

Petrology of an eclogite–metarodinite suite at Cima di Gagnone, Ticino, Switzerland

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Abstract

Mafic rocks composed essentially of garnet + clinopyroxene occur in association with metamorphic ultramafic rocks in the Penninic Adula–Cima Lunga Nappe around the summit of Cima di Gagnone, Ticino, Switzerland. The mafic rocks are interpreted as basaltic volcanics and minor intrusives and their mildly and strongly rodingitized products, later metamorphosed under eclogite facies conditions. They show continuous variation in their bulk chemical, mineralogical, and textural properties from eclogite (metabasaltic, Cpx with Jd > 10 percent) to metarodinite (CaO \approx 24 weight percent, Na₂O < 0.1 weight percent). Garnets in this series vary continuously from 84 percent pyrospite, 16 percent ugrandite (in eclogite), to 42 percent pyrospite, 58 percent ugrandite; some highly calcic metarodinites contain garnet with 80 percent ugrandite. Clinopyroxenes vary correspondingly from omphacite (45 percent Jd) to diopside (<1 percent Jd) and eventually to fassaite (23 percent Cats). The metarodinites differ from grosspyrites and kyanite eclogites in kimberlites by the absence of kyanite and presence of diopside rather than omphacite; they also differ in bulk composition.

Abundances of some trace and minor elements appear to have been little affected by the process of rodingitization or by the succeeding episodes of eclogite and amphibolite facies metamorphism. They suggest sea-floor tholeiitic basalt affinities and a possible oceanic origin for the ultramafic–mafic rock suite at Cima di Gagnone. An eclogite facies garnet lherzolite assemblage is preserved in one of the ultramafic bodies.

The K_D for Fe_{total}/Mg partitioning between garnet and clinopyroxene averages 6 in the eclogites, and rises, with increase in grossular component in the garnet, to 15 in the more calcic metarodinites. The Råheim and Green temperature calibration applied to eclogites with garnets containing up to 25 percent grossular component, and a new calibration by Ganguly suitable for low-Na pyroxene–calcic garnet pairs applied to the metarodinites, give comparable values and suggest conditions for the eclogite facies metamorphism were likely to have been 800°C, $P \approx$ 25 kbar. Similar values have previously been obtained for the Gagnone garnet lherzolite. These conditions do not overlap those estimated for the Tertiary Central Alpine Barrovian-style metamorphism in the area (600 \pm 100°C, $P <$ 10 kbar).

The Central Alpine metamorphic overprint did not reequilibrate Fe and Mg between garnet and pyroxene. However, it partially amphibolitized the eclogites and was responsible for hornblende and epidote growth in the metarodinites. Contact reaction zones (probably polymetamorphic) between metarodinite boudins and ultramafic rock contain hornblende, diopside, epidote, sphene, and chlorite; contact zones against eclogite contain the same minerals, except for garnet instead of diopside. Phlogopite and staurolite have been found in amphibolitized eclogite.

Assuming metasomatic rodingitization to be an exclusive accompaniment of serpentinization, the mafic-ultramafic rock suite at Cima di Gagnone must, at an early stage in its metamorphic history, have been in a low T , low P environment. This was followed by subduction and eclogite facies metamorphism, then tectonic uprise to shallower depths, and regional Barrovian-style metamorphism.

General introduction

This is a petrological study of a suite of mafic rocks surrounding and enclosed in lenses of metamorphic ultramafic rock in amphibolite facies gneisses in the vicinity of Cima di Gagnone, Valle Verzasca, Ticino, Switzerland. The mafic rocks comprise amphibolites, eclogites, metamorphosed rodingites, and rocks transitional between them, together with the products of contact reaction between mafic and ultramafic rock. This paper describes the completely gradational set of bulk chemical, textural, and mineralogical properties of the eclogite-metarodingite suite. These mafic rocks together with associated ultramafics record geologic, metamorphic, and tectonic events pre-dating the Tertiary Central Alpine, Barrovian-style regional metamorphism.

Regional introduction

The Cima di Gagnone area is located in the Lepontine Alps 25 km N of Locarno, between the Ticino and Verzasca Valleys. It is situated in the heart of the Central Alpine metamorphic belt close to the sillimanite isograd. Structurally, the area lies within the lower Pennine Nappes, between the N-S trending axial culmination of the Ticino (Leventina Nappe) to the E and the steeply plunging Maggia Zone to the W (Fig. 1). It includes the boundary zone between the Simano and the Cima Lunga (=Adula) Nappes (Preiswerk *et al.*, 1934). Isoclinally folded rocks in this area possess a flat-lying regional foliation and axial planes that dip progressively steeper south towards the S, where they merge into the steep, E-W trending zones close to the Insubric Line (Wenk, 1955).

Boundaries between structural units (nappes) in this general area have traditionally been drawn along metacarbonate zones, on the assumption, based on the lower-grade Pennine nappes, that the metacarbonate rocks are metamorphosed Mesozoic sediments. At Cima di Gagnone, they are impure calcareous sediments varying in thickness from a few centimeters to several tens of meters. The metacarbonate rocks are accompanied by a distinctive suite of rock types that are largely absent from the monotonous quartzo-feldspathic gneisses of the over- and underlying nappes, *viz.* ultramafic lenses and macrobou-

dins, amphibolites (some eclogitic), and semi-pelitic gneisses. Alpe Arami, 15 km SE of Cima di Gagnone, is probably the best-known locality in this zone. This suite has been tentatively interpreted as an association of highly metamorphosed oceanic rocks belonging to the Tethyan plate. The lherzolitic composition of the ultramafics (which include garnet peridotite) has suggested, on the other hand, correlation with the subcontinental mantle of the south European plate. The age of the suite—Mesozoic or pre-Mesozoic—is still unknown. Dal Vesco (1953) gave special attention to the mafic/ultramafic components of the suite. Additional information on the regional geology may be found in Wenk (1943, 1967, p. 418–430), Evans and Trommsdorff (1974), Heinrich (1978), and Stäubli (1978).

The grade of metamorphism in the Gagnone area has been determined in the course of numerous studies on progressive regional metamorphism in the Central Alps (Trommsdorff and Evans, 1974, and others). The assemblage kyanite + staurolite + muscovite + quartz characterizes the metapelitic gneisses, whereas in quartz veins all three Al_2SiO_5 polymorphs can be found (Heinrich, 1978), although early kyanite dominates over late andalusite and fibrolite (*e.g.*, at Passo di Gagnone, Fig. 2). Metacarbonate rocks contain calcite, dolomite, tremolite, diopside, quartz, calcic plagioclase, and scapolite. Almandine-andesine amphibolite can be shown in many examples to have formed from an earlier eclogitic assemblage. Ultramafic rocks are composed of olivine, anthophyllite or cummingtonite, tremolite, chlorite, chrome-spinel, and at least two generations of enstatite. The metamorphism involved at least an earlier high-pressure eclogite facies (defined by the parageneses omphacite + pyralpsite and olivine + pyrope) and a later lower-pressure amphibolite facies event. This last event, the Central Alpine metamorphism, was Tertiary in age, as has long been known from field and more recently from geochronological studies (Jäger *et al.*, 1967; Frey *et al.*, 1974).

Field relations of the mafic rocks

This study concentrates on the mafic rocks occurring in contact association with the ultramafic bodies around Cima di Gagnone (Fig. 2). Layers of variably

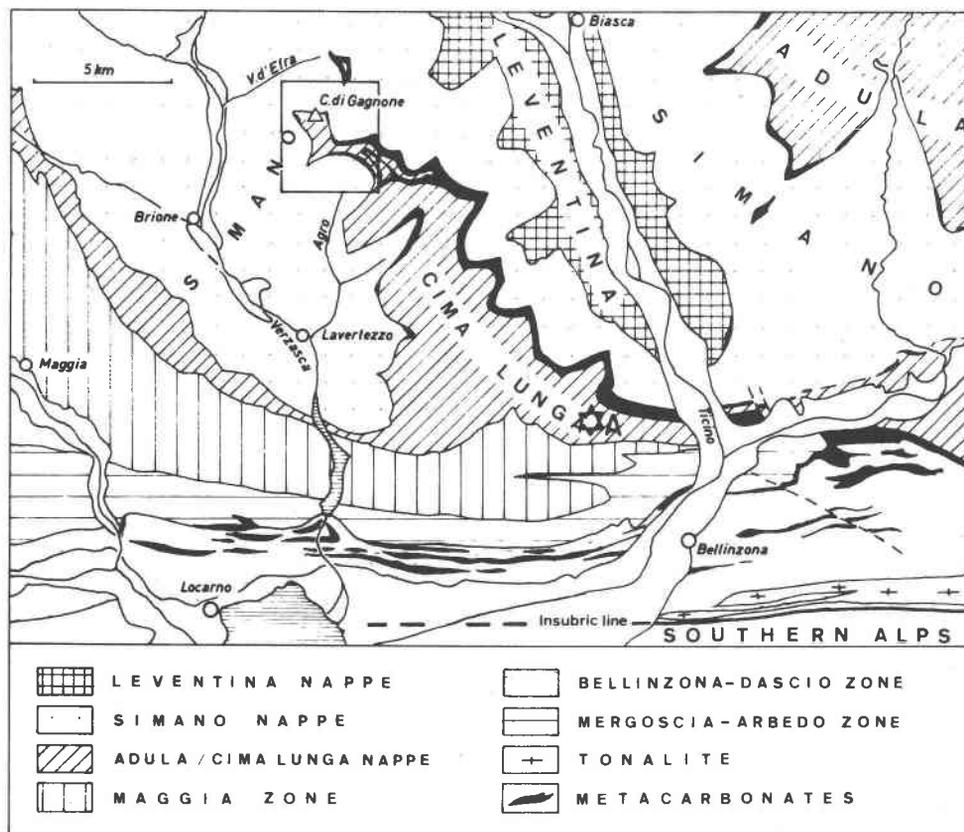


Fig. 1. Tectonic map of the southern Lepontine Alps. Small area around Cima di Gagnone is enlarged in Fig. 2. Alpe Arami is at A.

amphibolitized eclogite occur along the margins of the ultramafic bodies and in the nearby semi-pelitic gneisses. Metamorphosed rodingites (*i.e.*, calc-silicate rocks formed by Ca-enrichment of dikes or country rock during serpentinization) and rocks transitional between eclogite and metarodingite occur inside the ultramafics. They form thin boudinaged sheets, almost everywhere parallel to the compositional banding and foliation in the ultramafics.

The *eclogites*, which we will define here as consisting of pyralspite and pyroxene with more than 10 percent jadeite component, are generally banded rocks forming layers (Fig. 3) typically one to three meters thick and tens of meters long. The banded appearance is due to variations in the relative amounts of the principal minerals and of amphibole, epidote, and symplektite. A weak foliation parallel to the banding is marked by the alignment of pyroxene, hornblende, and zoisite. The grain size of the principal constituents is 1 to 2 mm, rounded garnets being the largest and most conspicuous.

The *metarodingites* are pink, relatively fine-grained highly calcic rocks, in places faintly banded, and are

commonly cross-cut by irregular thin veins of garnet, pyroxene (less common), and secondary hornblende. Compared to the eclogites, the metarodingites form thinner bodies, seldom exceeding one meter in thickness, and are strongly boudinaged. Recognizable metarodingite boudins may be as small as one or two centimeters across.

Dark, amphibole-rich *contact reaction zones* (Fig. 4) surround all metarodingite boudins. There are normally three such reaction zones around each boudin; from metarodingite to ultramafic rock, these are: (1) a green zone consisting of hornblende-epidote symplektite and diopside sharply bounded against the metarodingite, (2) a black zone of hornblende and epidote (locally with megacrysts of sphene), and (3) a coarse-grained heterogeneous zone of actinolite and chlorite, with chlorite becoming more abundant outwards. Reaction zones between eclogite and ultramafic rock are not so simple and clear as those around metarodingite. From eclogite to ultramafic, they are made up of the sequence: (1) garnet + symplektite (after omphacite), (2) hornblende, passing outwards into actinolite, and (3) chlorite ± actinolite.

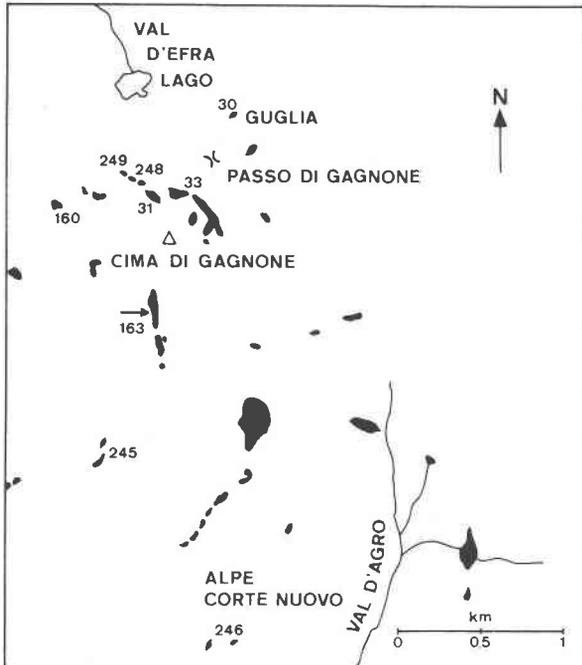


Fig. 2. Location of ultramafic bodies (with field numbers) in the region surrounding Cima di Gagnone, based on maps of the authors, Stäubli (1978), and Heinrich (1978). Arrow indicates location of section in Fig. 3.

The principal difference between the reaction zones against metarodingite and eclogite is the abrupt disappearance at the beginning of zone 1 of garnet in the former and of omphacite in the latter.

A sizeable fraction of the garnet-bearing mafic boudins in the region possess structural and textural characteristics *transitional* between eclogite and Ca-rich metarodingite. These rocks, as will be shown

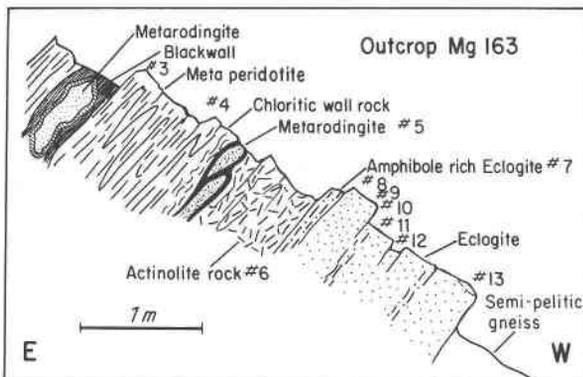


Fig. 3. Cross-section of the western, hanging-wall side of ultramafic lens 163, showing typical mode of occurrence of eclogite and metarodingite. Abbreviated specimen numbers are given (e.g., #11 is 163-4-11, Table 1).

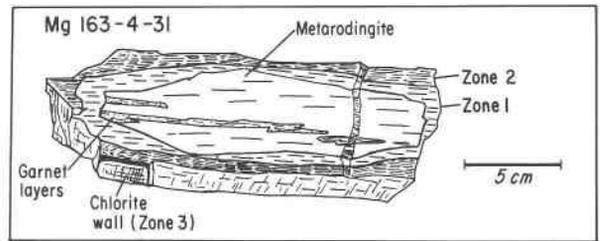


Fig. 4. Sketch of hand specimen of zoned metarodingite lens from ultramafic lens 163. Zones 1, 2, and 3 are described in the text.

below, have correspondingly intermediate bulk chemical and mineralogical properties. They are interpreted as rocks in which the original process of rodingitization was incomplete. The cut-off point between rocks labelled metarodingite and transitional can only be arbitrary. Transitional rocks have not as yet been found spatially separating eclogite and metarodingite in one single mafic layer. Indeed, a metarodingite body with an eclogite interior, such as might form by metamorphism of a basaltic dike with rodingitized margins, has not yet been discovered in the area.

In a few places, gradations from eclogite into pyralispitic garnet rock and from metarodingite into ugranditic garnet rock have been observed.

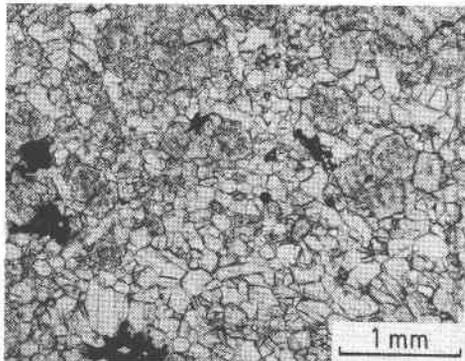
Microscopic petrography

Eclogite

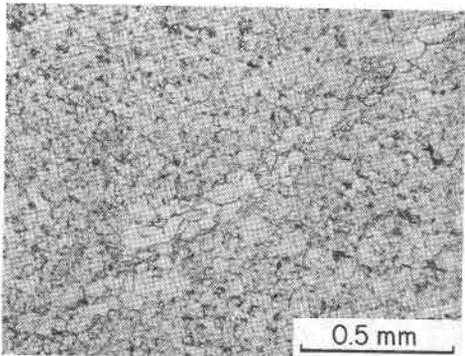
Different specimens show a process of amphibolitization developed to varying degrees. The early eclogitic assemblage includes garnet, omphacite, and rutile. Clinozoisite and pale brownish-green hornblende probably also belong to the eclogite assemblage (Heinrich, 1978). Unlike at Alpe Arami (Dal Vesco, 1953; Ernst, 1977), kyanite eclogite is rare, although pseudomorphs of corundum + phlogopite after kyanite are not uncommon (Heinrich, 1978). Accessory minerals include zircon, apatite, and quartz. Evenly distributed garnets tend to be porphyroblastic, and except in their marginal zones, contain tiny inclusions, some making S-shaped and planar trails. Some garnets are atoll-shaped, even in samples that escaped amphibolitization. Omphacite may be elongate and in some cases poikiloblastic to smaller garnets.

Amphibolitization reduces the modal amount of garnet and pyroxene, and is concentrated at the margins of the eclogite layers or parallel to the banding and foliation within the body. The garnets tend to

become atoll-shaped or highly irregular, but in many rocks seem to remain part of the assemblage, that is, show evidence of textural equilibrium with the hornblende. The products of identifiable garnet breakdown are epidote and hornblende, in some cases in symplectitic intergrowth with zoned poikiloblastic plagioclase. Omphacite is initially partially altered to a fine-grained symplectite which afterwards becomes replaced by hornblende. Formation of a more diopside clinopyroxene also seems to accompany increasing amphibolitization. Epidote grains tend to be zoned to more ferric compositions outwards; the less ferric examples are idioblastic and elongate. Rutile becomes rimmed and replaced by sphene, but on occasion has survived total amphibolitization. With increasing amphibolitization, the hornblende changes from very pale brown to green. Phlogopite has formed along favored S-surfaces in some of the amphibolitized eclogites. One highly amphibolitized eclogite was found to contain small amounts of staurolite.



(A)



(B)

Fig. 5. (A) Photomicrograph of sugary-textured metaroddingite (163-5-9). Contains garnet (with inclusions), diopside and ilmenite; plane-polarized light. (B) Photomicrograph of metaroddingite (31) with diopside veinlet, plane-polarized light.

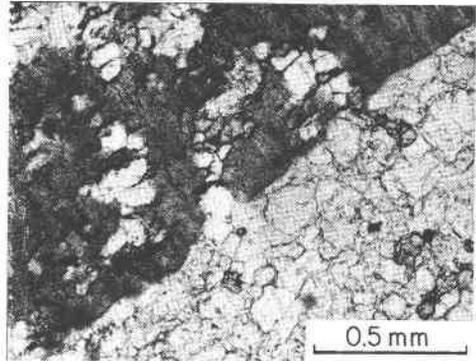


Fig. 6. Photomicrograph of metaroddingite (31-5-3D), showing contact with alteration zone 1, plane-polarized light. Unaltered rock contains garnet and diopside; altered zone contains diopside, texturally unchanged, and hornblende + epidote symplectite after garnet.

Metaroddingite

In thin section a typical rock contains roughly equal amounts of colorless, pale pink, or pale yellow garnet and colorless diopside clinopyroxene, and subordinate quantities of epidote and hornblende. Hornblende occurs instead of garnet in a few more magnesian samples, and in these epidote is also more abundant. Sphene is a fairly consistent accessory, commonly cored by rutile, although rutile may occur without sphene. Ilmenite is present in some samples, green spinel only rarely (*e.g.* 160-7B). Except close to and within the marginal reaction zones, earlier and later assemblages cannot be clearly distinguished. Secondary hornblende, chlorite, ilmenite, vesuvianite, and calcite (Dal Vesco, 1953) do not occur pervasively. The absence of kyanite and the presence of diopside rather than omphacite distinguish these rocks from grosspydites.

Clinopyroxene and epidote occur as 0.1-mm grains in a feebly schistose mosaic (Fig. 5a). Garnet tends to be coarser-grained, irregular in outline, and in some cases poikiloblastic; it does not form spherical porphyroblasts as in the eclogites. Thin veins parallel to or oblique to the foliation are composed of garnet or less commonly clinopyroxene, hornblende, or epidote, all somewhat coarser than in the neighboring rock (Fig. 5b). Hornblende generally occurs interstitially or poikiloblastically in the garnet-bearing metaroddingites.

The sharp boundary against zone 1 of the reaction contact (Fig. 4) is marked by the replacement of garnet (including veins) by hornblende-epidote symplectite (Fig. 6), or less commonly diopside-epidote symplectite. The compositional layering and the clinopyroxene fabric of the metaroddingite continue

right through this zone. With an increase in grain-size outwards, the symplektite pseudomorphs after garnet gradually disappear in favor of a coarse-grained xenoblastic granular hornblende-epidote aggregate (zone 2). The epidotes are noticeably zoned towards more ferric rims. Megacrysts of hornblende, epidote, sphene, and ilmenite occur patchily in the outer parts of zone 2. Passage into zone 3 (chlorite zone) is accompanied by a change towards an idioblastic actinolitic amphibole and the disappearance of epidote. The chlorite zone proper, closest to the ultramafic, contains tremolite. The transition into ultramafic schist is marked by the incoming of olivine and orthopyroxene.

At least zone 1 and probably also zone 2 was originally metarodingitic (*i.e.* garnetiferous, *etc.*). Bulk analyses (W. Richter, unpublished) show that the transition from metarodingite to reaction zone 1 was largely isochemical; it is the front of the hydration reaction: $\text{garnet} + \text{H}_2\text{O} \rightarrow \text{epidote} + \text{hornblende}$. The transition from zone 1 to zone 2 may have involved some mass transfer, but is in part an expression of the reaction: $\text{clinopyroxene} + \text{H}_2\text{O} \rightarrow \text{epidote} + \text{hornblende}$. Mass transfer of the type commonly found at ultramafic contacts (*e.g.* Brady, 1977) was obviously responsible for zone 3.

Bulk chemistry

Analyses of rocks spanning the entire range from eclogite to Ca-rich metarodingite (excluding marginal reaction contacts) are assembled in Table 1 in order of increasing CaO content. The eclogite end-members

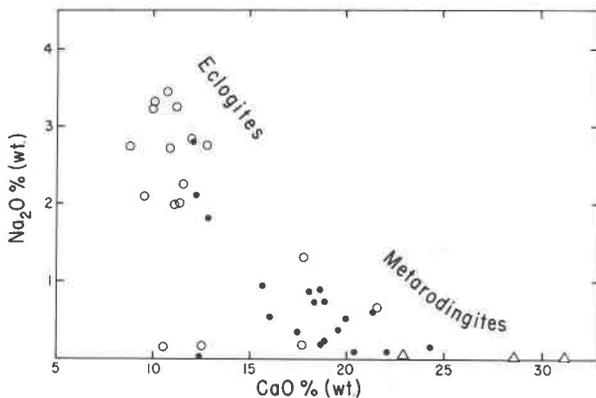


Fig. 7. Inverse correlation between CaO and Na₂O in Ticino eclogites, metarodingites and transitional rocks. Dots: Gagnone samples (Table 1); triangles: samples from Gesero, SE of Bellinzona; circles: samples from other Ticino localities from unpublished data, and Dal Vesco (1953), Fumasoli (1974), and Ernst (1977).

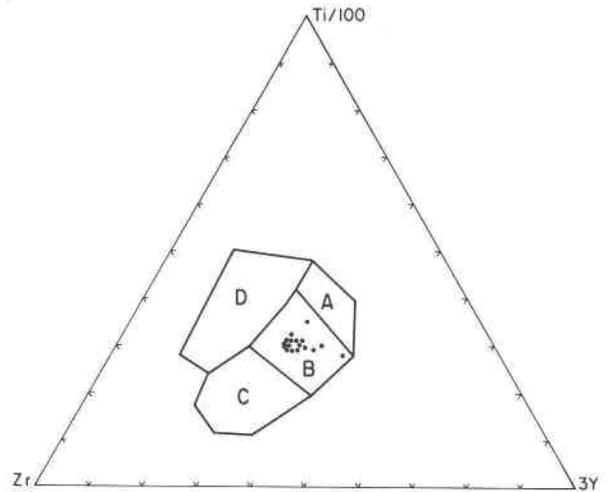


Fig. 8. Ti-Zr-Y diagram after Pearce and Cann (1973). Data from Table 1. (A) field of low-K tholeiites; (B) ocean-floor basalts, low-K tholeiites, and calc-alkali basalts; (C) calc-alkali basalts; (D) ocean island and continental basalts.

(*e.g.*, 163-4-11, 163-4-9, 163-4-8) are basaltic in their chemistry, except for their extremely low K₂O. Transition towards metarodingite has involved principally an increase in CaO coupled with a decrease in Na₂O (Fig. 7), to the point where the bulk chemistry can no longer be matched by any magmatic rock (nor by grosspyrite; Sobolev *et al.*, 1968). Alumina, Cr₂O₃, TiO₂, MgO, total iron, MnO, P₂O₅, and the trace elements Nb, Zr, Y, and Sr do not show any obvious systematic changes in this transition. Silica diminishes slightly. Alumina and MgO vary antipathetically, but the reason for this is not known. The chemical transition from metabasaltic eclogite to metarodingite is reflected in the norm (Table 1) by an increase in Di + An from 55 to 95 percent and a decrease in Ab + Ne from 21 to 0.2 percent. The degree of silica undersaturation in the norm has been somewhat exaggerated by counting all Fe as FeO.

Despite the rocks having undergone rodingitization and eclogite facies and amphibolite facies metamorphism, correlation in the suite among the trace elements Ti, P, Zr, and Y, no doubt reflective of original igneous fractionation, has remained excellent (Figs. 8 and 9). Relative abundances of these elements are the same in eclogites and metarodingites. Their proportions are consistent with the trace element chemistry of ocean-floor basalts, as depicted by Pearce and Cann (1973). Y/Nb and P/Zr ratios suggest tholeiitic affinities (Pearce and Cann, 1973; Winchester and Floyd, 1976). Strontium shows no correlation with CaO and apparently did not increase during rodingitization.

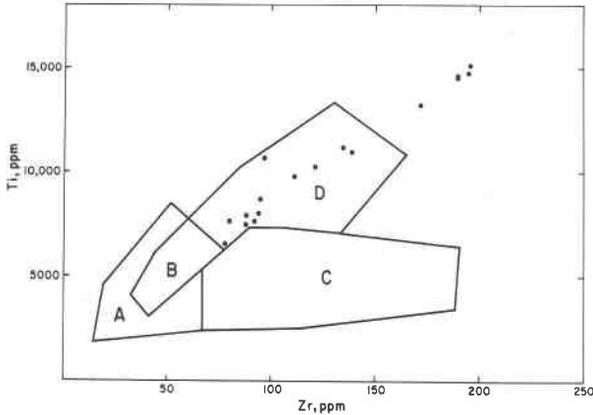


Fig. 9. Plot of Ti against Zr, after Pearce and Cann (1973). Data from Table 1. (A) Field of low-K tholeiites; (B) ocean-floor basalts, low-K tholeiites, and calc-alkali basalts; (C) calc-alkali basalts; (D) ocean-floor basalts. Field D should probably be extended to include more differentiated, higher Ti and Zr rocks.

Mineral chemistry

Microprobe analyses have been made of garnet, clinopyroxene, hornblende, epidote, and sphene. The chemistry of the first three strongly reflect bulk-rock

compositions, namely, whether they are eclogitic (basaltic), rodingitic, or compositionally intermediate.

Garnet

The analyses of cores and rims of garnet in 24 specimens of mafic rock from eight different ultramafic lenses in the Gagnone region (Fig. 2), plotted in Figure 10, have totals within the range 99.0 to 101.0 percent and cation contents relative to 24 oxygens within the limits: Si, 5.93–6.07; Al + Cr + Fe³⁺, 3.95–4.15, Ca + Mg + Mn + Fe²⁺, 5.92–6.02. Ferric iron was calculated so as to make Ca + Mg + Mn + Fe²⁺ = 6/16 of total cations, as required by the garnet formula. The values given for Fe₂O₃ (and corresponding estimates of percent andradite) in the analyses of garnet from five typical Gagnone rocks (Table 2) must be regarded as highly uncertain.

The garnets show an uninterrupted range in composition from 84 percent pyralpsite, 16 percent ugrandite to at least 58 percent ugrandite, mainly but not exclusively as a function of percent CaO in the rock (Fig. 11). Thus, the eclogites contain pyralpsite, whereas the metaroddingites and intermediate rocks

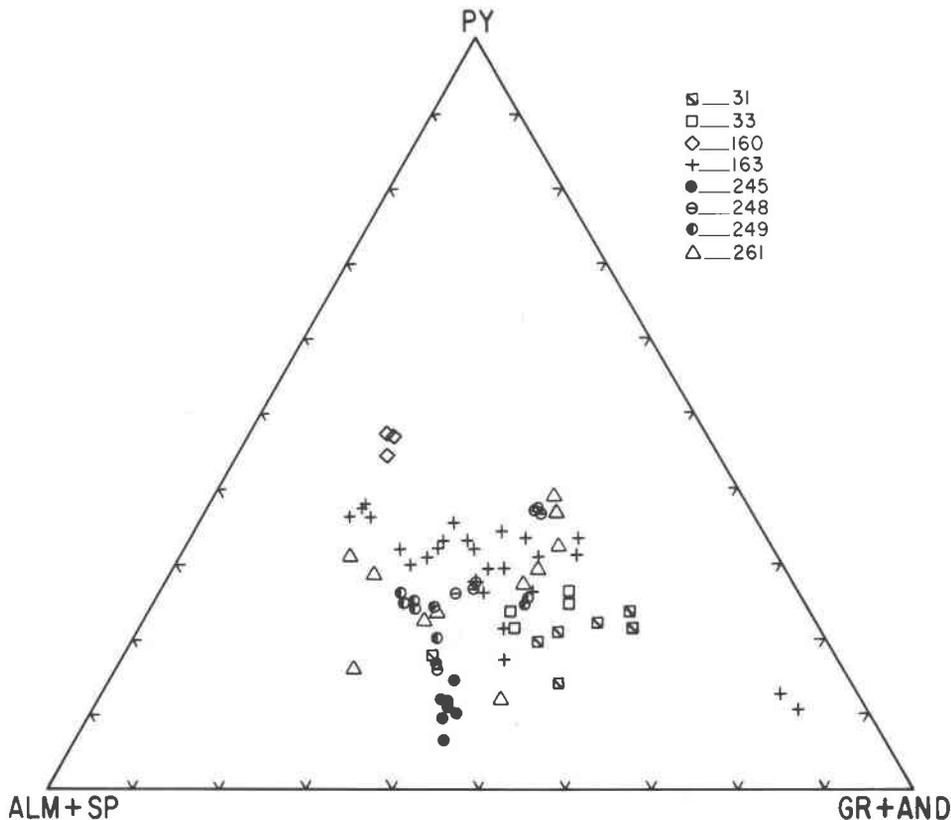


Fig. 10. Core and rim compositions of garnet from Gagnone eclogites, metaroddingites, and transitional rocks. Symbols refer to different ultramafic host bodies (Fig. 2).

Table 2. Microprobe analyses of garnet

	Eclogite 163K		Eclogite 248-6		Transitional 249-J		Metaroddingite 163-P		Metaroddingite 163-M
	C	R	C	R	C	R	C	R	
SiO ₂	39.28	39.53	38.07	39.10	39.26	39.23	39.67	40.06	38.69
TiO ₂	.04	.10	.10	.04	.10	.00	.11	.11	1.12
Al ₂ O ₃	22.59	22.56	22.04	22.38	21.75	21.90	22.21	22.59	18.20
Cr ₂ O ₃	.00	.00	.00	.04	.00	.00	.02	.10	.00
Fe ₂ O ₃	.02	.14	.00	.05	.73	.23	.22	.04	4.12
FeO	19.86	21.47	21.31	17.73	14.28	18.74	14.23	12.89	4.43
MnO	.39	.50	.34	.91	.96	1.30	.23	.19	1.00
MgO	8.52	9.76	4.04	6.89	6.44	6.33	6.99	8.36	3.17
CaO	9.27	6.40	13.34	12.58	15.81	11.93	15.90	15.60	28.66
	99.97	100.46	99.24	99.72	99.33	99.66	99.57	99.94	99.39
Cations relative to 24 oxygens									
Si	5.961	5.965	5.950	5.966	5.992	6.018	6.003	5.990	5.980
Al	4.041	4.013	4.061	4.026	3.913	3.961	3.962	3.982	3.336
Cr	.000	.000	.000	.005	.000	.000	.002	.012	.000
Ti	.005	.011	.012	.005	.011	.000	.013	.012	.131
Fe ³⁺	.002	.016	.000	.006	.084	.027	.025	.005	.479
Fe ²⁺	2.521	2.710	2.785	2.263	1.822	2.405	1.800	1.612	.573
Mn	.050	.064	.045	.118	.124	.169	.029	.024	.130
Mg	1.927	2.195	.941	1.567	1.465	1.447	1.576	1.863	.730
Ca	1.507	1.035	2.234	2.057	2.585	1.961	2.578	2.500	4.746

Fe₂O₃ and FeO based on total iron and ideal garnet formula. C, core composition. R, rim composition.

contain ugrandite or garnet of intermediate composition. The most calcic metaroddingites (e.g. 163M) contain grossular (Fig. 10) with less than 20 percent Py + Alm.

The garnets are almost ternary in composition; percent andradite averages 0.9 and exceeds 3.7 only in the most grossular-rich garnets (>75 percent gross), where calculated amounts reach 12 percent; percent spessartine averages 1.3 and does not exceed 3.8. The relative proportions of pyrope and almandine components vary considerably, although the spread is noticeably smaller within groups of samples from the same ultramafic lens (Fig. 10).

Zoning in the garnets tends to be concentrated in the marginal parts and takes the form of an increase in Fe/Mg ratio towards the rim; Ca zoning is not pronounced and consists, in most cases, of a small decrease towards the rim.

Crystallization temperatures evidently were above the garnet solvus (e.g., Ganguly and Kennedy, 1974). No rocks have been found with two microprobe-

detectable garnet populations related to a grossular-pyralspite solvus, which is apparently asymmetric towards pyralispite (Newton *et al.*, 1977; Cressey, 1978). However, evidence of exsolution in one of the meta-

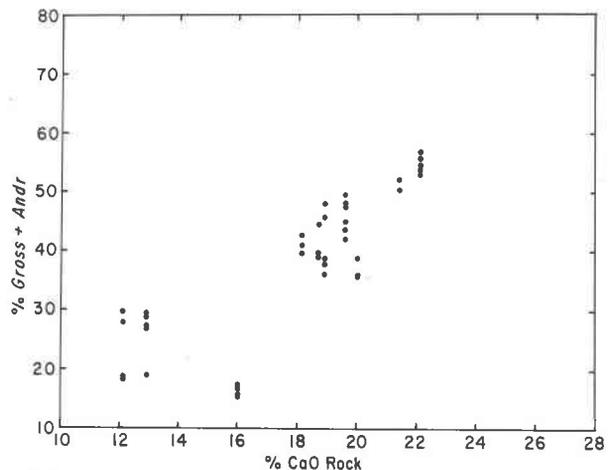


Fig. 11. Correlation between garnet composition and whole rock CaO percent.

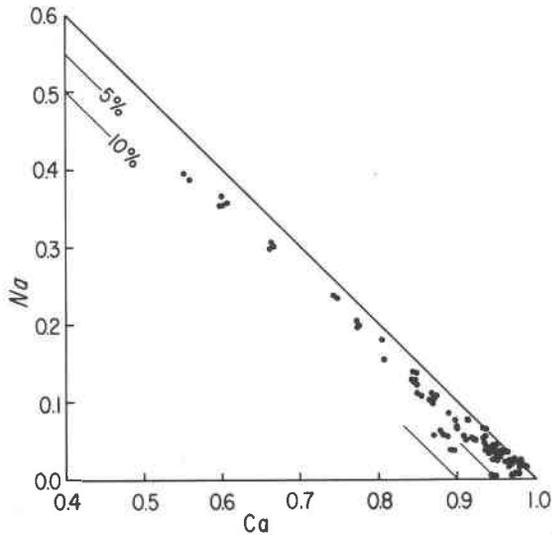


Fig. 12. Na vs. Ca per formula unit in analyzed Gagnone pyroxenes. Percentages refer to content of "orthopyroxene" end-member.

rodingite garnets (sample 33-5-1), with 24 percent pyrope, 24 percent almandine, and 51 percent grossular + andradite, was observed in transmission electron micrographs (Ghose *et al.*, 1976). Garnets intermediate in composition between grossular and pyralspite are, of course, not uncommon in kyanite eclogite and grosspydite xenoliths in kimberlite (*e.g.*, Sobolev *et al.*, 1968; Chinner and Cornell, 1974). Conditions no more extreme than those of the high-greenschist facies or low amphibolite facies are, in fact, capable of permitting compositions midway between grossular and almandine to crystallize (Ackermann *et al.*, 1972; Heritsch, 1973). A solvus is more likely to be encountered, however, as the ratio pyrope/almandine increases (Ganguly and Kennedy, 1974).

Clinopyroxene

A variation in jadeite component of pyroxene from more than 40 percent to almost zero parallels the decrease in bulk Na_2O of the rocks in the passage from eclogite to metarodingite. Cats component ($\text{CaAl}_2\text{SiO}_6$) is uniformly low, with the exception of the few highly calcic metarodingites, which contain diopside zoned marginally to fassaite with more than 20 percent Cats. These relationships are illustrated in Figure 12 in terms of Ca and Na atoms per formula unit of 6 oxygens and in Figure 13 in terms of the proportions of Di + Hed, Jd, and Cats. These diagrams give the results of full microprobe analyses of 67 clinopyroxenes in 26 samples of eclogite/meta-

rodingite from nine different ultramafic lenses. Total cation contents per formula unit ranged from 3.988 to 4.020 (average 4.006), and Si contents, excluding the fassaite pyroxenes (Fig. 13), ranged from 1.931 to 1.993 (average 1.965). Illustrative analyses (Table 3) are of clinopyroxene grains in contact with the garnets previously selected for Table 2.

The calculation of molecular components in the pyroxenes is based on the assumption that all iron is ferrous. Although this is unlikely to be entirely true, the most reasonable procedure for estimating the proportions of ferrous and ferric iron in microprobe analyses of pyroxenes, one in which total cations are equated to 4 and total oxygens to 6 (Ryburn *et al.*, 1976), is so critically dependent on the accuracy of the Si determination that the result is nearly worthless for the calculation of Fe_2O_3 content in low-iron pyroxenes. In our case, this procedure yielded acmite contents ranging up to 6 percent (average 2), and also several negative contents of Fe^{3+} , particularly for the omphacites. Thus, we have preferred to regard all Fe as ferrous, compute first Jd from the number of Na atoms, then TiCats ($\text{CaTiAl}_2\text{O}_6$) from Ti, Cats from the remaining Al, Di + Hed from the remaining Ca, and En + Fs from the remaining Mg + Fe + Mn. This ignores the Si determination. The $(\text{Mg,Fe})_2\text{Si}_2\text{O}_6$ content of the clinopyroxenes so calculated (Fig. 12) averages 3.8 percent and exceeds 5 percent in only one sample (160-4-8). Our calculated Cats components are similar to, but of course not identical to, Cats calculated from four-coordinated Al (*i.e.* 2.0 - Si), *e.g.*, Lovering and White (1969).

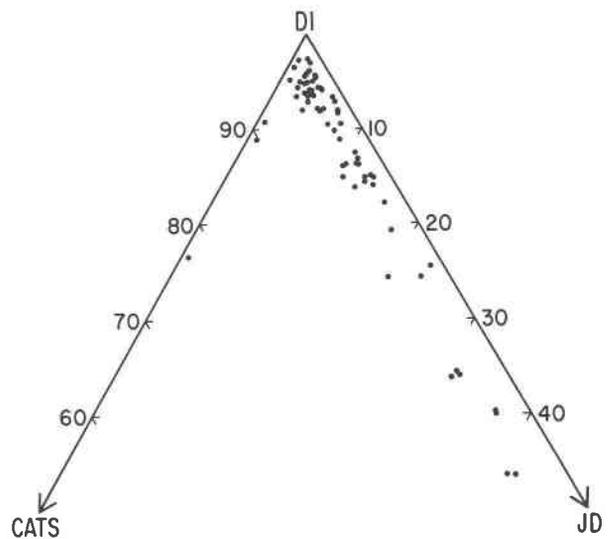


Fig. 13. Molecular proportions of diopside, jadeite, and Cats components in analyzed Gagnone pyroxenes.

Table 3. Microprobe analyses of clinopyroxene

	Eclogite 163K	Eclogite 248-6	Transitional 249-J	Metarodingite 163-P	Metarodingite 163-M	
					C	R
SiO ₂	55.48	53.27	54.24	53.82	51.31	45.90
TiO ₂	.16	.17	.10	.13	.43	2.17
Al ₂ O ₃	11.41	7.06	3.10	1.95	4.36	10.01
Cr ₂ O ₃	.06	.01	.02	.00	n.d.	n.d.
FeO	3.45	3.71	4.58	2.18	2.88	4.60
MnO	.01	.03	.13	.00	.04	.11
MgO	9.74	12.66	14.11	16.94	15.48	12.54
CaO	14.89	19.93	22.72	24.84	25.29	24.49
Na ₂ O	5.70	2.78	1.17	.34	.04	.05
K ₂ O	0.00	.00	.00	.00	n.d.	n.d.
	100.90	99.62	100.17	100.20	99.84	99.88
Cations relative to 6 oxygens						
Si	1.954	1.933	1.977	1.954	1.881	1.703
Al ^{IV}	.046	.067	.023	.046	.119	.297
Al ^{VI}	.428	.235	.110	.037	.070	.141
Cr	.002	.000	.001	.000		
Ti	.004	.005	.003	.004	.012	.061
Fe	.102	.112	.140	.066	.088	.143
Mn	.000	.001	.004	.000	.001	.003
Mg	.510	.685	.767	.916	.846	.694
Ca	.562	.775	.888	.966	.994	.973
Na	.390	.196	.083	.024	.003	.004
X	.952	.971	.971	.990	.997	.977
Y	1.046	1.038	1.025	1.023	1.017	1.042

Minor elements exhibit no obvious dependency on the relative proportions of jadeite and diopside component. TiO₂ averages 0.13 weight percent, Cr₂O₃ 0.02 percent, MnO 0.03 percent, whereas K₂O was not detected (<0.01 percent). The rare fassaite clinopyroxenes have, like their associated garnet, considerably higher TiO₂ (Table 3).

The metamorphic histories of the eclogites and metarodingites are believed to have been the same, namely an eclogite facies metamorphism followed by an amphibolite facies metamorphism, the effects of the latter being more noticeable at the margins of the mafic bodies. The occurrence at Gagnone of high Al^{VI} omphacitic pyroxene and high Al^{IV} tschermakitic pyroxene (fassaite) is a consequence of the range in bulk compositions of the mafic rocks. Their conventional correlation with high and low *P/T* environments respectively (White, 1964) is of course only valid in appropriate multiphase assemblages where

heterogeneous equilibria controlling the Al^{IV}/Al^{VI} ratio of pyroxene can operate. The fact that both jadeitic and tschermakitic pyroxenes can occur in high-pressure metamorphic facies is not in conflict with experimental data on their intrinsic stabilities (Hays, 1966; Kushiro, 1969).

Amphibole

Amphiboles from Gagnone eclogites and amphibolitized eclogites are complex hornblendes in which the edenite component becomes more significant as the associated clinopyroxene becomes more jadeitic (Fig. 14). Amphiboles from the metarodingites (<10 percent Jd in Cpx) show a trend from tremolitic hornblende to pargasitic hornblende (Fig. 14). In the eclogites, the amphibole mostly appears to be in textural equilibrium with pyroxene and garnet; on the other hand, most amphiboles do not form a part of the primary metarodingite texture.

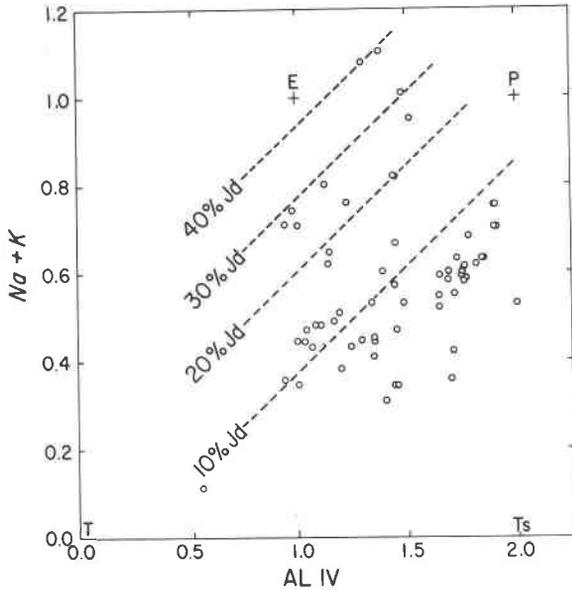


Fig. 14. Na + K vs. Al^{IV} in Gagnone amphiboles. Contours represent percentage of jadeite component in coexisting pyroxene. T, tremolite; Ts, tschermakite; P, pargasite; E, edenite.

The amphibole chemistry was determined by microprobe analysis of 25 samples from 8 ultramafic lenses. Those selected for Table 4 are of hornblendes in contact or near contact with the garnets and pyroxenes in Tables 2 and 3. Recast on an anhydrous basis of 23 oxygens, the range in cation contents was: Ca, 1.32 to 1.9 (eclogites), 1.9 to 2.03 (metaroddingites); Na + K, 0.34 to 1.1 (eclogites), 0.11 to 0.69 (metaroddingites); Al^{IV}, 0.37 to 2.00; and Al^{VI}, 0.28 to 1.28. Iron was taken as entirely ferrous in these calculations. Mg/(Mg + Fe), with one exception, fell between 0.83 and 0.63. The average content of minor elements is distinctly higher than in the clinopyroxenes, *viz.* TiO₂ 0.45 percent, Cr₂O₃ 0.04 percent, MnO 0.13 percent, and K₂O 0.26 percent.

Epidote

The eclogites contain two varieties of epidote: (1) a coarse-grained idioblastic elongate clinozoisite with 4 to 5 percent Ca₂Fe₃Si₃O₁₂(OH) endmember and (2) a xenoblastic epidote with 13 to 17 percent of this component. The former variety often defines a schistosity and appears to be in textural equilibrium with pyroxene, garnet, and amphibole. The latter appears to postdate the garnet and pyroxene. Some samples of eclogite contain both varieties of epidote.

Epidote in metaroddingites seems mostly of secondary origin. Its composition ranges from 13 to 24 percent Ca₂Fe₃Si₃O₁₂(OH). These variations in epi-

dote composition presumably reflect in part the Fe₂O₃/FeO ratio of the rock. There is no correlation between epidote composition and the principal variations in pyroxene, garnet, and amphibole noted earlier.

Sphene

Analyses showed substitutions of Fe³⁺ + Al for Ti from 4.5 to 10 percent. The nature of the coupled substitution was not determined.

Iron-magnesium partitioning

Because the Gagnone eclogites and metaroddingites are essentially two-phase (garnet + clinopyroxene) rocks, and the paragenetic relations of hornblende and epidote minerals are not entirely clear, we effectively have high-variance assemblages and consequently little opportunity to derive *P, T* information from considerations of heterogeneous equilibria. On the other hand, as is well known, the fractionation of Fe and Mg among silicate minerals at metamorphic temperatures, particularly when garnet is involved, is relatively sensitive to temperature.

Garnet-pyroxene

The rim compositions of coexisting garnet and clinopyroxene from Gagnone eclogites and metarod-

Table 4. Microprobe analyses of amphibole

	Eclogite 163K	Eclogite 248-6	Transitional 249-J		Roddingite 163-P
			C	R	
SiO ₂	46.27	47.89	48.26	54.13	42.64
TiO ₂	.49	.36	.33	.13	.53
Al ₂ O ₃	15.76	11.22	9.64	3.90	14.41
Cr ₂ O ₃	.05	.07	.03	.03	.00
FeO	8.15	8.54	9.12	7.18	10.19
MnO	.04	.09	.14	.12	.13
MgO	14.76	15.33	15.03	18.56	14.19
CaO	8.91	11.19	12.51	12.72	12.79
Na ₂ O	3.50	2.08	1.10	.40	2.09
K ₂ O	.30	.21	.19	.03	.10
	98.23	96.98	96.35	97.20	97.07
Cations relative to 23 oxygens (anhydrous)					
Si	6.521	6.865	6.989	7.629	6.242
Al ^{IV}	1.479	1.135	1.011	.371	1.758
Al ^{VI}	1.140	.761	.634	.277	.727
Cr	.005	.008	.003	.003	.000
Ti	.052	.039	.036	.014	.058
Fe	.960	1.023	1.105	.847	1.247
Mn	.004	.011	.016	.015	.016
Mg	3.100	3.275	3.244	3.898	3.097
Ca	1.346	1.719	1.941	1.920	2.006
Na	.959	.577	.309	.112	.592
K	.055	.037	.034	.005	.018
X	2.360	2.333	2.284	2.037	2.616
Y	5.261	5.117	5.038	5.054	5.145

ingites are plotted logarithmically in Figure 15. The values of K_D (Fe/Mg Gt)/(Fe/Mg Cpx) range from 5.5 to 15. Inclusion of analyses of garnet and pyroxene cores would extend the spread to K_D values as high as 30. This could mean that core pairs record lower temperatures or higher pressures of equilibration than rim pairs (Banno, 1970) or, more likely in our opinion, that they do not record a reliable exchange equilibrium. In general, the clinopyroxenes are more homogeneous than the garnets, and it is possible that, unlike the garnet interiors, the pyroxene interiors participated in the exchange equilibrium that the rim pairs record. The enrichment of Mg in the garnet rims would be consistent with this hypothesis.

Aside from real variations in physical conditions, the spread in values of K_D (Fig. 15) may be attributed to a combination of the following: imperfect exchange equilibration between the points analyzed, analytical error, variability in Fe^{3+}/Fe^{2+} ratio particularly in pyroxene, and the influence of other components in the two phases, notably grossular and jadeite. The correlations between K_D for rim pyroxene-garnet pairs and these additional components are illustrated in Figures 16 and 17. The positive correlation between K_D and percent grossular (*cf.* Mysen and Heier, 1972) parallels the effect already well known for natural garnet-biotite and garnet-hornblende pairs (*e.g.* Ganguly and Kennedy, 1974, and others), and ascribed to non-ideality in grossu-

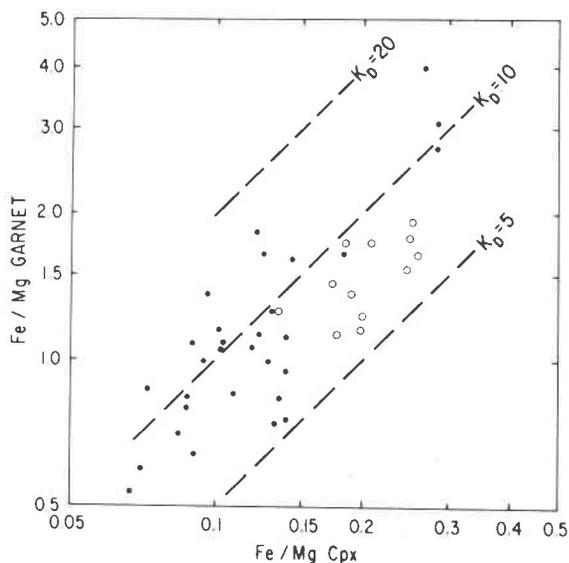


Fig. 15. Partitioning of total Fe and Mg between clinopyroxene and garnet rims. Circles, eclogites; dots, metarodingites.

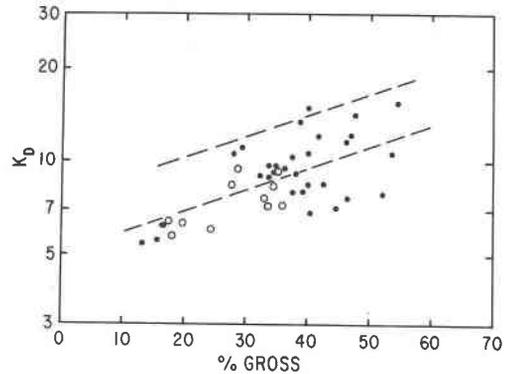


Fig. 16. Correlation between K_D for Fe/Mg partitioning between clinopyroxene and garnet rims and percent grossular component in garnet. Circles, eclogites; dots, metarodingites. Arbitrary dashed lines indicate the trend of change in K_D at $T = 1000K$ due to variation in percent gross alone, according to Ganguly (1978).

lar-pyralspite solid solutions (*e.g.* Hensen *et al.*, 1975). The trend lines on Figure 16 express, for $T = 1000K$, the dependence of K_D on percent grossular alone, according to a relation for $K_D = f(T, P, X)$ developed by Ganguly (1978), based on tabulated thermodynamic data and mixing models for natural garnets. The negative correlation between K_D and percent jadeite (Fig. 17) is probably not so much due to non-ideality of diopside-jadeite solution as to the inverse correlation of grossular and jadeite percent (due to the inverse CaO *vs.* Na₂O relation in the bulk rock, Fig. 7) combined with the correlation between K_D and percent grossular (Fig. 16). This explanation is supported by sample 160-4-7, which, despite the low jadeite content of its pyroxene, contains grossular-poor garnet and hence has a K_D near 6 (Fig. 17). Clearly, then, attempts to derive estimates of physical

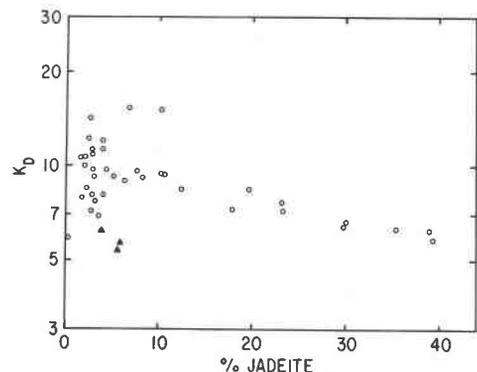


Fig. 17. Correlation between K_D for Fe/Mg partitioning between clinopyroxene and garnet rims and percent jadeite component in pyroxene. Triangles, specimen 160-4-7 which has garnet with only 16-17 percent ugrandite.

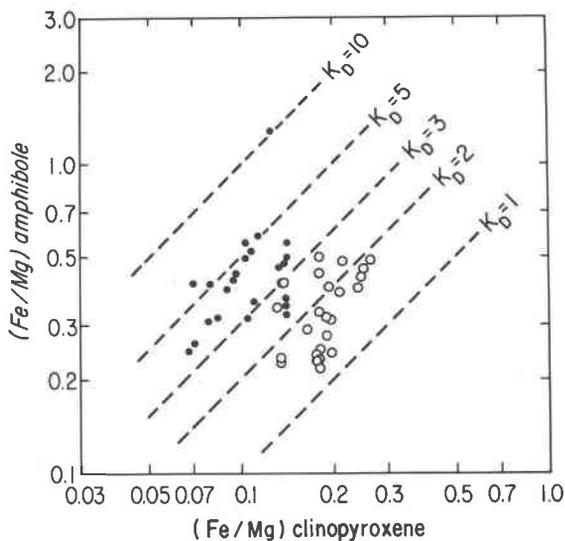


Fig. 18. Partitioning of Mg and total Fe between coexisting clinopyroxene and amphibole. Circles, eclogites; dots, metaroddingites.

conditions from the $K_D(\text{Mg-Fe})$ of pyroxene-garnet pairs cannot afford to ignore certain phase compositional effects.

Iron in Figures 15, 16, and 17 is total Fe. Except when percent grossular exceeds 60, the relative amount of Fe^{3+} in garnet is small, whereas in pyroxene it is possible that Fe^{3+} constitutes a sizeable part of total iron. Thus, it is probable that the ratio $(\text{Fe}^{2+}/\text{Mg Gt})/(\text{Fe}^{2+}/\text{Mg Cpx})$ is somewhat larger than the plotted K_D values. This uncertainty can have a significant effect on estimated temperatures.

For Gagnone eclogites, K_D values (total Fe) range from 5 to 10 (Figs. 15 and 16) and thereby include the values (≈ 6) found for eclogites from Alpe Arami (Ernst, 1977). Within the garnet compositional range of 15 to 25 percent grossular, typical of the product garnets in the experiments on tholeiitic bulk compositions by Råheim and Green (1974), Gagnone eclogites have K_D values of 6. According to Råheim and Green, this partitioning would correspond to 690°C at 10 kbar (an approximate lower P limit for eclogite, Green and Ringwood, 1967) and 793°C at 25 kbar (an approximate upper P limit for hornblende eclogite at 800°C , Lambert and Wyllie, 1972). The latter figure would be in excellent agreement with estimates of P and T made for the garnet peridotite (outcrop 160) at Gagnone (Evans and Trommsdorff, 1978). However, these temperatures might be somewhat high because of the $\text{Fe}^{3+}/\text{Fe}^{2+}$ uncertainty. On the other hand, the dP/dT slope of the experimental calibration includes the effect of increasing grossular

content with increasing pressure (at the expense of pyroxene) and may be too flat (Wood, 1976); hence the estimates of T at lower assumed pressure may be too low.

For the Gagnone metaroddingites, it is appropriate to use the relation: $T^\circ\text{K}(\ln K_D + 2.939) = 4801 + 11.07P(\text{kbar}) + 1586X_{\text{Ca}}^{\text{Gt}} + 1308X_{\text{Mn}}^{\text{Gt}}$ for low-Na pyroxene, developed by Ganguly (manuscript). For an average metaroddingite ($X_{\text{Ca}}^{\text{Gt}} = 0.4$, $K_D = 10$, see Fig. 16), Ganguly's expression gives 785°C for $P = 10$ kbar, or 806°C for $P = 20$ kbar, or 817°C for 25 kbar.

What is important is that, despite their different K_D values, the available thermobarometric data indicate similar equilibration conditions for the eclogites and the metaroddingites (and, of course, the transitional rocks). This is consistent with the geological evidence for an eclogite facies event affecting all the mafic inclusions in the Gagnone region.

Having no accurate fix on pressure, we can only infer a temperature minimum (690°C) for this metamorphism. Nevertheless, this indicates negligible overlap with conditions estimated for the later Central Alpine regional metamorphism in the Gagnone region ($600 \pm 100^\circ\text{C}$, < 10 kbar; Evans and Trommsdorff, 1974). The Central Alpine metamorphism evidently caused little or no reequilibration of Fe and Mg between pyroxene and garnet. Temperature and pressure estimates for Gagnone eclogites, metaroddingites, and garnet peridotite converge on the rounded-off values of $T = 800^\circ\text{C}$ and $P \approx 25$ kbar.

Pyroxene-amphibole

Compositional factors can also be recognized in the partitioning of Fe and Mg between pyroxene and coexisting amphibole (Fig. 18). Although the K_D correlates with jadeite content, this in turn correlates with several compositional variables in the amphibole, as we have seen earlier (Fig. 14). It is therefore not so easy from the present data to identify the principal compositional variable influencing the K_D for pyroxene-amphibole pairs. Clearly, for temperature comparisons among different areas, such data should be used with caution. Furthermore, the nature of the event being measured is not entirely clear. For Gagnone eclogites the K_D values fall in the range 1.2 to 3.0, and for Gagnone metaroddingites the range is 2.3 to 10. Again, the eclogite pairs cover the range found for pyroxene-amphibole pairs in Alpe Arami eclogites (Ernst, 1977). Scatter due to the unknown

proportions of Fe^{3+} in both pyroxene and amphibole has, of course, not been considered.

Origin of metaroddingites

On the basis of our analytical data, the eclogites must represent metabasaltic extrusives or intrusives. On the other hand, it is appropriate at this point to list the evidence and lines of reasoning for believing that the garnet–diopside (*i.e.* calc–silicate) boudins are metaroddingites, the metamorphosed products of rodingitized dikes. Alternate possibilities are that they are metamorphosed calcareous sediment or metamorphosed garnet or plagioclase pyroxenite dikes or layers in the ultramafic rocks.

(1) The metaroddingites are everywhere in contact with, in fact usually totally enclosed by, the metamorphosed ultramafic rock. Mafic layers and boudins in the surrounding semi-pelitic gneisses are amphibolites, amphibolitized eclogites, or eclogites, but nowhere metaroddingites. These observations make a sedimentary protolith unlikely.

(2) Albeit rarely, metaroddingite appears to cross-cut, at low angles due to tectonic flattening, the compositional layering in the ultramafic rock. An intrusive origin is therefore indicated.

(3) Their basic mineralogy, structures, and physical appearance in the field resemble the rodingite dikes enclosed in serpentinite in lower-grade sections of the upper Pennine nappes (*e.g.*, Peters, 1963; Dal Piaz, 1967; Dietrich, 1969; Keusen, 1972). Essentially comparable are details such as the size of the rock bodies, the boudinage structure, absence of pronounced compositional layering, presence of thin cross-cutting garnet and pyroxene veins, and the presence of a marginal chloritic contact reaction zone.

(4) The bulk chemistry of the metaroddingites shows complete gradation into the basaltic chemistry of the eclogites. Such a gradation would be unlikely if the protolith were sedimentary. It would, on the other hand, be the predictable result of variable degrees of metasomatic rodingitization of basaltic dikes. Enrichment in CaO and eventual nearly complete depletion in alkalis are the principal changes recognized in studies of rodingitized dikes enclosed in serpentinite (Coleman, 1967, p. 49; Keusen, 1972, Leach and Rodgers, 1978).

(5) The abundances of minor (*e.g.*, TiO_2 :1.24 to 2.46 percent) and trace elements (Table 1) throughout the gradational sequence eclogite–metaroddingite are consistent with basaltic chemistry. Indeed, certain abundance ratios show remarkable regularities (Fig.

8) and suggest correlation with tholeiitic sea-floor basalts. The concentrations of Cr, Zr, and Sr distinguish these rocks from the accompanying ultramafics and make it improbable that they represent boudinaged pyroxenite layers.

(6) The enclosing ultramafic rocks are chlorite–enstatite \pm cummingtonite–olivine schists, some of which are totally devoid of or possess only very minor amounts of tremolite, the only phase in the ultramafics carrying more than 0.6 percent CaO at this metamorphic grade (Evans and Trommsdorff, 1974; Rice *et al.*, 1974). Thus, some of the ultramafics are CaO-depleted, although not Al_2O_3 -depleted, relative to a lherzolite protolith (Trommsdorff and Evans, 1969; Stäuble, 1978). This CaO depletion could well have taken place during serpentinization accompanying rodingitization (Page, 1967; Barnes and O'Neil, 1969; Coleman, 1977).

(7) In areas of lower grades of metamorphism in the Pennine Alps, a serpentinite origin for some of the olivine-rich ultramafic rocks was established through the preservation of a microfolded antigorite–magnetite fabric inside olivine porphyroblasts (*e.g.* Trommsdorff and Evans, 1974, Plate 1c). In the Gagnone area, traces of a possible serpentinite-stage fabric appear to have been totally removed.

(8) Relics of titanian clinohumite and the presence of a characteristic breakdown intergrowth of olivine + ilmenite in ultramafic body #160 (Fig. 2) may, by analogy with titanian clinohumite-bearing Penninic serpentinites, be a further indication of a precursor serpentinite stage for the Gagnone ultramafics (Evans and Trommsdorff, 1978).

Geologic history and significance of the eclogite–metaroddingite suite

The formation of the metaroddingites at least predated the growth of the postkinematic hornblende, diopside, epidote, and chlorite-bearing reaction zones. Similarly, eclogite formation predated an episode of amphibolitization which can be attributed to the Central Alpine metamorphism. The fact that the mineralogy and textures of the metaroddingites show smooth gradations into those of eclogite, and that these variations can be ascribed entirely to compositional variations, suggests that the entire suite shared an eclogite facies metamorphism sometime during their geological history. Geothermometry has confirmed this fact. It is still not known, however, if this high-pressure metamorphism was early Alpine or pre-Mesozoic. It probably correlates with the formation of garnet peridotite at Gagnone (Evans and

Trommsdorff, 1978), Alpe Arami (Dal Vesco, 1953; Möckel, 1969; Ernst, 1977) and perhaps Monte Duria (Fumasoli, 1974). The temperature of the eclogite facies metamorphism, as consistently recorded by element partitioning in both eclogite and metarodingite at Gagnone, was higher than that of the Central Alpine phase, and considerably higher than that of the possibly correlative early Alpine high-pressure events in other parts of the Alps (Bearth, 1974; Ernst, 1973; Compagnoni, 1977).

If the hypothesis of partial metasomatic rodingitization is accepted for the Gagnone suite, the necessary accompanying event of serpentinization must have taken place prior to eclogite facies metamorphism. The involvement of oceanic rocks is suggested by their trace-element chemistry. The early geologic history of the Gagnone mafic/ultramafic suite therefore appears to have involved subduction metamorphism of partially serpentinized oceanic lithosphere. This conclusion does not have to apply, of course, to all rocks in the Cima Lunga Series, for example, the carbonate and pelitic metasediments. Except for one¹ ultramafic body at Cima di Gagnone (no. 160, Fig. 2), traces of these early high-pressure events within the ultramafics have been wiped out by thorough recrystallization during the Central Alpine metamorphism.

Although we believe the balance of evidence favors an early rodingitization event, the possibility of mass transfer of CaO and alkalis during the various metamorphisms cannot entirely be dismissed. For example, in some of the ultramafics we find clear textural evidence of the replacement of tremolite by magnesioicummingtonite (Rice *et al.*, 1974), with no concomitant crystallization of a Ca-bearing phase elsewhere in the rock. Diffusion on the scale of centimeters involving several components, perhaps earlier but certainly also during the Central Alpine metamorphism, was clearly responsible for the high-variance contact reaction zones developed between the ultramafic and mafic rocks. The size of some of the metarodingite bodies (*e.g.* > 2 m across), however, argues in favor of an infiltrational process of mass transfer rather than a diffusional one (Fletcher and Hofmann, 1974). Thus, a low-temperature, near-surface environment for the formation of these rocks seems very reasonable.

The recognition of calc-silicate rock associated with metamorphic ultramafic rock as metamorphosed rodingite supplies a piece of information of

considerable value in reconstructing the geological history of any crystalline terrane. The association is indicative of passage through a pressure-temperature range in which serpentinization was possible. The subsequent metamorphic history may be easy to reconstruct, as in the case of contact metamorphism (*e.g.*, Frost, 1975), or rather subtle, as in the present case of subduction zone, eclogite facies metamorphism, followed by upward migration of fragmented material during alpine tectonism, and subsequent Barrovian-style regional metamorphism.

Acknowledgments

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¹ Note added in proof: Two additional bodies show evidence of having formerly been garnet peridotite.

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