

The Mineralogy of the Glaucophane Schists and Associated Rocks from Île de Groix, Brittany, France

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Abstract. Ninety-seven mineral phases consisting of ten chloritoids, fifteen epidotes, sixteen garnets, four sphenes, seven rutiles, seven pyroxenes, thirteen blue amphiboles, two green amphiboles, eleven phengites, two paragonites, a mariposite, seven chlorites, and two specimens of albite were obtained from the metamorphic rocks of Île de Groix, and their chemical, physical, optical and X-ray properties determined. The chloritoids are all optically positive, monoclinic polymorphs with large 2V, moderate refractive indices and characterized by high densities. Their fluorine contents have been used to propose a new upper limit for OH → F substitution in the chloritoid structure, suggesting that partial pressure of fluorine might modify the stability of chloritoids from that determined in pure H₂O. The epidotes belong to the Al-Fe epidote series and are "epidote" sensu stricto. The almandine-rich garnets and the chloromelanites are metastable relics in the glaucophane schists. The grossular contents of the calcareous schist garnets are believed to have become depressed under high CO₂ pressure and the low Tschermak's contents of the pyroxenes are to be explained by equilibria involving epidote at high P_{H₂O} and low temperature when the Tschermak's components will break down to epidote group minerals. The sphenes contain appreciable amounts of combined water, fluorine substituting for oxygen and aluminium substituting for silicon and titanium. The presence of H₃O⁺ is suspected in a specimen of blue amphibole. The barroisite has a composition between glaucophane and hornblende. On account of its high Fe³⁺ content it is believed to have formed under higher P_{O₂} than the blue amphiboles. The paragonites which occur in the chloritoid veins are unstable in the potassium-rich aluminous schists. The phengites show a tendency towards sericitic composition due to post-glaucophanisation readjustments under the lower pressure conditions of the greenschist facies. Some of the Fe³⁺ contents of the chlorites are interpreted as due to oxidation of ferrous iron, e.g. 2[Fe(OH)₂] → 2FeOOH + H₂. The minerals show strong chemical control of the host rock and their Mn contents are directly related to those of the minerals from which they have evolved through retrogression.

Chloritoids and epidotes that are not associated with garnets contain higher amounts of manganese; similarly, the two blue amphiboles with the highest Fe:Mg ratios were obtained from rocks in which garnet has not appeared. It is therefore believed that ottrelite and piemontite would be stable only at the lowest subfacies of the greenschist facies. Also, the iron-rich amphiboles must have evolved from low-grade iron-aluminium chlorites, since on the appearance of garnet in a schist iron-aluminium chlorites react with quartz to give almandine and Mg-rich chlorites. The Fe²⁺:Mg ratios of the blue amphiboles therefore reflect the grade of the original schist in which the minerals formed.

Introduction

Materials Analysed and Mode of Presentation. The modal group average of the mineral composition of the glaucophane schists of Île de Groix is given in Table 1. Ninety-seven mineral phases belonging to ten mineral species were obtained from these schists. Petrographic evidence (Makanjuola, 1967) revealed that the garnets,

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Table 1. The modal group averages of the mineral composition of the glaucophane schists of Île de Groix

	Aluminous schists	Calcareous schists	Basic schists ^a	Magnesian schists
Phengite	48.0	5.5	0.9	Mineralogy consists essentially of talc, actinolite and chlorite with accessory magnetite and mariposite
Quartz	24.0	10.1	5.4	
Chloritoid	6.3	—	0.1	
Plagioclase	5.4	5.1	—	
Chlorite	5.0	5.3	2.4	
Glaucophane	2.9	12.2	43.7 ^b	
Pyroxene	—	—	12.4	
Rutile/ilmenite	2.6	1.7	0.9	
Sphene/rutile	—	—	4.5	
Sphene	0.4	0.2	—	
Epidote	2.3	52.0	16.3	
Calcite	—	2.3	—	
Garnet	2.1	4.6	13.2	
Magnetite	0.5	0.9	—	
Apatite	0.2	trace	trace	
Tourmaline	0.1	trace	—	
Biotite	0.1	—	—	
Hematite	0.1	0.1	0.1	
Pyrite	—	—	0.1	
Total	100.0	100.0	100.0	
No. averaged	30	7	15	

^a Average is for the pyroxene-bearing varieties of the basic schists, hence the absence of plagioclase and calcite in the mineralogy.

^b Includes value for barroisite.

pyroxenes and sphenes, especially of the basic schists, predated glaucophane evolution. These minerals occur in some specimens of the glaucophane schist and therefore must have readjusted their compositions under the glaucophane schist facies set of physical conditions. The chloritoids, epidotes, rutiles and phengites must have co-evolved with glaucophane, while the plagioclase, paragonite, barroisite and secondary chlorites post-date glaucophane. The distribution of the minerals studied among the rock groups of the island is shown in Table 2. Their physical, optical, chemical and X-ray properties, provided data for the detailed characterisation of the minerals. For ease of presentation these ten mineral species have been grouped into three, namely, the orthosilicates and the oxides, the chain silicates, and the sheet and framework silicates. A final chapter has been devoted to the study of element partitioning among coexisting mineral phases, the influence of the host rock composition on the mineral chemistry and a reconstruction of the mineral paragenetic succession.

Previous Work. Publications on mineralogical and geochemical studies of the crystalline schists and associated rocks of Île de Groix have been rather few. Barrois (1884) quoted the analysis of a vein chloritoid by Renard; Lacroix (1893) cited two incomplete analyses of glaucophane, which were later quoted by Cogné

Table 2. Distribution of the analysed minerals among the rock groups^a

	Group I	Group II	Group III	Group IV	Group V
Actinolite	—	—	1	—	—
Albite	—	1	—	1	—
Barroisite	—	—	—	1	—
Chlorite	1	—	1	1	4
Chloritoid	8	—	—	—	2
Epidote	1	4	—	10	—
Garnet	4	3	—	9	—
Glaucofane	1	2	—	10	—
Mariposite	—	—	1	—	—
Paragonite	—	—	—	—	2
Phengite	7	1	—	3	—
Pyroxene	—	—	—	7	—
Rutile	—	—	—	7	—
Sphene	—	—	—	4	—

^a Group I: the aluminous schists; Group II: the calcareous schists; Group III: the magnesian schists; Group IV: the basic schists. Group V: the vein and pegmatitic bodies. These group names are the same as those previously established (Makanjuola, 1967).

(1960); Brindley and Harrison (1952) determined the cell dimensions of a chloritoid from the island, which they later refined (Harrison and Brindley, 1957). Velde (1967a) investigated the Si⁴⁺ content of phengites and also (Velde, 1967b) reported chemical analyses of a chloritoid and a glaucophane. Velde's (1967a) work was the first publication dealing mainly with the geochemical aspects of the rocks of the island. A petrographic study of the glaucophane schists and prasinites has been reported by Felix (1970).

Scope of the Present Work. The present work provides data on the chemical, physical, optical and X-ray properties of the minerals of the crystalline schists of the island. Detailed analyses of the collected data have resulted in the determination of the interdependencies of the properties of the mineral species, calculation of ionic substitution ranges within their lattice structures, the partitioning of elements among coexisting mineral phases and a reconstruction of the mineral paragenetic succession. All these will contribute to a better understanding of the complex geological history of the polymetamorphic rocks of Île de Groix, and it is hoped that they will also stimulate discussions that will lead to further investigations.

Sampling Methods. Only fresh rock specimens were collected from outcrops that were *in situ*. Selections were mainly subjective since many of the samples were collected because they showed promise of mineralogical and, or, geochemical interest.

Purity of the Analysed Materials. Representative grains from all the analysed specimens were mounted in liquids that matched their refractive indices, and examined in transmitted light. They were all over 99 percent free of admixed impurities and inclusions. Although many of the thicker chloritoid grains were almost opaque, their thin edges transmitted sufficient light to show that they were fresh and free from inclusions.

Analytical Methods and Precision. The specific gravity of individual minerals was determined with a pycnometer, using a modification of the method described by Holmes (1921). Precision was about ± 0.005 .

Conventional classical methods for silicate analysis, as described by Washington (1930), have been closely followed, but with modifications wherever and whenever necessary. The concentrations of the alkalis were determined using a flame photometer, and those of titania, manganous oxide, total iron and phosphorus were determined spectrophotometrically. Ferrous iron and fluorine were determined by volumetric methods.

The X-ray diffraction pattern for all the minerals was obtained on a Siemens Kristalloflex horizontal diffractometer with scintillation counter detector and pulse height selection: both copper and cobalt targets with appropriate filters were used. Each specimen was scanned twice in order to check reproducibility of the diffraction pattern and thereby the dependability of the data obtained from them. Both press and smear mounts were used. The press slides were employed in the case of platy minerals in order to correct their tendency for preferred orientation. Silicon (chemical standard), quartz (purified α -quartz) and potassium bromate were used as internal standards for different minerals. Peak positions of diffraction charts were estimated at half peak heights.

The rutiles were the only minerals analysed by means of X-ray spectrography. Sample and standard pellets were made from 0.5 g of material mixed with cellulose binder and scanned using tungsten radiation. The standard pellets were prepared from titanium oxide (analar) and spepure iron.

The optical properties of the minerals were determined on the petrographic microscope equipped with either a three-axis or a four-axis universal stage as found convenient. The optic axial angles were either measured directly or indirectly by using Berek's (1923) procedure. Most of the optical determinations were carried out with a sodium lamp. The refractive indices were determined by the single (temperature) variation method, and the values were accurate to ± 0.001 . Values up to 1.74 were read directly on the Abbe, and the higher values on the Leitz refractometers. The refractive indices and data correlate well with $(\text{Fe}^{3+} + \text{Fe}^{2+} + \text{Ti} + \text{Mn})$ in the blue amphiboles, chlorites, and garnets, with $\text{Mg}/(\text{Mg} + \text{Fe}^{3+} + \text{Fe}^{2+} + \text{Mn})$ in the chloritoids, Fe^{3+} in the epidotes, and $(\text{Fe}_2\text{O}_3 + \text{FeO} + \text{MnO})$ in the garnets. The relevant graphs are available in Mäkanjuola (1967) or are easily constructed from the data presented here.

Nomenclature and Formula Recalculation. The nomenclature used by Deer, Howie and Zussman (1963) and their methods of recalculation have been followed. The minerals generally re-calculated to within 5 percent of the ideal formulae. Any departures from the foregoing are discussed in the section on the specific mineral.

Small amounts of alkalis (and Ca in the case of the chlorites and the chloritoids) are here assumed to substitute in the mineral structures of garnets, epidotes, chloritoids and chlorites—which are often thought to completely exclude them. These minerals have, in general, equilibrated with alkali-rich minerals and there is no doubt that they can absorb small amounts of alkalis, presumably substituting for divalent cations with an equivalent amount of a more highly charged cation, as for instance $(\text{Na}, \text{K}) R^{3+} = 2R^{2+}$, or $(\text{Na}, \text{K}) R^{4+} = R^{2+} + R^{3+}$. Conversely, if the alkalis are regarded as contaminants, then their small amounts in these minerals testify to the relative freedom from contamination due to the closely associated alkali-rich micas, amphiboles and pyroxenes. The inclusion of the alkalis into the structural formulae hardly affects the ratios of cations to oxygen.

Results and Discussions

The Orthosilicates and the Oxides (Rutile)

The orthosilicates and oxides separated from these rocks include ten chloritoids, fifteen epidotes, sixteen garnets, seven rutiles and four sphenes.

The Chloritoids. The results of the chemical analyses of the chloritoids are given in Table 3. In order to be able to compare their compositions with the structural

Table 3. Chemical analyses of the chloritoids

	I								V		Previous analyses	
	LH5	LM5	LM6	LH15	LM20	LM22	MK68	PM6	VM1	VM2	1	2
SiO ₂	24.05	24.12	24.00	24.15	24.66	24.07	24.18	24.11	23.20	22.86	24.90	24.76
TiO ₂	0.24	0.26	0.66	0.27	0.24	0.26	0.28	0.72	0.24	0.28	nd	0.50
Al ₂ O ₃	41.61	40.55	37.19	38.00	37.07	37.62	37.00	36.22	41.67	42.12	40.36	40.15
Fe ₂ O ₃	2.89	4.91	7.30	7.08	7.08	6.20	5.55	4.61	2.71	2.55	nd	6.00
FeO	19.80	19.19	20.94	20.10	19.76	21.55	23.75	24.73	20.56	21.99	26.17	17.30
MnO	0.40	0.28	0.30	0.28	0.44	0.30	0.63	0.78	1.21	0.35	nd	0.59
MgO	3.84	3.92	2.39	3.48	3.71	3.12	1.70	1.79	3.33	2.79	2.54	3.60
CaO	0.16	0.09	0.06	—	0.21	0.31	—	0.04	0.19	0.11	nd	0.13
Na ₂ O	0.09	0.07	0.08	0.03	0.05	0.05	0.03	0.05	0.05	0.03	nd	0.20
K ₂ O	0.11	0.06	0.13	0.11	0.19	0.15	0.12	0.10	0.09	0.05	nd	0.08
H ₂ O ⁺	7.00	6.98	6.89	6.51	6.77	6.49	6.48	7.02	7.13	6.93	6.23	5.94
H ₂ O ⁻	0.05	0.05	0.16	0.04	0.17	0.06	0.05	0.09	0.05	0.05	nd	0.03
F	0.06	0.14	0.07	0.15	0.13	0.09	0.14	0.22	0.14	0.26	nd	nd
O = F	100.30	100.62	100.17	100.20	100.48	100.27	99.91	100.48	100.57	100.37	—	99.28
	0.03	0.06	0.03	0.06	0.06	0.03	0.06	0.08	0.06	0.11	—	—
Total	100.27	100.56	100.14	100.14	100.42	100.24	99.85	100.40	100.51	100.26	—	99.28
S. G.	3.53	3.56	3.56	3.64	3.54	3.59	3.64	3.67	2.54	3.59	—	—
R. l. α	1.714	1.716	1.717	1.716	1.716	1.718	1.718	1.718	1.715	1.715	—	—
β	1.717	1.718	1.720	1.720	1.719	1.721	1.722	1.722	1.718	1.719	—	1.717
γ	1.722	1.726	1.725	1.725	1.725	1.727	1.726	1.728	1.725	1.725	—	1.725
2V (γ)	85°	85°	84°	89°	73°	78°	64°	53°	56°	57°	—	45–50°

Numbers of ions calculated on the theoretical basis of (O, OH, F) = 14 per unit formula

Si	1.962	1.967	2.005	2.011	2.045	2.015	2.047	2.029	1.899	1.886	—	2.05
Al	3.000	3.000	3.000	3.000	3.000	3.000	3.000	3.000	3.000	3.000	—	3.00
Al	0.999	0.897	0.662	0.730	0.624	0.712	0.693	0.590	1.021	1.096	—	0.87
Ti	0.001	0.016	0.042	0.017	0.015	0.017	0.018	0.046	—	—	—	0.13
Fe ⁺³	—	0.087	0.296	0.253	0.361	0.271	0.289	0.292	—	—	—	—
Σ	1.000	1.000	1.000	1.000	1.000	1.000	1.000	0.928	1.021	1.096	—	1.00
Ti	0.014	—	—	—	—	—	—	—	0.015	0.017	—	0.24
Fe ⁺³	0.177	0.215	0.163	0.190	0.081	0.119	0.065	—	0.167	0.140	—	—
Mg	0.467	0.476	0.298	0.432	0.459	0.389	0.215	0.224	0.407	0.343	—	0.44
Fe ⁺²	1.350	1.309	1.463	1.400	1.370	1.509	1.682	1.740	1.408	1.517	—	1.18
Mn	0.027	0.019	0.021	0.020	0.031	0.021	0.045	0.056	0.084	0.024	—	0.02
Ca	0.014	0.008	0.006	—	0.018	0.028	—	0.004	0.017	0.010	—	0.01
Na	0.015	0.011	0.013	0.005	0.008	0.008	0.005	0.008	0.008	0.005	—	0.03
K	0.012	0.006	0.014	0.012	0.021	0.016	0.013	0.011	0.010	0.005	—	—
Σ	2.076	2.044	1.978	2.059	1.988	2.090	2.025	2.043	2.116	2.061	—	1.92
OH	3.807	3.797	3.839	3.615	3.745	3.625	3.660	3.939	3.894	3.814	—	3.20
F	0.008	0.036	0.019	0.040	0.034	0.024	0.038	0.053	0.036	0.068	—	—
Σ	3.815	3.833	3.858	3.657	3.779	3.649	3.698	3.992	3.930	3.882	—	3.20

I = Chloritoids from the aluminous schists.

V = Chloritoids from the veins.

Analysis 1. Vein chloritoid from Île de Groix. Analyst: Renard (1884).

Analysis 2. Chloritoid from glaucophane and chloritoid mica schist (Velde, 1967 b). Analyst: A. Nétillard.

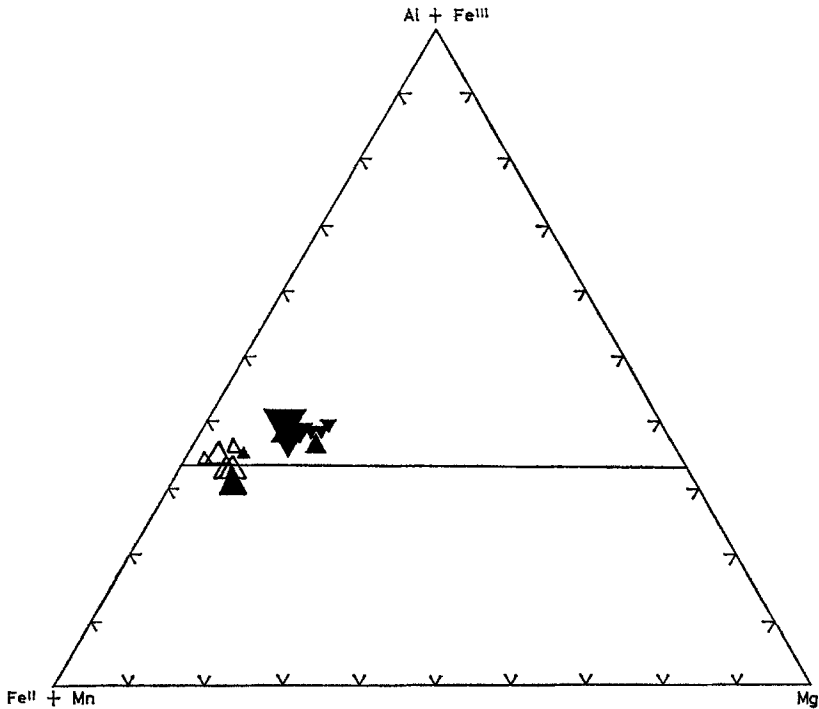


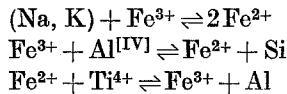
Fig. 1. Chemical composition of natural chloritoids showing variations in the mixed octahedral layer $[(\text{Fe}^{2+}, \text{Mg}, \text{Mn})_2(\text{Al}, \text{Fe}^{3+})]$ of the structure. Analyses from the present work are shown as filled triangles, the open triangles are from Halferdahl (1961)

formula $[(\text{Fe}^{2+}, \text{Mg}, \text{Mn})_2][\text{Al}, \text{Fe}^{3+}]\text{Al}_3[\text{O}_2(\text{SiO}_4)_2](\text{OH}, \text{F})_4$, determined by Brindley and Harrison (1952), their ions were calculated on the theoretical basis of $(\text{O}, \text{OH}, \text{F}) = 14$ per unit formula. The composition of the mixed octahedral layer has been used to judge the dependability of the analyses (Halferdahl, 1961). Fig. 1 compares the mixed octahedral layer compositions with the ideal structural formula. The sizes of the triangles are inversely proportional to the degree of reliability of the analyses if the chloritoids have simple stoichiometric substitutions. One of the analyses of the chloritoid specimens from Île de Groix is less reliable than the others.

In the generalised formula of chloritoids, silicon ions are located at the tetrahedral sites, Fe^{2+} , Mg , Mn , Fe^{3+} , and Al occupy the brucite-type layer and Al ions are exclusively arranged in the corundum-type layer. The Si^{4+} ions approximate the ideal 2 for 14 $(\text{O}, \text{OH}, \text{F})$, see Table 3. The major ionic substitutions in the chloritoids take place in the mixed octahedral layer. The ranges of these substitutions proposed by Halferdahl (1961) are compared with those calculated from the present analyses. Except for the fluorine-hydroxyl substitution, all the observed values of ionic substitution are well within Halferdahl's (1961) range. The five analyses upon which Halferdahl based his $\text{F}/(\text{F} + \text{OH})$ substitution range are considered to be too few for any far-reaching conclusions. The present ana-

lyses show that chloritoids contain fluorine in excess of Halferdahl's upper limit, and a new upper limit of 2 mol. percent is therefore proposed.

In three of the analyses there are sufficient aluminium ions to account for the four possible Al in the structural formula. But Fe^{3+} ions were also determined in these specimens. Fe^{3+} is known to replace Fe^{2+} through oxidation, for example in the substitutions 2Fe^{3+} for 3Fe^{2+} and Fe^{3+} for $\text{Fe}^{2+} + \text{H}^+$ suggested by Yoder (1957). If however, there had been a substitution of the type $2\text{Fe}^{3+} \rightleftharpoons 3\text{Fe}^{2+}$, then the mixed octahedral site should show a deficiency of $1/3\text{Fe}^{3+}$, whereas the site is in excess of 2.000 for the three analyses. If, on the other hand, the substitution $\text{Fe}^{3+} \rightleftharpoons \text{Fe}^{2+} + \text{H}^+$ had been appreciable there should exist a simple correlation between Fe^{3+} and O. Therefore, although Yoder's substitution is certainly important, other possible substitutions in these three specimens could have been:



The first of these substitutions will account for the alkali content of these chloritoids, the second for the deficiency of Si ions in all three and hence the presence of tetrahedral Al, and the third for the Ti of the mixed octahedral layer.

The three analyses showing the highest contents of manganese are from non-garnet-bearing rocks. It is therefore suggested that in a pelitic schist with chloritoid but no associated biotite undergoing progressive metamorphism, most of the manganese is incorporated in the chloritoid. At later stages when garnet appears the manganese content of the rock is redistributed among the early-formed chloritoid and the newly evolving garnet.

The *b* crystallographic axis of the chloritoids coincides with the α vibration direction. This is in agreement with the general conclusion reached by Halferdahl (1961) who gave reasons why he thought Simpson (1915) erroneously concluded that *b* corresponds with β in the chloritoid from Collier Bay. But Hietanen (1951) determined $b = \beta$ in the monoclinic polymorphs of a number of chloritoid twin lamellae that she measured. In this instance the confusion may arise from the twinning. All the studied specimens are optically positive and with fairly large $2V$, which are however in close agreement with measurements for chloritoids (Halferdahl, 1961). The measured refractive indices (Table 3) are also in agreement with Halferdahl's (1961) demonstration that increasing Mg content lowers the refractive indices. It is interesting to note that these chloritoids that have crystallised within very contrasting paragenesis should belong to the same class of monoclinic polymorph.

Renard's analyses (Barrois, 1884) contain too few determined oxides for any possible extensive comparison; but the values in Nétillard's analysis (Velde, 1967b) are very close to the present ones, although his SiO_2 and Na_2O values are too high and the FeO and H_2O^+ too low. As the purity of the powder is not stated, the high sodium content could have resulted from paragonite or glaucophane admixtures. However, since no quartz was discovered as inclusions in the chloritoids studied, it is rather difficult to account for the high silica content of this chloritoid.

Al	2.219	2.170	2.455	2.377	2.109	2.428	2.387	2.408	2.478	2.538	2.427	2.331	2.320	2.547	2.289
Fe ³⁺	0.831	0.865	0.561	0.615	0.858	0.531	0.555	0.546	0.560	0.436	0.539	0.631	0.685	0.481	0.713
Ti	0.003	0.002	0.016	0.013	0.004	0.023	0.038	0.033	0.010	0.014	0.016	0.018	0.023	0.018	0.021
Σ	3.053	3.037	3.032	3.005	2.971	2.982	2.980	2.987	3.048	2.988	2.982	2.980	3.028	3.046	3.023
Ca	1.948	1.949	1.929	1.915	1.952	1.957	1.893	1.951	1.946	1.940	1.907	1.888	1.918	1.928	1.917
Mg	0.008	0.016	0.006	0.017	0.003	0.006	0.008	0.014	0.002	0.010	0.031	0.037	0.013	0.021	0.002
Fe ²⁺	0.044	0.037	0.053	0.065	0.034	0.065	0.082	0.065	0.051	0.058	0.065	0.073	0.053	0.048	0.059
Mn	0.021	0.016	0.009	0.017	0.015	0.013	0.004	0.027	0.005	0.005	0.010	0.011	0.006	0.011	0.016
Na	0.013	0.010	0.010	0.007	0.014	0.017	0.038	0.050	0.008	0.010	0.022	0.020	0.008	0.017	0.012
K	0.009	0.007	0.013	0.006	0.007	0.009	0.007	0.007	0.007	0.007	0.009	0.007	0.009	0.009	0.010
Σ	2.043	2.035	2.020	2.027	2.025	2.067	2.032	2.114	2.019	2.030	2.044	2.036	2.007	2.034	2.016
OH	0.783	0.826	0.991	0.990	1.088	0.947	0.974	0.923	0.845	1.041	1.048	1.047	0.920	0.907	1.005
Fe _x	0.277	0.292	0.187	0.205	0.286	0.177	0.185	0.182	0.186	0.145	0.180	0.210	0.227	0.160	0.238

Epidotes from the aluminous schists (I), the calcareous schists (II), and the basic schists (IV).

The chloritoids have a compositional variation of 10 to 25 percent Mg for Fe and 0 to 9 percent of Fe³⁺ for Al. Halferdahl (1961) reported 0 to 14 percent Fe³⁺ for Al. This therefore implies that the experiments of Halferdahl (1961), Ganguly and Newton (1968) and Hoschek (1969) on the apparent upper stability of pure Fe²⁺ chloritoid are not strictly applicable to these and many other natural chloritoids. The discovery of up to 2 percent F substituting for OH (Halferdahl, 1961, found 0-0.25 percent in the only five samples analysed for fluorine), suggests that partial pressure of fluorine might modify the stability of chloritoids from that determined experimentally using pure H₂O. The possible substitution of Fe³⁺ + O for Fe²⁺ + OH, if a primary feature of the mineralogy (i.e. not an alteration product as it can be with amphiboles), suggests that oxidising conditions may reduce the amount of H₂O in the chloritoid by up to 10 percent, which could also seriously alter the stability field of the mineral.

The Epidotes. The epidotes from Île de Groix belong to the Al-Fe series of Strens' (1966) epidote system. In them the substitution of Al for Si (0.00 to 0.15) is very limited. They were analysed for phosphorus in order to find out if this element substitutes for silicon in this mineral, as suggested by Koritnig (1964) and Myer (1966). They either contained no phosphorus or it was below detection level. They all contained aluminium in excess of the amounts required to balance silicon deficiency and to fill the AlO and the AlOH sites of the epidote structure determined by Ito *et al.* (1954). The excess aluminium and the Fe³⁺ contents are allocated at the "between chain" sites, which therefore contain Al and Fe³⁺ in a ratio between 0.13 and 1.23. The assumption that the Fe³⁺ will be mainly located at the "between chain" sites is supported by Miyashiro and Seki (1958), who explained that

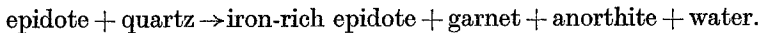
epidotes in which the octahedral sites outside the chains are occupied by Fe^{3+} are more stable than those in which they are filled with Al. Ito *et al.* (1954) also allocated 0.97 Fe^{3+} to these sites in the epidote whose structure they determined.

The small amounts of Ti^{4+} calculated from the epidote analyses are grouped with the octahedral cations on account of their ionic radii. They are probably located at the distorted octahedral or the AlOH sites.

In Table 4, OH shows the largest deviation from the ideal value. Although traces of fluorine were detected, there was no correlation between water deficiency and fluorine content. The wide range of OH values may therefore reflect the difficulty of dehydrating epidotes, as discussed by Smethurst (1935), or may in part be due to an oxy-epidote substitution of the type $\text{Fe}^{2+} + \text{OH}^- \rightarrow \text{Fe}^{3+} + \text{O}^{2-}$.

Besides the alkalis, Mg, Fe^{2+} and Mn^{2+} are also taken as substituting for Ca. But on account of ionic radii the amounts of the first two elements cannot be considerable without distorting the epidote lattice. The alkalis on the other hand with their larger ionic radii may conveniently proxy for calcium. For example, Sakurai and Nagashima (1956) recorded the analyses of a clinozoisite from Arai, Saitama Prefecture, that contained 1.14 percent Na_2O , 1.19 percent K_2O and 1.71 percent MgO . From considerations of crystal chemistry, Strens (1964) concluded that the paired substitutions $(\text{Na}, \text{K})^+\text{Si}^{4+}$ for $\text{Ca}^{2+}\text{Al}^{3+}$ probably take place in such alkali clinozoisite; this may also be the case in these epidotes.

The pistacite, $\text{Ca}_2\text{Fe}^{3+}\text{Si}_3\text{O}_{12}(\text{OH})$, components of the epidotes lie between 15 and 29 molecular percent. Specimens with the highest Fe^{3+} contents are, however, those from non-garnet-bearing schists or schists in which garnet is just beginning to appear. If one assumes the appearance of garnet on Île de Groix to be controlled by $P_f - P_s - T$ rather than by bulk chemistry, this observation is contrary to Strens' (1964) hypothesis that epidotes formed in equilibrium with garnets should contain more Fe^{3+} than the lower grade epidotes. Strens' equation is:



But the epidotes of the glaucophane-epidote assemblages of Île de Groix did not evolve in equilibrium with garnets. Petrographic evidence (Makanjuola, 1967), showed that for the most part, especially in the basic schists, epidotes formed retrogressively from the garnets. Composite grains, consisting of garnet cores and peripheral epidotes are very common in these schist. This is not to say that garnets are not stable in glaucophane schist facies rocks; garnets occur in equilibrium with clinozoisite in for example the glaucophane-bearing rocks of Japan and California. The glaucophane-epidote schists of Île de Groix belong to the glaucophane greenschist facies of Winkler (1965) and are believed to have formed under high $P(\text{O}_2 + \text{H}_2\text{O})$, such that Strens' equation



proceeded from right to left. In contrast, the large amounts of CaFe^{2+} found in the glaucophane schist facies garnets (Pabst, 1931; Lee *et al.*, 1963) have been attributed to the foregoing reaction proceedings from left to right at high pressures and low temperatures.

The b cell dimensions and the differences between the 2θ values of the 020 planes of the epidotes and those of the silicon internal standard, $2\theta(020_{\text{Ep}} - 111_{\text{Si}})$,

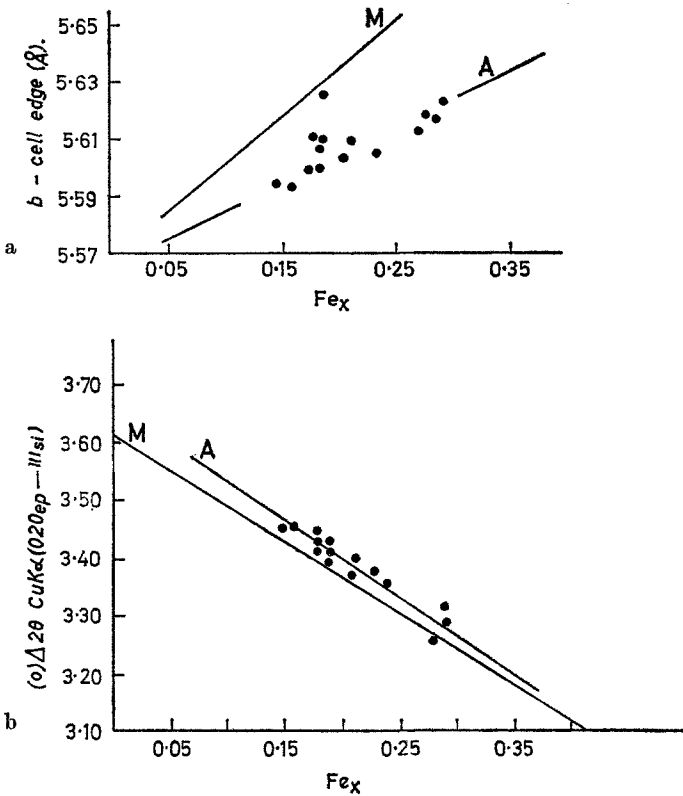
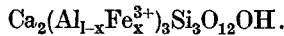


Fig. 2. (a) Epidote *b*-cell dimensions shown as functions of Fe_x . (b) Epidote $2\theta CuK\alpha$ (020_{Ep} , -111_{Si}) shown as functions of Fe_x

were calculated. Figs. 2a and b show both quantities plotted against the Fe_x values calculated for each specimen, in the general epidote formula



In both figures the equations of the straight lines that best fit the plotted data fail to agree with Myer's (1965, 1966) determinative curves for hydrothermal epidotes and epidotes from metamorphic zones. All the plotted points fall below Myer's line (M) in Fig. 2a and above it in Fig. 2b. Although Myer (1966) indicated that the function $Fe_x = 2.985 - 0.827(\Delta 2\theta CuK\alpha_{020_{Ep}} - 111_{Si})$ adequately described the values measured from the twenty epidotes from various geological environments, only two of these are from glaucophane schist metamorphic zones, and the regression equations fail to apply in the present case.

There are straight line relationships between Fe_x on the one hand and the three principal refractive indices α , β and γ on the other. A good correlation also holds between the densities of the epidotes and the values $100 Fe_2O_3 / (Fe_2O_3 + Al_2O_3)$ calculated from the analyses.

The Garnets. Garnet did not crystallize within the magnesian schists, it was rare in the amphibole-bearing sub-type of the basic schists, common in the aluminous

	1.986	1.962	1.851	1.993	1.714	1.893	1.977	1.852	1.815	1.859	1.757	1.623	1.815	1.914	1.890	1.728	Fe ⁺²
Ca	0.587	0.486	0.479	0.461	0.590	0.736	0.728	0.930	0.879	0.862	0.855	0.841	0.966	0.805	0.784	0.908	Ca
Mg	0.088	0.119	0.176	0.147	0.135	0.241	0.247	0.223	0.296	0.289	0.219	0.342	0.224	0.231	0.267	0.294	Mg
Mn	0.277	0.324	0.324	0.279	0.491	0.116	0.122	0.056	0.062	0.068	0.123	0.169	0.091	0.126	0.063	0.093	Mn
Na	0.008	0.010	0.011	0.010	0.008	0.011	0.013	0.009	0.005	0.010	0.011	0.008	0.013	0.008	0.008	0.007	Na
K	0.009	0.011	0.011	0.010	0.012	0.007	0.015	0.010	0.014	0.007	0.011	0.012	0.013	0.012	0.015	0.010	K
Σ	2.955	2.912	2.852	2.900	2.950	3.004	3.102	3.080	3.071	3.095	2.976	2.995	3.122	3.096	3.027	3.040	Σ
Almandine	68	68	66	69	58	63	64	60	60	61	60	54	59	63	63	57	Almandine
Grossular	6	8	4	6	10	12	12	21	18	19	18	20	26	17	14	21	Grossular
Andradite	14	9	13	10	10	13	12	10	10	9	11	9	5	9	12	9	Andradite
Spessartine	9	11	11	10	17	4	4	2	2	2	4	6	3	4	2	3	Spessartine
Pyrope	3	4	6	5	5	8	8	7	10	9	7	11	7	7	9	10	Pyrope

Garnets from the aluminous schists (I), the calcareous schists (II), and the basic schists (IV).
 a Deviation in all cases in ± 0.001.

schists, plentiful in the calcareous schists and abundant in the pyroxene-bearing sub-type of the basic schists. It has been demonstrated (Makanjuola, 1967) that the evolution of garnets in the rocks predated that of glaucophane. Therefore, strictly speaking, they are not glaucophane schist garnets. They are metastable relics out of equilibrium with the blue amphiboles and the epidotes. None of the analysed garnets showed any visible zoning although work with an electron microprobe may show zoning: most of them contain inclusions.

Calculations from the chemical analyses show that they are almandine garnets in which the $Fe_3^{2+}Al_2Si_3O_{12}$ molecule predominates over any of the other components by a ratio of at least 3:1 (Table 5). They contain low amounts of alkali which for ionic radii considerations are added to the divalent cations. Although there are some grossulars known with 1-3 percent alkalis, the alkali content of most garnets is negligible and the alkalis are often not even determined in analyses.

The proportion in which end-member molecules are combined in garnets have been investigated and found to be strongly influenced by the chemical and the physical environments under which they have formed. Heritsch (1926), Buddington (1952) and Tröger (1959), among others, discovered obvious relationship between the composition of the host rock (a chemical environment) and the garnet evolved in it; see also Hutton (1962). Variations in the compositions of garnets, which are related to the grade of metamorphism, have been established by Goldschmidt (1921), Miyashiro (1953), Engel and Engel (1960), Sturt (1962), and Atherton (1965). These investigators noticed two trends with increase in grade: Mn^{2+} decreases and Fe^{2+} in-

creases, Mg increases and Ca decreases. The rocks of Île de Groix belong to several chemical groupings, thus providing a spatial distribution of differing chemical environments. Also during metamorphic reconstitution they were not closed systems at least to the volatile phases such as water and CO_2 . Furthermore they are polymetamorphic schists. During each metamorphic episode there would have been spatial distribution of physical environments, i.e. metamorphic zoning, and during succeeding episodes physical conditions must have changed with time. All these differing conditions have left their imprints upon the mineral assemblages by either modifying their chemical compositions or totally reconstituting them (Makanjuola, 1967). With regards to the garnets, the influence of the composition of the host rock becomes apparent when one compares the chemical compositions of specimens derived from different rock groups that are known to have formed under similar set of physical conditions. Compositional differences that are attributable to metamorphic grade are best seen by comparing individual garnet compositions within a given chemical grouping.

The chemical composition of the garnets is expressed in terms of five end-member molecules in Table 5. The molecular averages of these end-members for each rock group are given in Table 6, which also contains the group averages of the oxides FeO, CaO, MgO, Fe_2O_3 and MnO for the host rocks. As is usual with metamorphosed argillaceous sediments the garnets from the aluminous schists are rich in almandine. The pyrope content of the garnets is directly related to the magnesium content of the respective rock. On the other hand their grossular contents fail to reflect the host rock composition because the garnets of the basic schist show higher concentrations than those from the calcareous schists. Essene (personal communication) has suggested that the higher grossular content of the

Table 6. Group averages of garnet end-member molecules and some oxides of the rocks

	Garnets from the:			Average for all garnets
	aluminous schists	calcareous schists	basic schists	
Almandine	68	62	60	63
Grossular	6	11	19	13
Pyrope	5	7	9	7
Andradite	11	11	9	10
Spessartine	10	9	3	7

	Rocks from the:		
	aluminous schists	calcareous schists	basic schists
Av. FeO (wt.%)	4.4	6.0	8.3
Av. CaO (wt.%)	1.2	12.5	8.0
Av. MgO (wt.%)	1.9	3.6	6.0
Av. Fe_2O_3 (wt.%)	3.0	7.2	3.6
Av. MnO (wt.%)	3.7	0.5	2.7

garnets of the basic schist over those of the calcareous schists could be due to the fact that high CO_2 pressures will depress the grossular content of a garnet at relatively low temperatures. The andradite molecule shows an almost uniform distribution among all the garnets of the schists. Its values most probably reflect the number of mineral phases among which the available calcium and ferric iron are partitioned, e.g. epidote, glaucophane and magnetite.

Two specimens KM11 and LH11 are here compared in order to illustrate the influence of grade on the garnet compositions. The host rocks from which these two specimens were obtained contained approximately the same modal amounts of garnet. Both rocks were also free from any other manganese phase such as chloritoid or biotite (Makanjuola, 1967). Relevant data about these two rock and mineral specimens are as follows:

	Rocks		Garnets	
	KM11	LH11	KM11	LH11
Wt. %				
MnO	0.67	0.19	7.16	1.70
FeO	2.98	4.01	25.33	28.05
CaO	13.42	16.12	6.81	8.52
MgO	6.66	1.37	1.12	2.00
Garnet mode			3.99	3.81

Both specimens are from the calcareous schists, but while garnet is just being evolved in KM11, specimen LH11 has reached an advanced stage of metamorphism. Furthermore the mineral assemblage of KM11 include calcite. The influence of grade upon the concentration of MgO and MnO is quite evident, especially in respect of MgO. Both garnets are almandine and hence appropriately enriched in FeO. The CaO contents, however, vary contrary to expectation, but again, the depletion of specimen KM11 in CaO may be due to a higher partial pressure of CO_2 .

The Sphenes. The sphene analyses presented in Table 7 show appreciable amounts of combined water, fluorine and aluminium. On account of these, substitutions of the type $\text{O}_1 \rightarrow (\text{OH}, \text{F})$, proposed by Sahama (1946), and $3(\text{Ti}^{4+} + \text{Si}^{4+}) \rightarrow 4\text{Al}$ are believed to be present in the minerals. The values recorded in Table 7 as Al_2O_3 were obtained by difference, i.e. weight percent $\text{Al}_2\text{O}_3 = \text{weight percent } R_2\text{O}_3 - (\text{MnO} + \text{total } \text{Fe}_2\text{O}_3 + \text{TiO}_2 + \text{P}_2\text{O}_5)$. Therefore, there exists the probability that the values for alumina might include those of the oxides of niobium, tantalum and vanadium, elements sometimes found in fair amounts in some sphenes of igneous origin (Sahama, 1946). But if these transitional elements were present at all their concentrations would be negligible in metamorphic sphenes.

The Rutiles. The rutiles were the only oxide mineral obtained from the schists, and they were the only specimens analysed by a combination of X-ray spectroscopy and wet chemical methods. Their analyses given in Table 8 show substitutions of the type $\text{Fe}^{2+}\text{Mn}^{2+}$ for Ti^{4+} . The amounts of both replacing cations are generally low in all the analyses. The powders of the more iron-rich specimens are of darker brown colour.

Table 7. Chemical analyses of sphenes

	LM1	PYM1	LH3	LH6		Lm1	PYM1	LH3	LH6	
SiO ₂	30.08	29.32	29.72	29.13	Ions calculated on the theoretical basis of (O, OH, F) = 5 per unit formula					
TiO ₂	37.92	40.42	37.00	35.55		Si	0.978	0.951	0.966	0.954
Al ₂ O ₃	2.11	2.50	2.80	3.97		Al	0.022	0.049	0.034	0.046
Fe ₂ O ₃	0.48	0.14	0.09	0.84		Σ	1.000	1.000	1.000	1.000
FeO	—	—	—	1.00		Al	0.059	0.047	0.073	0.108
MnO	0.03	0.06	0.11	0.25		Fe ⁺³	0.012	0.004	0.002	0.021
MgO	27.82	26.41	28.46	27.78		Fe ⁺²	—	—	—	0.027
CaO	0.07	0.12	0.10	0.12		Ti	0.927	0.951	0.904	0.875
Na ₂ O	0.10	0.10	0.09	0.11		Mg	0.001	0.003	0.006	0.012
K ₂ O	0.15	—	0.19	0.07		Σ	0.999	1.005	0.985	1.043
H ₂ O ⁺	0.43	0.24	0.40	0.22		Mn	0.004	—	0.005	0.002
H ₂ O ⁻	0.11	0.10	0.06	0.08		Na	0.004	0.007	0.006	0.007
F	0.46	0.63	0.88	0.65		K	0.004	0.004	0.004	0.005
Total	99.76	100.04	99.90	99.77		Ca	0.970	0.918	0.991	0.975
F ≡ O	0.19	0.27	0.37	0.27		Σ	0.982	0.929	1.006	0.989
Total	99.57	99.77	99.53	99.50	OH	0.015	0.052	0.087	0.048	
S. G.	3.52	3.55	3.54	3.51	F	0.048	0.065	0.090	0.067	
2V _γ	30°	35°	31°	36°	Σ	0.063	0.117	0.177	0.115	

Table 8. Partial chemical analyses of the rutiles

	LH5	LM5	LH11	LH13	LH4	LM4	LM17
FeO	0.07	1.74	0.16	0.32	0.16	0.15	0.22
MnO	0.15	0.40	0.07	0.10	0.15	0.15	0.22
H ₂ O ⁻	0.09	0.16	0.11	0.08	0.08	0.17	0.16
Specific gravity	4.243	4.275	4.244	4.250	4.246	4.246	4.253
Unit cell dimensions in Å							
<i>a</i>	4.588	4.592	4.587	4.589	4.588	4.586	4.591
<i>c</i>	2.958	2.959	2.958	2.959	2.956	2.957	2.956

The Chain Silicates

The chain silicates consist of seven pyroxenes, thirteen blue sodic amphiboles, a tremolite-actinolite and a green barroisite. All but two of these minerals were obtained from the basic schists, the exceptions being a ferroglaucophane from the aluminous schists and a tremolite-actinolite from the magnesian schist.

The Pyroxenes. The pyroxenes are discussed first because, except for the two minerals mentioned above, all the other amphiboles were directly or indirectly derived from them. The chemical analyses of the seven pyroxenes, their calculated formulae and their compositions expressed in terms of end-member molecules are given in Table 9. This table also contains the specific gravity and the optical

Table 9. Chemical analyses of the pyroxenes (all from the basic schists)

	LM1	PYM1	LH3	LH6	LH9	LH12	LH16
SiO ₂	52.50	54.23	54.54	53.10	53.77	53.99	53.47
TiO ₂	0.03	0.20	0.23	0.27	0.03	0.41	0.42
Al ₂ O ₃	12.60	8.46	8.89	11.95	9.84	10.43	9.93
Fe ₂ O ₃	5.70	5.73	7.07	5.60	5.34	5.23	8.96
FeO	2.88	3.70	2.89	2.93	2.80	3.20	2.19
MnO	nd	0.03	0.03	0.05	0.13	0.05	0.03
MgO	6.09	7.29	6.48	6.10	7.77	7.20	5.94
CaO	13.81	13.93	12.38	12.99	13.19	12.63	11.66
Na ₂ O	6.29	6.32	7.20	6.75	7.40	7.28	7.28
K ₂ O	0.06	0.18	0.06	0.23	0.05	0.14	0.14
Total	99.96	100.07	99.77	99.97	100.32	100.56	100.02
S. G.	3.366	3.365	3.366	3.350	3.347	3.369	3.388
R.I. α	1.685	1.686	1.685	1.685	1.684	1.685	1.688
β	1.689	1.694	1.697	1.692	1.692	1.694	1.698
γ	1.697	1.700	1.702	1.698	1.699	1.699	1.704
2V γ	76°	74°	76°	74°	78°	72°	78°
Ions calculated on the basis of O = 6 per unit formula							
Si	1.896	1.968	1.977	1.917	1.939	1.939	1.937
Al	0.104	0.032	0.023	0.083	0.061	0.061	0.063
ΣT	2.000	2.000	2.000	2.000	2.000	2.000	2.000
Al	0.432	0.330	0.357	0.425	0.357	0.381	0.361
Ti	0.001	0.006	0.006	0.007	0.001	0.011	0.012
Fe ⁺³	0.155	0.157	0.193	0.152	0.145	0.142	0.244
Fe ⁺²	0.087	0.112	0.088	0.089	0.085	0.096	0.066
Mn	—	0.001	0.001	0.002	0.004	0.002	0.001
Mg	0.328	0.394	0.350	0.328	0.418	0.385	0.321
$\Sigma M1$	1.003	1.000	0.995	1.003	1.010	1.017	1.005
Na	0.440	0.445	0.506	0.472	0.517	0.507	0.512
Ca	0.534	0.542	0.481	0.502	0.510	0.486	0.453
K	0.003	0.009	0.003	0.010	0.002	0.002	0.007
$\Sigma M2$	0.977	0.996	0.990	0.984	1.029	0.995	0.972
Compositions expressed as molecular percentages of end-members							
Ts	11	3	2	8	6	6	6
Ac	16	16	19	15	14	14	25
Jd(Na-Fe ⁺³)	29	29	31	32	36	36	27
Jd(Al ^{VI} -Al ^{IV})	34	30	36	34	29	32	30
	32 ± 3	30 ± 1	34 ± 3	33 ± 1	33 ± 3	34 ± 2	28 ± 2
Di + Hd + En + Fs	42	51	44	42	49	48	39
Total	100 ± 3	99 ± 1	99 ± 3	98 ± 1	101 ± 3	102 ± 2	99 ± 2

data for these minerals. The chemical formulae are expressed as (M2)(M1)T₂O₆, where M2 consists of Ca, Na and K; M1 of Al, Fe³⁺, Fe²⁺, Mg, Mn, and Ti and T of Si and Al^[IV]. The low values of Al^[IV] (between 1 and 5 percent of the sum of tet-

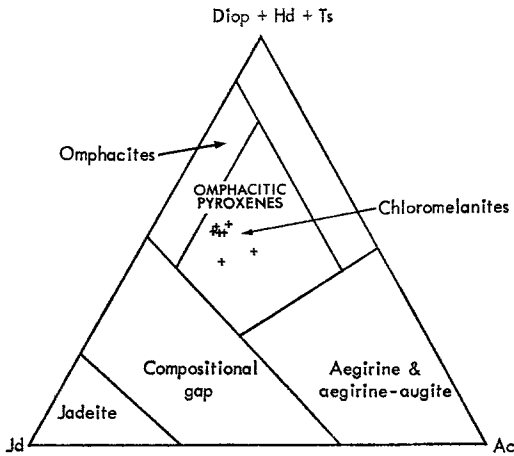


Fig. 3. Triangular composition diagram of the pyroxenes from Île de Groix

rahedrally coordinated ions) indicate that the minerals were formed under high pressure (Thompson, 1947; Yoder, 1950; Coleman and Clark, 1968). The sums of the M1 cations are slightly higher than the expected 1.00, except in specimens PYM1 and LH3 where they equal that value, but also with a single exception, the sums of the M2 cations are correspondingly less than 1.00.

In the calculation for the end-member molecules two methods were used in determining the jadeite components, i.e. $\text{Al}^{\text{VI}} - \text{Al}^{\text{IV}}$ and $\text{Na} - \text{Fe}^{3+}$. Although these two values fail to coincide, as they should, their deviations from the mean values (maximum ± 3) are within experimental errors. Their failure to agree may also mean that other substitutions are significant or that site deficiencies are present. Their low Ts contents (Al^{IV}) is likely to be explained by equilibria involving epidote at high $P_{\text{H}_2\text{O}}$ and relatively low temperatures when the Ts molecule will break down to epidote group minerals.

In the compositional diagram (Fig. 3) the pyroxenes plot in the field of omphacitic pyroxenes derived from rocks of basaltic compositions. Their composition field in the diagram, outside the indicated composition gap, supports the miscibility limits and mean composition of sodium-aluminium pyroxene (jadeite) and sodic diopside (omphacite), discussed by Dobretsov (1964). They are classified as chloromelanite on account of their relatively high 14 to 25 molecular percent of aemite (Winkler, 1965).

In Table 10 the observed peaks and the b cell dimensions calculated from the 020 spacings are listed and compared with the indexed observed peaks of the Tiburon omphacite (100-RGC-58) reported by Coleman and Clark (1968). The close similarities of the listed values suggest that these chloromelanites also have a $P2$ symmetry; but this has yet to be confirmed by single-crystal photographs.

Sakata (1957) showed that the parameter likely to vary most in this type of pyroxenes is b (see also Clarke *et al.*, 1962). A smooth curve results from a plot of b versus $(\text{Fe}^{2+} + \text{Fe}^{3+})$, with b increasing with increasing iron contents. The fit for b versus Al^{VI} is not as good as for b versus $(\text{Al}^{\text{VI}} - \text{Fe}^{3+})$; in both cases b decreases with increasing Al (Makanjuola, 1967). This is an expected result because aluminium has a smaller ionic radius than both ferric and ferrous ions.

Table 10. X-ray diffraction powder data for pyroxenes from the Basic Schists of Île de Groix (observed d in Å)

hkl^a	LM1	PYM1	LH3	LH6	LH9	LH12	LH16	100-RGC-58 ^a
110	6.305	6.460	6.340	6.310	6.280	6.325	6.315	6.38
020	4.366	4.444	4.381	4.366	4.362	4.396	4.390	4.40
221	2.970	2.970	2.970	2.967	2.970	2.964	2.970	2.970
310	2.885	2.888	2.890	2.881	2.895	2.880	2.885	2.893
131	2.537	2.538	2.539	2.531	2.538	2.535	2.538	2.541
002	2.520	2.516	2.521	2.519	2.522	2.516	2.516	2.521
221	2.472	2.476	2.476	2.473	2.475	2.469	2.471	2.474
311	2.255	2.260	2.258	2.256	2.256	2.252	2.251	2.258
112, 022	2.194	2.196	2.195	2.193	2.192	2.189	2.189	2.193
330, 331	2.112	2.112	2.114	2.114	2.114	2.109	2.110	2.112
421	2.083	2.084	2.086	2.085	2.085	2.080	2.080	2.082
041	2.016	2.015	2.017	2.015	2.015	2.012	2.012	2.014
150	1.729	1.731	1.731	1.728	1.731	1.725	1.726	1.726
b (Å)	8.710	8.734	8.734	8.710	8.694	8.704	8.750	8.771 ± 6

^a The hkl reflection planes and the d -spacing in the last column were listed for the omphacite specimen (100-RGC-58) by Coleman and Clark (1968). Additional d -spacings measured in some but not others have been omitted; values were obtained using a smear mount with internal standard and Co K_{α} radiation (Fe filter).

The observed optical and physical properties, though slightly higher, are comparable with those of omphacite 100-RGC-58 from Tiburon Peninsula, Marin County (Coleman and Clark, 1968). These slight increase are attributable to the higher iron content of the Île de Groix pyroxenes.

The amphiboles are represented by thirteen blue and two green varieties. The chemical compositions of the host rocks appear to determine their type and abundance. For example, the magnesium-rich rocks, such as the basic and the magnesian schists, show high contents of amphiboles, but the aluminous and the calcareous schists with low magnesia carry them only as accessory phases. The sodic amphiboles are restricted to the sodium-bearing basic, calcareous and aluminous schists, and tremolite-actinolite to the sodium-deficient magnesian schists, see Table 1.

The Blue Amphiboles. Blue amphiboles are most abundant in the basic schists, moderately developed in the calcareous schists, fairly common in the aluminous schists and absent in the magnesian schists. The results of the analyses of thirteen specimens together with their optical and physical properties are listed in Table 11. The analyses were recalculated to the amphibole formula, $AX_2Y_5Z_8O_{22}(OH, F)_2$. All but one analysis, LH 16, fall within the limits suggested by Phillips (1963) for good analyses. In analysis LH 16 the sum of the X-site cations is below the expected value of 2 by 0.47. The OH group in all cases but this one specimen, is below 2. If the low water content is due to the inherent difficulty in determining water in amphiboles (Borley, 1963), then the actual water content of specimen LH 16 must have been even higher the determined value. These observations strongly suggest the presence of H_3O^+ at A sites as sometimes met with in hydromuscovites. The $(Na^+ + Ca^{2+} + K^+)$ totals in the other specimens approximate so

Table 11. Chemical analyses of the blue amphiboles

	I		II		IV								
	PM6	LH11	LH13	LM1	PYM1	LH3	LH4	LM4	LH6	LH12	LH16	LM17	LM3
SiO ₂	55.19	57.76	57.22	56.97	57.11	56.72	57.01	57.71	56.38	56.82	56.48	57.68	55.92
TiO ₂	0.03	0.06	0.04	0.03	0.02	0.17	0.03	0.14	0.05	0.23	0.04	0.05	0.03
Al ₂ O ₃	10.48	10.08	9.03	10.92	9.45	12.00	10.73	13.43	10.86	10.91	10.96	10.32	10.21
Fe ₂ O ₃	4.60	3.55	4.83	3.29	5.06	2.53	4.77	1.01	4.22	2.40	4.53	4.93	5.04
FeO	14.37	8.16	9.82	9.25	8.90	8.05	8.12	7.81	8.54	9.80	9.38	5.56	10.85
MnO	0.10	0.08	0.06	0.02	0.01	0.01	0.03	0.01	0.07	0.02	0.03	0.03	0.04
MgO	5.91	10.53	8.65	9.44	9.61	10.37	9.58	11.16	9.92	9.65	9.63	10.96	7.98
CaO	0.60	0.60	0.52	0.90	1.11	1.19	0.73	1.08	0.74	1.10	0.80	1.34	0.54
Na ₂ O	6.92	6.89	7.29	7.29	6.62	6.73	7.08	6.90	7.13	6.86	5.41	6.62	6.82
K ₂ O	0.04	0.14	0.13	0.04	0.11	0.20	0.09	0.07	0.11	0.08	0.05	0.45	0.12
H ₂ O ⁺	1.81	2.13	2.08	2.00	2.03	2.03	2.03	2.16	2.00	2.00	2.25	2.15	2.10
H ₂ O ⁻	0.11	0.21	0.18	0.16	0.11	0.10	0.12	0.05	0.10	0.13	0.21	0.13	0.13
F	0.02	0.02	0.02	0.02	0.02	0.02	0.03	0.01	0.02	0.02	0.01	0.02	0.01
	100.18	100.21	99.87	100.33	100.16	100.10	100.35	100.54	100.14	100.02	99.78	100.24	99.79
F ≡ O	0.01	0.01	0.01	0.01	0.01	0.01	0.01	—	0.01	0.01	—	0.01	—
Total	100.17	100.20	99.86	100.32	100.15	100.11	100.34	100.54	100.13	100.01	99.78	100.23	99.79
S. G.	3.266	3.169	3.167	3.169	3.172	3.187	3.143	3.143	3.193	3.149	3.167	3.164	3.246
R.l. α	1.635	1.627	1.626	1.624	1.633	1.629	1.632	1.625	1.632	1.631	1.628	1.624	1.639
β	1.643	1.632	1.631	1.633	1.637	1.639	1.635	1.632	1.635	1.634	1.638	1.632	1.645
γ	1.651	1.638	1.639	1.369	1.642	1.645	1.640	1.639	1.642	1.646	1.639	1.639	1.650
2V _α	48°	46°	44°	39°	37°	38°	39°	43°	40°	41°	37°	44°	40°
Blue amphibole ions calculated on the theoretical basis of (O, OH, F) = 24 per unit formula													
Si	7.863	7.941	7.995	7.877	7.926	7.792	7.863	7.689	7.810	7.875	7.866	7.883	7.852
Al	0.137	0.059	0.005	0.123	0.074	0.208	0.137	0.302	0.190	0.125	0.134	0.117	0.148
Σ Z	8.000	8.000	8.000	8.000	8.000	8.000	8.000	8.000	8.000	8.000	8.000	8.000	8.000
Al	1.618	1.575	1.483	1.656	1.472	1.735	1.606	1.846	1.583	1.658	1.634	1.545	1.541
Ti	0.003	0.007	0.004	0.003	0.003	0.018	0.003	0.015	0.005	0.024	0.004	0.005	0.003
Fe ³⁺	0.493	0.367	0.509	0.420	0.529	0.263	0.494	0.104	0.440	0.250	0.467	0.508	0.533
Fe ²⁺	1.712	0.938	1.148	1.070	1.033	0.925	0.936	0.887	0.990	1.136	1.073	0.636	1.274
Mn	0.012	0.009	0.008	0.003	0.001	0.001	0.003	0.001	0.008	0.003	0.003	0.003	0.005
Mg	1.256	2.158	1.801	1.946	1.989	2.123	1.970	2.258	2.048	1.994	1.964	2.233	1.671
Σ Y	5.094	5.054	4.953	5.098	5.027	5.065	5.012	5.111	5.074	5.065	5.145	4.930	5.027
Na	1.912	1.837	1.975	1.945	1.781	1.793	1.893	1.815	1.914	1.844	1.436	1.754	1.856
K	0.007	0.025	0.024	0.007	0.020	0.035	0.017	0.011	0.020	0.015	0.008	0.079	0.022
Ca	0.092	0.088	0.078	0.134	0.165	0.175	0.108	0.157	0.110	0.163	0.118	0.196	0.081
Σ A + X	2.011	1.950	2.077	2.086	1.966	2.003	2.018	1.983	2.044	2.022	1.562	2.029	1.959
OH	1.721	1.953	1.938	1.844	1.879	1.860	1.868	1.956	1.848	1.849	2.054	1.961	1.959
F	0.009	0.009	0.009	0.009	0.009	0.009	0.013	0.004	0.009	0.003	0.004	0.009	0.004
Σ	1.730	1.962	1.947	1.853	1.888	1.869	1.881	1.960	1.857	1.852	2.058	1.970	1.963
Fe(*)	23	19	26	20	26	13	23	5	22	12	22	25	26
Fe(t)	58	30	39	35	34	30	32	28	33	36	35	22	43

Blue amphiboles from the aluminous schists (I), the calcareous schists (II) and the basic schists (IV).

$$\text{Fe}^*(*) = 100 \text{Fe}^{3+}/(\text{Fe}^{3+} + [\text{Al}]^6 + \text{Ti}).$$

$$\text{Fe}(t) = 100 \text{Fe}^{2+}/(\text{Fe}^{2+} + \text{Mn} + \text{Mg}).$$

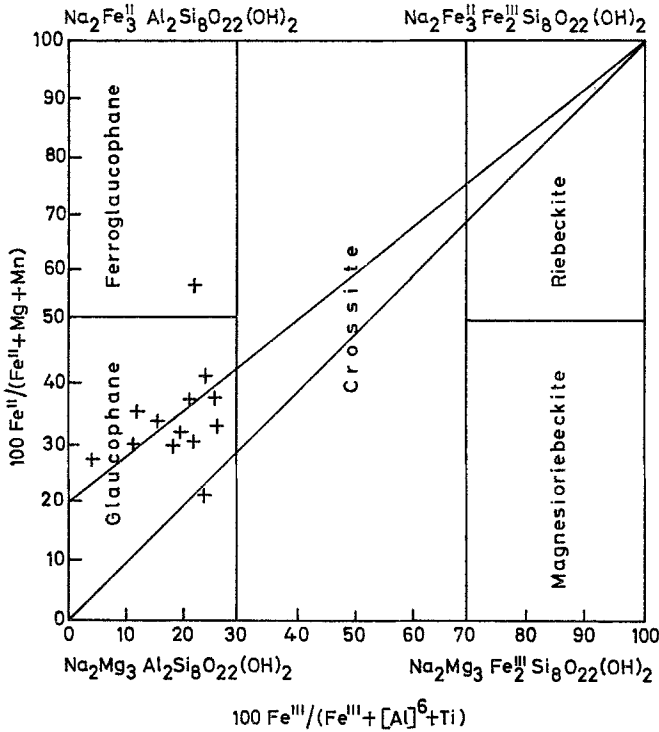


Fig. 4. Blue amphibole composition fields (after Miyashiro, 1957)

close to 2.0 that the A site must be regarded as virtually unoccupied; see Table 11. There is very limited substitution between Na^+ and Ca^{2+} and between Si^{4+} and Al^{3+} . There is however a wide range of $\text{Fe}^{2+} \rightleftharpoons \text{Mg}^{2+}$ and $\text{Fe}^{3+} \rightleftharpoons \text{Al}^{3+}$ substitutions. The replacement of OH by F is very small and uniform.

The structural investigation of glaucophane II by Papike and Clark (1966) supports the suggestion of Ernst (1963) that the high-pressure low-temperature polymorph of glaucophane is characterised by a high degree of cation ordering. The blue amphiboles under investigation are believed to have formed under physical conditions similar to the above. Their analyses on recalculation are shown generally to fit the ordering scheme by which the M_4 position is occupied by Na and Ca, the M_2 by Al and Fe^{3+} and the M_3 by Mg and Fe^{2+} . Specimens LM4 and LH12, however, show slight deficiency of $\text{Al} + \text{Fe}^{3+}$, thus necessitating the presence of some Mg at M_2 sites.

The thirteen blue amphibole analyses are plotted on Miyashiro's (1957) chemical diagram after grouping Mn with Fe^{2+} and Ti with Fe^{3+} because of their similar ionic radii and comparative influence on optical properties (Borg, 1967). One of the analyses, PM6, plots in the ferroglaucophane field (Fig. 4). All the other analyses plot within the glaucophane composition field. A diagonal drawn between $G_{80}Fg_{20}$ and riebeckite provides a line of better fit for the plotted points than the glaucophane-riebeckite tie line proposed by Borg (1967). A similar trend is noticeable on the diagram by Coleman and Papike (1968), and even on Borg's (1967) diagram. It is therefore thought that the failure of blue amphibole compositions

to fall on Borg's tie line is a natural trend and not due to inaccurate analyses or impurities. The ferroglaucophane was the only specimen obtained from the aluminous schists. But this does not mean that the blue amphiboles of the aluminous schists are exclusively iron-rich varieties. The optical properties of the other blue amphiboles studied petrographically in these schists are those of glaucophane. The two blue amphiboles with the highest $100\text{Fe}^{2+}/(\text{Fe}^{2+} + \text{Mg} + \text{Mn})$ ratios in Table 11 were obtained from rocks in which garnet had not appeared. It is therefore suggested that the iron-rich blue amphiboles are produced from low grade chlorites and the glaucophane in types I and II rocks from high grade chlorites. The reason for this being that the low grade chlorites are iron-aluminium chlorites, but on the appearance of garnet these react with quartz to give almandine and Mg-rich chlorites (Winkler, 1965). This argument appears to be applicable equally well to any other ferromagnesian mineral associated with almandine garnet. For example, the blue amphiboles of the basic schists that evolved from the pyroxenes coexisting with almandine garnets are also glaucophane. In these schists, therefore, the Fe^{2+}/Mg ratios of the amphiboles reflect the grade of the original schist in which the mineral formed.

The indices of refraction of the blue amphiboles are related to the total iron content and generally increase with increasing iron content.

The Green Amphiboles. Consist of a barroisite (LH 14) and an actinolite (BM 1). The barroisite was separated from a glaucophane-garnet basic schist and the actinolite from a chlorite-actinolite magnesian schist. The chemical analyses, refractive indices, 2V and the specific gravity of the two amphiboles are given in Table 12.

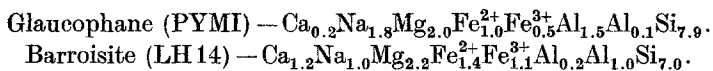
Following Winchell (1932), BM 1, which contains 12.85 molecular percentage ferroactinolite, has been classified as actinolite. In this actinolite the octahedral and the tetrahedral aluminium are low and so also is the Fe^{2+} substitution for Mg. The sum of the X-site cations is close to 2, therefore the A-site may be regarded as virtually empty. Fluorine was below the detection level of the method used but since OH approximates to the ideal 2, if this element is present at all it must be in trace amounts.

Specimen LH 14 has characteristics that are similar to those of the dark green mineral with a composition intermediate between glaucophane and hornblende named barroisite by Murgoci (1922). Binns (1967) suggested that this name, barroisite, be used for sodium-rich aluminous amphiboles with about 10 percent Al_2O_3 and half the available X-sites occupied by calcium. According to Binns, barroisites have compositions that are intermediate between glaucophane and actinolite. The chemistry of specimen LH 14 is quite close to Binns' requirements, but the distribution of ions in its structural sites, as well as its physical and optical characteristics, are those of a mineral intermediate between glaucophane and hornblende. Al^{IV} is much restricted in both glaucophane and actinolite but may reach a value of 2 in hornblendes. Specimen LH 14 has $\text{Al}^{\text{IV}} \simeq 1.0$. In glaucophanes and actinolites the sum of the X-cations is about 2, whereas in hornblendes it is generally greater than 2. In Table 12, ΣX for LH 14 is 2.334. There is therefore a partial filling of the A-sites which are empty in most glaucophanes and actinolites. Also, the refractive indices and the specific gravity of LH 14 are greater than those of glaucophane and actinolite but close to those of hornblendes. The specimen exhibits the substitution scheme proposed by Iwasaki (1960) for Na-Ca-

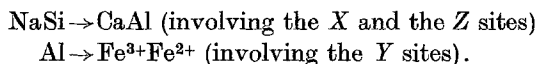
Table 12. Chemical analyses of the green amphiboles

	Weight percentages		Ions on the basis of (O, OH, F) = 24		
	BM1	LH 14		BM1	LH 14
SiO ₂	56.66	47.58	Si	7.879	7.032
TiO ₂	0.01	—	Al ^{iv}	0.121	0.968
Al ₂ O ₃	1.87	6.84	Σ	8.000	8.000
Fe ₂ O ₃	2.12	10.05	Al ^{vi}	0.187	0.224
FeO	4.99	11.63	Ti	0.001	—
MnO	0.39	0.03	Fe ⁺³	0.222	1.117
MgO	19.00	9.90	Fe ⁺²	0.581	1.438
CaO	11.65	7.71	Mn	0.046	0.004
Na ₂ O	1.17	3.62	Mg	3.938	2.181
K ₂ O	0.14	0.40	Σ	4.975	4.964
H ₂ O ⁺	2.10	2.13	Na	0.316	1.037
H ₂ O ⁻	0.05	0.10	K	0.025	0.076
F	nd	0.10	Ca	1.736	1.221
	100.15	100.09	Σ	2.077	2.334
F ≡ O	—	0.04	OH	1.948	2.101
Total	100.15	100.05	F	—	0.047
			Σ	1.948	2.148
S. G.	3.070	3.320			
R.l. α	1.622	1.667			
β	1.639	1.672			
γ	1.642	1.675			
2V	68° (-)	86° (+)			

rich amphiboles. For example, a glaucophane analysis from Table 11 is here compared with the analysis of specimen LH 14



To derive the barroisite from the glaucophane, the following substitutions must be effected:



This is a trend towards hornblende composition and among the schists of Île de Groix, barroisites are found in various stages replacing glaucophane.

Compared with known analyses of other barroisites this specimen contains less titanium and aluminium but more ferric iron. The mineral is believed to have formed under higher $P_{(\text{O}_2)}$ than the glaucophanes. Its Al₂O₃ is less than the 10% proposed by Binns (1967), but this is also true for many known barroisite analyses. Thus it seems that barroisites can and do have variable contents of aluminium most of which is tetrahedrally located. Specimen LH 14 falls into the katophorite

category in the scheme for amphibole nomenclature devised by Whittaker (1968) but like the barroisitic hornblende whose structure was determined by Heritsch *et al.* (1957) its composition plots near the common corner of the katophorite and common hornblende cells (Whittaker, personal communication).

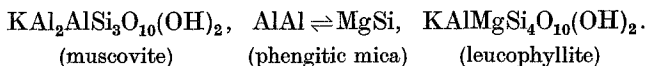
The large measured $2V$ for LH 14 does not agree with Murgoci's (1922) postulate that the optic axial angle of barroisite becomes smaller with increase of iron. More data need to be collected about barroisites before far-reaching conclusions could be drawn on the relationship between the mineral's chemical composition and its optical and physical properties.

The Sheet and the Framework Silicates

These are represented by thirteen white micas, a green mica, seven chlorites and two specimens of plagioclase. Petrographic studies of the host schists (Mankanjuola, 1967), show that the potassic white micas are glaucophane-schist facies minerals that have remained stable in the schists, probably after minor chemical and physical readjustments. The sodic white micas, the chlorites and the plagioclase are post-glaucophanisation minerals, Mankanjuola (1967), formed at the expense of the original chloritoids, blue amphiboles and garnets of the rocks.

The micas are $2M_1$ polymorphs and are represented by eleven potassic white micas, two sodic white micas and a magnesium-rich green mica. Their distribution among the various rock groups of the island is shown in Table 2. In thin sections the potassic white micas are found in contact with garnets, pyroxenes, blue amphiboles, chloritoid, chlorites and albite. They must therefore, at one time or another have been in equilibrium with each of these minerals. The paragenetic differences are, however, in no way reflected in their compositions or other properties. They all exhibit the chemical characteristics of high pressure, low temperature white micas of the glaucophane schist facies. The sodium white micas and the green mica on the other hand reflect the chemistry of their surroundings. The sodium white micas only formed in the potassium-deficient, aluminium-rich chloritoid veins and not in the potassium-rich aluminous schists. The green mica with its high content of MgO occurred only in the magnesian schists.

The chemical and other determined data of the micas are listed in Table 13. Their analyses were recalculated in terms of the general formula $W_2X_2Y_{2-3}Z_4O_{10}$; where $X = K, Na, Ca$ or H_3O^+ ; $Y = Al, Fe^{3+}, Mg, Fe^{2+}, Mn$ or Ti ; $Z = Si$ or Al and $W = OH, O$ or F . From these calculations specimens VM1 and VM2 are shown to be paragonites. The phengitic micas contain higher silica and magnesia than muscovite. Schaller (1950) explained the high Mg content as due to necessary compositional readjustment brought about by increases in Si when a mica changes its composition from close to muscovite to near leucophyllite



This reaction, when it proceeds from left to right, leads to a decrease in $Al^{[IV]}$. It should therefore be pressure dependent and is consistent with the occurrence of phengitic micas under the high pressure conditions of the glaucophane schist facies.

In Table 13, phengite analyses show between 0.06 and 0.35 Na⁺ at the X-sites. Since X-ray studies revealed no inhomogeneity, sodium is regarded as substituting for potassium. Velde (1967 a), however, reported paragonite in association with phengites in two schist specimens from Pointe des Chats, Île de Groix. He claimed that these paragonites originated under the same physical conditions as the phengites. The two specimens of paragonite obtained during the present investigation are vein minerals. Petrographic studies (Makanjuola, 1967) showed that they post-date phengite and glaucophane in the schists and that they were produced in the veins in association with chlorite and magnetite through retrograde reactions involving albite and chloritoid. Under the low temperature, high pressure conditions of the glaucophane schist facies, the composition of a paragonite would be modified towards that of Na-phengite. Unlike in the veins, retrograde reactions involving chloritoid or glaucophane breakdown in the schists gave rise to chlorite and magnetite. Even when a joint breakdown of chloritoid and glaucophane occurs in the same schist, the liberated aluminium and sodium are not known to crystallise into separate paragonite phases, instead they are incorporated into the original phengites of the schists; see Table 13 analyses LH8 and LH10. This might be responsible for the trend towards sericitic composition noticeable among the phengites. The miscibility limit of Na-phengite in K-phengite is not known. To know the maximum solid solution one has to compare the sodium contents of K-phengites in equilibrium with Na-phengites. Results of the sub-solidus solvus regions of paragonite and muscovite are not applicable to these high pressure micas.

X-ray studies show that the two paragonite specimens contain separate muscovite phases. Their contents of K₂O may therefore be taken as a direct measure of the muscovite mixed with them. The basal spacing $d_{(002)}$ of these coexisting micas were measured. Also, using Zen and Albee's (1964) relationship: $d_{(002)} \text{ paragonite} = 12.250 - 0.2634 d_{(002)} \text{ muscovite} \pm 0.006 \text{ \AA}$, the basal spacings of the paragonites were calculated from those of the muscovite coexisting with them and the following values obtained:

Specimen VM1 muscovite = 9.978 Å (measured)
paragonite = 9.623 Å (measured); 9.532 Å (calculated)

Specimen VM2 muscovite = 9.986 Å (measured)
paragonite = 9.624 Å (measured); 9.619 Å (calculated).

Comparison of the measured and calculated values of the basal spacings of the paragonites shows VM2 within and VM1 outside the predicted deviation. McNamara (1965) who observed similar disagreements concluded that the value $0.2634 d_{(002)}$ muscovite was always too small.

The phengites have high refractive indices on account of the replacement of Al^[IV] by Si and Al^[VI] by Fe²⁺ and Mg (Deer *et al.*, 1963), and their specific gravities are directly related to Σ (Fe²⁺, Fe³⁺, Mn and Ti).

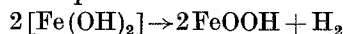
The Chlorites. Seven 14 Å polymorph chlorite specimens were analysed. The host rocks from which they were obtained are given in Table 2 and their chemical, physical, X-ray and optical properties are listed in Table 14. On account of the difficulties encountered in the accurate determination of water in iron-rich

Al	3.035	3.091	2.939	2.967	3.059	3.014	2.927	3.090	2.774	3.018	3.066	3.186	3.926	3.873	Al
Fe ⁺³	0.321	0.219	0.361	0.337	0.387	0.353	0.208	0.142	0.144	0.287	0.346	0.257	0.054	0.049	Fe ⁺³
Fe ⁺²	0.133	0.082	0.108	0.153	0.123	0.117	0.242	0.238	0.116	0.136	0.145	0.231	0.017	0.026	Fe ⁺²
Mn	—	—	—	—	0.001	0.001	0.003	—	0.003	0.001	0.002	0.001	0.002	0.002	Mn
Mg	0.414	0.566	0.537	0.554	0.328	0.578	0.489	0.473	1.048	0.681	0.346	0.301	0.005	0.025	Mg
Ti	0.031	0.029	0.035	0.036	0.004	0.004	0.025	—	0.005	0.024	0.004	0.039	0.005	0.010	Ti
Σ Y	3.934	3.987	3.980	4.047	3.902	4.067	3.894	3.943	4.090	4.147	3.909	4.015	4.009	3.985	Σ Y
Na	0.130	0.081	0.150	0.300	0.354	0.168	0.147	0.166	0.091	0.145	0.218	0.056	1.419	1.408	Na
K	1.654	1.663	1.674	1.684	1.631	1.763	1.715	1.642	1.670	1.863	1.683	1.672	0.553	0.302	K
Ca	0.079	0.006	0.007	0.011	0.032	0.019	0.033	0.101	0.024	0.006	0.016	0.004	0.066	0.141	Ca
Σ X	1.863	1.750	1.831	1.995	2.017	1.950	1.895	1.909	1.785	2.014	1.917	1.732	2.038	1.851	Σ X
OH	4.052	4.229	4.109	3.967	3.889	3.735	4.310	4.070	4.342	3.496	4.304	3.914	3.765	4.143	OH
F	0.182	0.138	0.189	0.043	0.374	0.161	0.265	0.210	—	0.052	0.230	0.211	0.184	0.182	F
Σ	4.234	4.367	4.298	4.010	4.263	3.896	4.575	4.280	4.342	3.548	4.534	4.125	3.949	4.325	Σ

White micas from the aluminous schists (I), the calcareous schists (II), the magnesian schists (III), the basic schists (IV), and veins (V).

chlorites, a calculation of the chemical formulae on an anhydrous basis gives better cation values (Makanjuola, 1967). The classification scheme proposed by Phillips (1964) has been followed.

All the analysed chlorites show, by their contents of octahedral and tetrahedral aluminium, the common characteristics of greenschist and blueschist facies chlorites (Foster, 1962; McNamara, 1965). In these specimens, the tetrahedral aluminium which may reach a value of 2 out of 4 occupied tetrahedral sites in silicates (Nelson and Roy, 1958) varies between 1.003 and 1.450. In most of the analyses the balancing octahedral aluminium is smaller than the corresponding tetrahedral aluminium. This is a general tendency among chlorites (Foster, 1962). Electrostatic balance is restored by adding Fe³⁺ to Al^[VI] but by so doing the sum of octahedral R³⁺ becomes greater than Al^[IV]. Some of the ferric iron is therefore considered to be an oxidation product of ferrous iron, e.g.



(Holzner, 1938).

The chlorites have crystallised in equilibrium with ilmenite, sphene and rutile. The small quantities of titanium in the analyses (0.004 to 0.058) are therefore not considered to be due to impurities. McNamara (1965) plotted the values $\{(R^{3+} + R^{4+})^{[VI]} - (R^{3+})^{[IV]}\}$ of his chlorites against the occupancy of octahedral sites. A better fit was obtained when the values for titanium were added and allowances made for its tetravalent nature.

The chlorites show substitutions of the type MgSi → Al^[VI]Al^[IV]. But by far the most important cations are the pair Mg and Fe²⁺. These occupy between 8 and 9 out of 12 octahedral sites. The ratio Fe²⁺:Mg varies between 0.19 and 1.74. Manganese, a minor cation, shows

Table 14. Chemical analyses of the chlorites

	I	III	IV	V			
	LH5	BM1	LM3	VM1	VM2	VM3	VM4
SiO ₂	25.16	30.54	25.75	25.04	24.96	25.50	27.27
TiO ₂	0.36	0.03	—	0.20	0.17	0.10	0.02
Al ₂ O ₃	17.80	20.36	19.70	18.81	21.39	21.33	19.71
Fe ₂ O ₃	7.24	3.04	3.94	6.91	4.26	2.91	4.85
FeO	25.00	8.83	23.27	28.95	29.26	21.54	17.34
MnO	0.59	0.20	0.06	0.35	0.26	0.28	0.13
MgO	11.77	24.83	15.34	9.66	9.37	17.40	19.76
CaO	0.09	0.13	0.10	0.02	0.16	0.08	0.04
Na ₂ O	0.05	0.03	0.11	0.05	0.09	0.02	0.10
K ₂ O	0.20	0.50	0.18	0.24	0.28	0.07	0.10
H ₂ O ⁺	11.26	11.33	11.29	9.70	9.60	11.25	11.10
H ₂ O ⁻	0.10	0.12	0.20	0.12	0.22	0.25	0.10
Total	99.62	99.94	99.94	100.05	100.02	100.73	100.52
S. G.	3.006	2.812	2.980	3.080	3.120	2.906	2.910
α	—	—	1.627	—	—	1.623	1.617
β	1.630	1.614	1.629	1.642	1.643	1.626	1.619
γ	1.632	1.616	—	1.644	1.644	—	—
2V	Small (-)	32° (-)	Small (+)	Small (-)	Small (-)	43° (+)	44° (+)
b (Å)	9.284	9.260	9.268	9.293	9.288	9.264	9.257
d_{001} (Å)	14.16	14.14	14.12	14.12	14.13	14.18	14.22
Chlorite ions calculated on the theoretical basis of (O, OH) = 36 per unit formula							
Si	5.541	5.995	5.417	5.533	5.481	5.100	5.572
Al	2.459	2.005	2.583	2.467	2.519	2.900	2.428
Σ	8.000	8.000	8.000	8.000	8.000	8.000	8.000
Al	2.057	2.706	2.301	2.432	3.018	2.128	2.318
Ti	0.058	0.003	—	0.033	0.028	0.016	0.004
Fe ⁺³	1.172	0.448	0.624	1.150	0.705	0.437	0.746
Fe ⁺²	4.500	1.450	4.094	5.349	5.375	3.603	2.964
Mn	0.107	0.033	0.011	0.065	0.049	0.048	0.022
Mg	3.776	7.266	4.809	3.181	3.067	5.186	6.018
Na	0.021	0.026	0.046	0.021	0.040	0.007	0.039
Ca	0.054	0.125	0.048	0.069	0.079	0.017	0.027
Σ	11.745	12.057	11.933	12.300	12.361	11.442	12.138
OH	16.164	14.835	15.841	14.297	14.064	15.009	15.127

Chlorites from the aluminous schists (I), magnesium schists (III), basic schists (IV) and veins and pegmatitic bodies (V).

higher concentrations in chlorites that are derived from chloritoids and garnets; two such manganese-enriched specimens are LH5 and VM1.

The specific gravity and the refractive indices of the chlorites are functions of their heavy cation contents as there is a straight line relationship between these and Σ (Fe³⁺, Fe²⁺, Mn and Ti). Their b cell edges also vary with their iron (Fe²⁺) and manganese contents.

The chlorites appear to have crystallised under the physical conditions of the greenschist facies through retrograde decomposition of the original ferromagnesian minerals of the schists. A pronounced chemical control by the parent materials from which they evolved are noticeable in their chemical compositions. The clinocllore is from a magnesian schist, the ripidolites from non-chloritoid bearing veins, the sheridanite from a glaucophane schist, and the thuringites and the aphrosiderite from a chloritoid schist and chloritoid veins. The chlorites of the quartz-albite-chlorite vein rocks, that Brière (1920) identified through petrographic studies as clinochlores, have been found to be ripidolites.

The Framework Silicate. Two plagioclase specimens were obtained from a basic schist and a calcareous schist. Their chemical analyses, density and optical data are listed in Table 15. They both have albite composition with very low anorthite and negligible orthoclase molecules.

Table 15. Chemical analyses of the plagioclases

	KM4	LM3		KM4	LM3
SiO ₂	67.11	68.08	Ions on the basis of O = 8 per unit formula		
Al ₂ O ₃	20.23	20.30	Si	2.952	2.966
CaO	0.66	0.32	Al	1.049	1.043
Na ₂ O	11.40	11.57	Σ	4.001	4.009
K ₂ O	0.24	bd.	Na	0.972	0.978
Total	99.64	100.27	K	0.013	—
S. G.	2.620	2.612	Ca	0.031	0.015
R. l. α	1.524	1.525	Σ	1.016	0.993
β	1.532	1.530	Ab	96	98
γ	1.535	1.535	An	3	2
2V γ	74°	75°	Or	1	—

There is an abundance of petrographic evidence that the albites, chlorites and paragonites are the latest additions to the minerals of the crystalline rocks of Île de Groix. They usually occur in close association with areas of glaucophane breakdown (chlorite and albite) or chloritoid breakdown (chlorite and paragonite). Their chemical compositions also confirm the foregoing conclusion. The high aluminium and ferrous iron content of the chlorites and the low anorthite content of the albites are all related to grade. These three minerals belong to the greenschist facies of metamorphism; they are not blueschist facies minerals.

Conclusions

Partitioning of Elements among Coexisting Mineral Phases

An appreciable number of the coexisting minerals of the basic schists were analysed. One of the striking features is the very narrow range of compositional variations among members of each group of mineral species, in spite of the influence of the host rock compositions. These minerals must therefore have evolved under closely

related metamorphic conditions. The partitioning of the major oxides also provides a chemical explanation for the common textural features of minerals associating in composite grains. For example, many thin sections of the basic schists contain grains in which glaucophane, epidote, rutile and barrosite are in the process of replacing pyroxene, garnet, sphene and glaucophane, respectively. Pyroxene and glaucophane, apart from both being chain silicates, contain about equal amounts of Al_2O_3 , SiO_2 , MgO and Na_2O and they are also equally low in MnO . Garnet and epidote are calcium-rich orthosilicates with about the same contents of SiO_2 , Al_2O_3 and MnO . Glaucophane and barrosite are double-chain, hydroxyl-bearing sodium, magnesium silicates. These mineral replacements are, however, neither simple nor volume for volume transformations. The structures and chemistry of these minerals, though similar, are nevertheless different in many essential details. The necessary readjustments must therefore have entailed a complex reaction among garnet, pyroxene and sphene to produce the assemblage glaucophane, epidote and rutile.

Influence of the Host Rock Composition on the Mineral Chemistry

Many of the diagrams in Fig. 5 show strong linear correlations between the ratios Fe^{2+} or Mg or $\text{Mn}/(\text{Fe}^{2+} + \text{Mg} + \text{Mn})$ and $\text{Fe}^{3+}/(\text{Fe}^{3+} + \text{Al} + \text{Ti})$ for the minerals and those for the rocks. They illustrate the dependence of the mineral compositions on the host rock bulk composition. It is not possible to separate, on Fig. 5A, the blue amphiboles of the metabasalts where Mg and Al predominate from those of the metasediments where Al and Fe^{2+} predominate. In the blueschist facies metamorphism, the end-member molecules albite and chlorite will react to give a blue amphibole. The two amphiboles with the highest $\text{Fe}^{2+}/(\text{Fe}^{2+} + \text{Mg} + \text{Mn})$ ratios are both from rocks in which garnet has not appeared, i.e. low grade schists. This is consistent with the derivation of sodic amphiboles from albite and low grade Fe-Al chlorites; at higher grades the chlorites react with quartz to give almandine and Mg -rich chlorites. In these blueschists the $\text{Fe}^{2+}/(\text{Fe}^{2+} + \text{Mg} + \text{Mn})$ ratios of amphiboles reflect the grade of the original schists in which the mineral formed.

The $\text{Fe}^{3+}/(\text{Fe}^{3+} + \text{Al} + \text{Ti})$ correlation for the epidotes, though suggestive of a definite trend, shows a great deal of scatter. As already stated, the basic schist epidotes were derived through garnet retrogression such that the equation $\text{Ca}_2\text{Al}_2\text{Fe}^{3+}\text{Si}_3\text{O}_{12}(\text{OH}) \rightleftharpoons \text{Ca}_2\text{Fe}^{2+}\text{Al}_2\text{Si}_3\text{O}_{12} + \frac{1}{2}\text{H}_2\text{O} + \frac{1}{2}\text{O}_2$, proceeded from right to left. This increase in the partial pressure of oxygen and water would be reflected in the $\text{Fe}^{3+}:\text{Al}$ ratios of the epidotes more than of the rocks, and should mask the influence of the bulk rock composition on Fe^{3+} content of the epidotes.

The chloritoids reflect by their compositions the $\text{Mg}:\text{Fe}^{2+}$ and the $\text{Al}:\text{Fe}^{3+}$ ratios of the aluminous schists; see Figs. 5C and 5D. The influence of mineral assemblages on element partitioning is on the other hand illustrated by the $\text{Mn}/(\text{Mn} + \text{Fe}^{2+} + \text{Mg})$ ratio correlation, Fig. 5E. The dots in this figure represent chloritoids coexisting with garnets; these show good correlation with the bulk rock composition. The plus signs represent chloritoids from non-garnet-bearing aluminous schists, i.e. lower grade rocks. As there are no other Mn -phases to share the manganese, this element becomes concentrated in the chloritoids alone. The amount of chloritoid in the rocks will be inversely proportional to its Mn content. This will suggest that chloritoids must readjust their compositions at the appear-

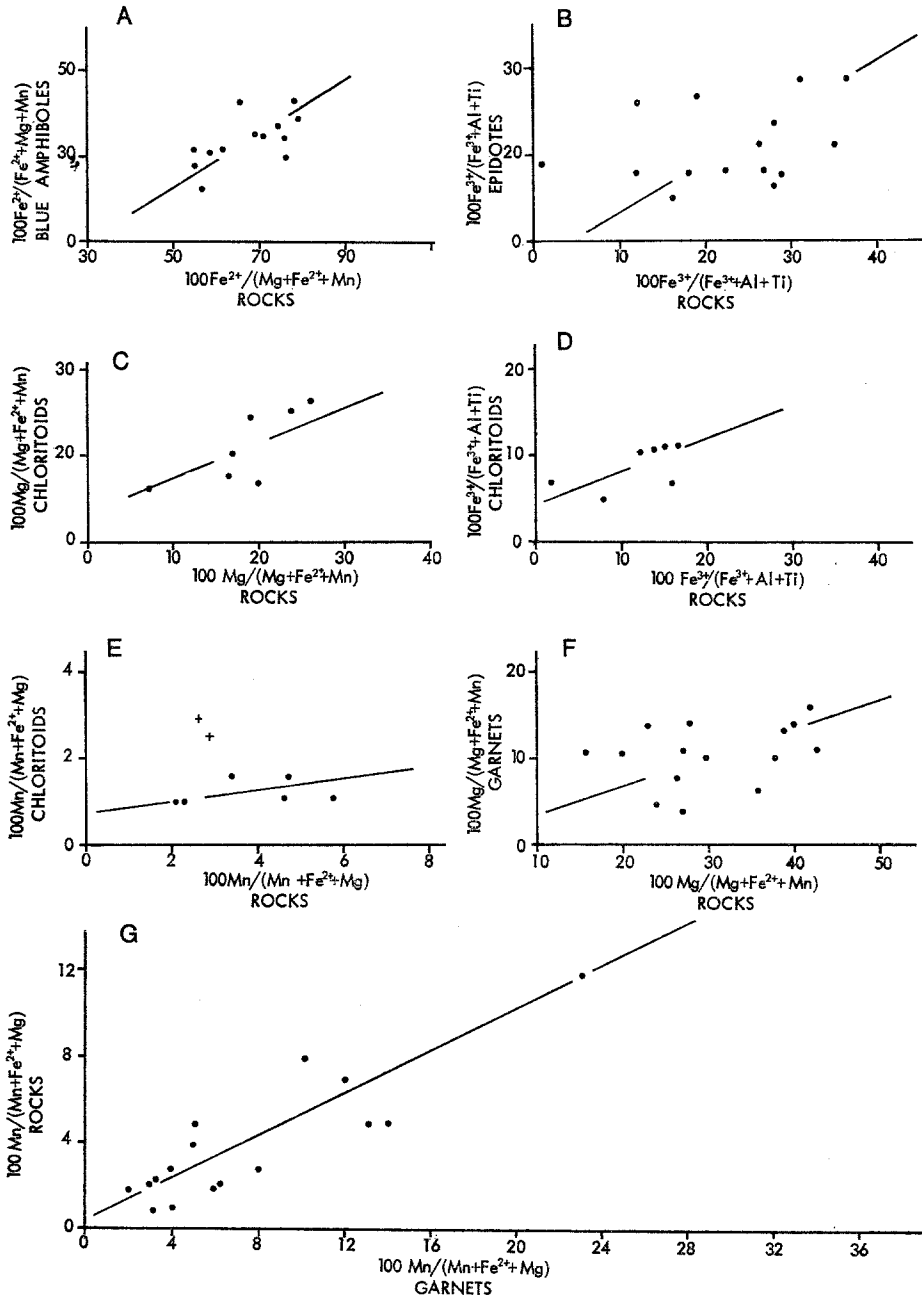


Fig. 5. Correlation diagrams of the rock and mineral oxides

ance of garnets in a rock undergoing progressive metamorphism. It also implies that ottrelite, the Mn-rich chloritoid, would be confined to the lower subfacies of the greenschist facies in regionally metamorphosed rocks. Similarly, in non-chloritoid-bearing garnet schists manganese is preferentially concentrated in the

garnets, with the result that $Mn/(Mn + Fe^{2+} + Mg)$ ratios are commonly higher in the minerals than in the rocks; see Fig. 5G. The scatter observable in Fig. 5F is attributed to the unequal lowering of the magnesium contents of the garnets with the appearance of glaucophane in the schists.

Reconstruction of the Mineral Paragenesis

The minerals recognised in the different rock types and from which the different species were obtained that form the subject of the present work were listed in paragenetic association by Mekanjuola (1967, p. 37, Table 2.4). The calcareous and the magnesian schists failed to exhibit the same multi-mineral assemblages observed in the aluminous and the basic schists, on account of their more restricted chemical compositions. Also, the original pyrogenic minerals of basaltic rocks, from which the basic schists were evolved, are less stable during metamorphism. Examples of assemblages present in the aluminous and the basic schists are as follows:

the basic schists:

pyroxene + garnet + sphene;
 pyroxene + garnet + sphene + glaucophane + epidote + rutile + phengite;
 garnet + glaucophane + barroisite + epidote + rutile;
 epidote + glaucophane + chlorite + albite + magnetite + phengite;
 epidote + chlorite + phengite + ilmenite + albite + calcite;

the aluminous schists:

garnet + phengite + shimmer aggregate;
 garnet + glaucophane + chloritoid + phengite + sphene + rutile;
 epidote + glaucophane + phengite + pseudomorphs (after lawsonite ?);
 chloritoid + chlorite + phengite + magnetite;
 glaucophane + chlorite + albite + magnetite.

Most of the foregoing are undoubted disequilibrium assemblages. For example, chlorite and albite found in association with glaucophane in some of the assemblages ought to react to give a blue amphibole. Occurrence of "armoured" minerals and composite grains are evidence that most of the assemblages are products of arrested reactions such as could have resulted from paragenetic readjustment under waning temperature conditions. The listed minerals may therefore be reconstructed into the following equilibrium assemblages:

Basic schists	Aluminous schists
Garnet + pyroxene + sphene	garnet + muscovite + (kyanite ?) + sphene
↓	↓
glaucophane + epidote + rutile + phengite	chloritoid + glaucophane + epidote + phengite + rutile + (lawsonite ?)
↓	↓
barroisite + epidote + phengite	↓
↓	muscovite + chlorite + quartz + calcite
calcite + chlorite + albite + muscovite + actinolite	

Certain pseudomorph porphyroblasts have been interpreted as "ghost" relics of lawsonite (Makanjuola, 1967), and the shimmer aggregates as the alteration products of kyanite (Chinner, 1967). The arrows in the table show the direction of mineral succession with decreasing temperature. Garnet-pyroxene assemblages are common in eclogites and granulites. For example, on the diagram showing possible phase relations between basalt, amphibolite, granulite and eclogite postulated by Essene and Fyfe (1967), the two last named rocks have a common interface, and their mineralogy must naturally show some similarities. The analyses of Île de Groix pyroxenes fall in the field of eclogite pyroxenes. The garnets are, however, poorer in pyrope and richer in almandine and grossular than most eclogite garnets; they are similar in composition to the garnets from glaucophane schist terrains. On the strength of the foregoing it is believed that eclogite was produced among the basic schists of the island. The pyrope content of the garnets and the Tschermak components of the pyroxenes must have become depressed with the appearance of glaucophane and epidote among the minerals of the basic schists. Such views are consistent with the observations of Coleman *et al.* (1965) that omphacites from the blueschist eclogites are richer in jadeite and co-exist with garnets containing less pyrope (< 20 molecular percent) than is typical in eclogites formed under different temperature and pressure conditions. The close field association of eclogitic and blueschist assemblages in Île de Groix is of special petrological and petrogenetic interest which is being examined (Makanjuola, in preparation).

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