*, 1975), while in the New Zealand occurrences tourmaline belongs to a late hydrothermal phase (Black, 1970), and in both the Natal and Lukusuzi River occurrences grandierite predates kornerupine and tourmaline (de Villiers,

1940; Vrána, 1979). The reaction with boron must be seriously considered. This element may be derived from a melt phase, the presence of which is obvious for the magmatic occurrences. The former presence of an interstitial melt phase in several other occurrences cannot be excluded. Partial melting could have taken place to a limited extent in the central Italian xenoliths. De Roever and Kieft (1976) suggest partial anatexis for the Maratakka xenoliths and it is interesting to notice the common absence of at least one of the leucocratic minerals (quartz, plagioclase, K feldspar) in almost all non-vein rocks (Table IV). In Transbaikaliya grandidierite-bearing aplitic and granitic veinlets cut hornfelses with interstitial segregations of quartz and K-feldspar (Zav'yalova *et al.*, 1975). Partial melting may lead to the aluminous environments favourable for the formation of grandidierite, and the conditions of partial melting are compatible with some independent *P-T* estimations on the stability of this mineral (see below). Consequently a grandidierite-forming reaction involving aluminium minerals and boron from a melt-phase must be considered for some of the magmatic as well as the metamorphic rock types.

#### *Stability of grandidierite*

The pegmatitic and aplitic occurrences indicate that grandidierite is stable under relatively high-temperature conditions. Estimates of temperatures for the metamorphic occurrences yielded minimum values of 600 °C for the Cuvier Island aureole (Black, 1970) and 700 °C for the Maratakka xenoliths (de Roever and Kieft, 1976), both based on mineral assemblages. Krogh (1975), using various geothermometers, calculated temperatures of 850–900 °C for the Vestpolltind area. Pressure estimates range from less than 300 bars (Cuvier Island) to 10 kb (Vestpolltind). As for the central Italian occurrences Puxeddu (1971) estimated the lower limits of temperature of the Mt. Cimino magma at its origin between 850 and 880 °C on chemical grounds, and because of the similarity between the two magmas approximately the same temperatures could apply to the Mt. Amiata magma. The mineral contents of the grandidierite-bearing xenoliths within the volcanic rocks of Mt. Cimino and Mt. Amiata point to temperatures of the sanidinite facies. The occurrence of grandidierite in xenoliths in lavas suggests that at high temperatures this mineral is stable at very low pressures.

In synthetic systems grandidierite was first mentioned by Rosenberg and Foit (*op. cit.*) as a high-temperature breakdown product of alkali-free tourmalines. Olesch and Seifert (1976) synthesized the Mg end-member at  $P_{H_2O} = 1$  kbar and tem-

peratures above 780 °C, and a grandidierite with  $X_{Fe} = 0.1$  at 700 °C with QFM buffer. From an experimental study of kornerupine, Werdning and Schreyer (1978) suggested that grandidierite is restricted to environments with relatively high boron/water ratios.

In general grandidierite is a high-temperature mineral, probably stable from very low to high pressures. It is likely to be the natural high-temperature breakdown product of tourmaline and a stable equivalent of this mineral in boron-enriched aluminous rock systems. While tourmaline is a common mineral in contact aureoles around acid intrusives, grandidierite is often associated with thermometamorphic rocks in relation to intrusive bodies which are generally more basic: granodioritic (Tizi-Ouchen, Mchinji area), dioritic (Cuvier Island), essexitic-granodioritic (Transbaikaliya), gabbroic (Maratakka). If the influence of total-rock chemistry (e.g. Na content) can be neglected, this can be explained by temperature differences. The high temperature of more basic intrusions favours the growth of grandidierite rather than tourmaline, provided that enough boron is present in the country rock at the peak of thermometamorphism. If the boron is supplied by the magma during cooling, tourmaline may be expected to be the stable boron mineral also around more basic intrusives.

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