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ORIGIN AND EVOLUTION OF THE PIÊN MAFIC-ULTRAMAFIC COMPLEX, SOUTHERN BRAZIL[1]

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ABSTRACT

A suite of mafic and ultramafic rocks exposed at Piên, SE Brazil, were affected by at least three recrystallization episodes, the earliest one producing high-grade granulites about 2,000 m.y. ago (Transamazonic cycle), later partly converted into amphibolite-facies lithologies during the Brasiliano cycle (650 m.y.) and serpentinites and talc shists. The predominant ultramafic rocks are meta-olivine pyroxenites, meta-websterites, meta-norites, serpentinites (mainly former harzburgites) and magnesian schists (mainly former orthopyroxene-rich rocks). The mafic suite presents granoblastic meta-gabbros, amphibolites and amphibole gneisses, compositionally akin to high-alumina basalts and olivine tholeiites. Bulk chemical composition shows a trend of Fe enrichment and probable continuity in chemical tendencies among ultramafic and mafic members. Mineral analyses of olivines, spinels and especially orthopyroxenes and clinopyroxenes in ultramafitites show a chemical trend compatible with their derivation from recrystallized differentiated (layered?) complexes. Compositional variations of various coexisting mineral assemblages suggest that the granulitic high-grade re-equilibration occurred at temperatures from about 750 to 880°C and minimum pressures of about five to seven kb. Present petrography, outcrop patterns and chemical data suggest that the main primary minerals in ultramafitites were Mg-rich olivine, Al-orthopyroxene and Al-subcalcic clinopyroxenes. Experimental studies of high-pressure crystallization of basalts show that such minerals can fractionate away from an olivine tholeiite magma—at a pressure interval from about seven to about 12 kb.—yielding a series of ultramafic rocks and a residue which on many accounts matches the composition of the Pien metagabbros and associated rocks.

INTRODUCTION

The basement areas in the states of Santa Catarina and Paranā, southeastern Brazil (fig. 1), are lithologically characterized by the presence of low- to medium-grade schists constituting distinctive folded belts (Hasui et al. 1975) and a product of Braziliano tectonometamorphic events (650 m.y.). Strips of folded belts are separated by the Taboleiro Group, now mapped as the so-called Joinville median massif (Hassui et al. 1975). Its rocks are divided into two lithologic units. One consists mainly of migmatites with some rare interlayered granulitic lenses, while the second unit, mainly found in the Piên area and the nearby Piçarras-Barra Velha region (fig. 1), exhibits ultramafic and associated types as main lithologies. Other geologic units in the area are intrusive granites, early Paleozoic effusive and supracrustal rocks, scattered remnants of Devonian sandstones and a few NW-trending diabase dykes geologically related to the Paranā basalts (for details, see Girardi 1974, and bibliography therein). It is the purpose of this study to present data on the Piên occurrence of ultramafic and mafic rocks, with special emphasis on the possible relationship between them.

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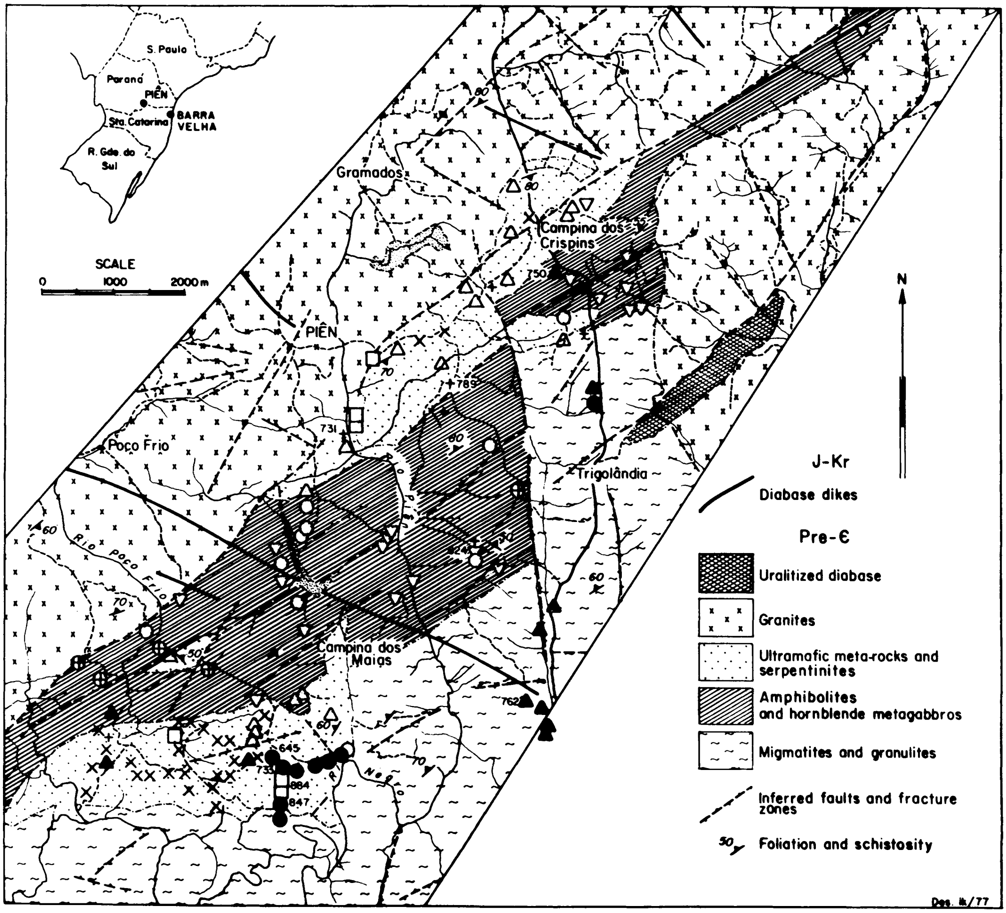


FIG. 1.—Geology of the Piên area. Numbers represent temperatures (in °C) computed according to a two-pyroxene thermometer (Wells, 1977). *Piên rock symbols.* Open squares: meta-olivine pyroxenites; crosses: meta-websterites; upright triangles: serpentinites; x: magnesian schists; large dots: meta-norites; downright triangles: meta-gabbros; open circles: amphibolites; filled triangles: granulites; crosses within a circle: amphibole gneisses.

GENERAL GEOLOGY AND PETROGRAPHY OF THE PIÊN ULTRAMAFIC AND MAFIC ROCKS

The present account on the local Piên geology is based on published data and on detailed work by one of the authors (Girardi 1974). Figure 1 presents the main geological features of the ultramafic bodies and surrounding rocks. Outcrops are observed mainly as unconnected boulders and thus mapped contacts are, with few exceptions, only inferred or approximate at best.

Outcrops of Taboleiro migmatites are very scarce, and present interlayered small lenses of granulitic rocks. Predominant minerals are andesine, mesoperthites and quartz in the rare acid granulites, while the other granulite types show plagioclase (An_{33} in intermediate granulites to An_{65} in basic varieties), salitic pyroxene, hypersthene, opaques, and sometimes quartz and alkali feldspar. The ultramafic and associated rocks were marked on the map (fig. 1) as two different units, the one predominantly ultramafic, the other predominantly mafic.

The mafic unit or "gabbroic" suite is exposed as a lenticular body with granulitic hornblende meta-gabbros, foliated amphibolites and amphibole gneisses as main rock types. The predominant hornblende metagabbros are granoblastic leuco- to mesocratic rocks, with plagioclase (An_{30}–An_{72}), amphiboles and a salitic pyroxene as main minerals. Amphibole is observed as an earlier olive-brown variety (ferrous pargasite to hornblende, see Leake 1968), mainly after clinopyroxene, and later green actinolite replacing the brown type. Amphibolites are melanocratic rocks with amphiboles (brown hornblende and late green actinolite) and plagioclase (An_{35}–An_{60}) as main minerals and minor amounts of relict clinopyroxene. In amphibolitic gneisses, granoblastic felsic bands (plagioclase- and quartz-rich) alternate with lepidoblastic ones in which olive and green amphiboles and some intermediate plagioclase and clinopyroxene are the principal constituents. Accesory minerals in mafic rocks are garnets, opaques, apatite, sphene, and minor quantities of secondary minerals.

The ultramafic unit appears in the field as a petrographically complex suite of rocks. Lithologically, the predominant types are meta-olivine pyroxenites (including some harzburgites, appr. 9% of observed outcrops), meta-websterites (with some ortho- and clinopyroxenites, appr. 6 %), melanocratic meta-norites (12%), and especially serpentinites (36%) and magnesian schists (36%). Meta-olivine pyroxenites present orthopyroxene as the predominant mineral, followed by about 10% of magnesian olivine and minor clinopyroxene; important accesory minerals are spinel (1–8 %), some opaques and varying quantities of secondary serpentine (mainly after olivine). Meta-websterites are rocks with clinopyroxenes (salite or diopside) and bronzite; amphiboles (pargasite-hornblende and actinolite, Leake 1968) are always present—sometimes making up 30% of the rock—and are secondary after both pyroxenes. The meta-norites have as main constituentes: bronzite (up to 5%); augitic clinopyroxene and plagioclase (An_{45} to An_{65}, rarely exceeding 15% modal percent); and hercynite (2–7%). Opaques and a secondary amphibole are always present in minor quantities. One unusual meta-norite is devoid of clinopyroxene and presents 10% of sapphirine, partly in symplectitic intergrowth with orthopyroxene. An unusual rock (sample MP 260) was found as an isolated boulder; it is a garnet pyroxenite, with about 35% garnet, 25% clinopyroxene 2% orthopyroxene, 3% slightly zoned plagioclase, 1% green spinel, and secondary amphibole; accesories are titanite and opaques (see section on geobarometry for a discussion). Lizardite and chrysotile are usually the main minerals of serpentinites. Antigorite, identified in X-ray patterns, is subordinate and, in some specimens, seen as "bastitic" intergrowths probably after orthopyroxene. Magnetite-hematite grainlets form border zones to the serpentine pseudomorphs; chromite, talc and dolomite, together with rare relictic minerals (olivine, some orthopyroxehe and a few spinel grains) are accesory constituents in some serpentinites. The magmesian schists present a distinct metamorphic fabric. Talc is the main mineral, while chlorite and tremolite are minor constituents and often appear as porphyroblasts.

CONTACT AND STRUCTURAL RELATIONSHIPS

The several rock types constituting the ultramafic unit are found as scattered and unconnected boulders and outcrops. The two lenticular ultramafic areas (fig. 1) show a great variation in petrography from one outcrop to the next with a tendency to a large-scale zonation, as indicated by the fact that serpentinites and magnesian schists on the one hand, and meta-norites on the other, are predominant in some areas. No field evidence was observed to indicate that the ultramafic unit is a rhythmically layered body.

The contacts between the mafic and ultramafic units are in part clearly tectonic, but in most cases no direct observation is possible. Locally, an intermingling of outcrops was observed, for example, scattered and minor lenses of meta-websterites and serpentinites within gabbroic rocks. Both petrography and spatial association suggest a genetic link between the mafic and ultramafic units, although no geological relationships could be established.

There are no clear exposures to indicate the exact nature of the non-tectonic contacts of the Taboleiro migmatites with the mafic-ultramafic rocks.

TEXTURES

Among ultramafic rocks, only the magnesian schists present a clearly oriented fabric. Meta-olivine pyroxenites, meta websterites and meta-norites are in general uniformly granoblastic, with equilibrium boundaries between adjacent grains. However, there is in most rocks, an overprint of a secondary nature. Serpentine and chlorite replace mostly olivines and pyroxenes. In most meta-norites, meta-olivine pyroxenites and meta-websterites, two generations of amphiboles are observed: the earlier one is a brownish-olivine pargasite (to hornblende), while the later one, frequently replacing earlier pyroxenes and hornblende-pargasite, is a green actinolite. The brownish amphibole exhibits in part straight boundaries when contacting pyroxene grains, but in most cases textures suggest a replacement relationship. The spinels, although xenomorphic, seem to be in textural equilibrium with pyroxenes and olivines. In serpentinites, spinel is rare and relict. Chrome-rich varieties are very rare; no chromite pods were observed. Serpentinites show several generations of serpentines, with earlier lizardite cut by chrysotile veinlets, but the boundaries of the formerly predominating olivine crystals—as recognized by magnetites grainlets—clearly depict a previous granular equilibrium texture. Some textures suggest the former presence of banding and layering on a micro-scale (but see Evans 1977, p. 401 ff., 429): for instance, rather regular banding produced by an alternation of plagioclase and two-pyroxene layers is shown by some meta-norites.

Textures in rocks of the gabbroic suite are either granoblastic or oriented. Meta-gabbros present an imprint of late amphibole crystallization. Brown-olivine hornblende (to pargasite) is mostly a late mineral after clinopyroxene; it is rarely observed as better-formed crystals, apparently coexisting with plagioclase. A younger actinolite is either fibrous or forms continuous borders to brown amphibole. Garnet is rare. Chlorites, epidotes and biotites are retrograde minerals. Amphibolites and amphibole gneisses, with a more oriented fabric, show similar mineralogical features; in most of them, clinopyroxene is relictic and being substituted by brown amphibole.

Most granulites within the regional migmatites are granoblastic rocks with pyroxenes and plagioclase. Retrograde reactions convert minor parts of the pyroxenes into serpentine, talc and a green hornblende.

RADIOMETRIC AGES

Rb/Sr and K/Ar age determinations (Girardi et al. 1974) indicated that the Piên mafic-ultramafic complex is, with the similar Barra Velha, Cana Brava and Barro Blanco occurrences, one of the oldest Brazilian massifs of its kind. It presents a complicated polymetamorphic history. The oldest detectable age reflects the action of a 2000 m.y. event related to a granulitic metamorphism, occurring during the Transamazonic cycle, which extensively reworked large zones of the South American cratonic areas. Migmatization at Pien is in part very old, a sequel to the closing phases of the Transamazonic cycle, but mostly a much younger syntectonic by-product of the Brasiliano cycle (peak activity at about 650 m.y.) (Cordani et al. 1973).

CHEMICAL DATA

Results will be here presented on bulk chemistry of rocks (obtained by means of wet chemical, optical spectroscopy and X-ray fluorescence methods) and mineral composition (spot analyses with the microprobe), to be discussed later on (for details, see Girardi 1974). Averages of rock analyses are found in table 1. Tables 2 to 6, presenting individual rock and mineral analyses, are deposited at ASIS/NAPS.

(*a*) *Bulk chemistry.*—Averages or ranges in composition of the most important Piên rock types, on a water-free basis, are presented in table 1 (cols. 1 to 6). Most commonly, mafic-ultramafic massifs present chemical gradations between their extreme members (e.g., Bowes et al. 1970; Jackson and Thayer 1972). In the Piên case, there is on the whole no break in trend between the two groups of rocks, but chemistry (as well as petrography) clearly indicates the lack of intermediate "missing links", a fact well expressed in several variation diagrams (e.g., figs. 2, 3 and 4).

TABLE 1[1]

AVERAGE COMPOSITION OF PIÊN ROCKS AND OF RELATED PETROGRAPHIC TYPES

	1	2	3	4	5	6	7	8	9	10	11	12	13	14	15	16	17	18	19	20
SiO_2	45.87	47.59	50.60	42.6 -52.0	48.96	53.05	44.95	45.07	45.48	40.66	43.38	45.60	41.3	44.9	41.7	45.31	46.14	47.0	47.3	47.99
TiO_2	0.03	0.07	0.29	0.08 -0.70	1.12	0.70	0.12	0.19	0.71	0.07	0.08	0.02	0.01	0.02	0.006	0.02	0.05	2.0	2.9	3.05
Al_2O_3	1.06	7.24	9.65	3.5- 14.6	17.24	17.26	4.13	4.19	3.56	0.85	1.05	0.74	0.39	0.94	0.66	1.71	1.97	13.1	16.1	15.00
FeO*	7.49	8.11	9.85	7.6- 12.8	11.97	8.85	7.97	8.26	8.50	9.19	7.88	7.5	9.0	8.1	8.5	7.01	9.09	11.0	10.8	13.66
MnO	0.08	0.02	0.10	0.02- 0.2	0.17	0.11	0.13	0.11	0.14	0.14	0.13	0.15	0.14	0.12	0.13	0.09	0.13	0.2	0.3	0.20
MgO	45.24	34.55	21.06	17.0 -28.6	6.56	5.47	39.40	39.47	37.75	48.83	47.00	45.0	48.5	44.9	48.2	45.01	41.53	14.6	7.2	6.73
CaO	0.03	2.27	7.48	4.8- 17.1	9.84	9.61	2.89	2.43	3.10	0.26	0.40	0.9	0.36	0.92	0.46	0.55	0.81	10.2	12.4	9.46
Na_2O	0.19	0.14	0.89	0.2- 0.8	2.86	4.35	0.32	0.22	0.57	—	0.08	0.01	0.02	0.02	0.05	0.18	0.26	1.7	2.5	2.63
K_2O	0.01	0.01	0.08	0.01- 0.6	1.03	0.53	0.07	0.05	0.13	—	—	0.08	0.22	0.05	0.05	0.02	0.015	0.1	0.1	0.75
P_2O_5	—	—	—	—	0.25	0.07	0.02	0.01	0.06	—	—	0.05	0.06	0.03	0.25	0.10	0.01	0.2	0.2	0.53
N	4	3	4		6	3						3	5	2	2	53	16			

NOTE: All values on a water-free basis, recalculated to 100%. *Iron oxides recomputed as FeO*, ferrous iron oxide. N is number of analyses for averages.

1: Average Piên serpentinite.
2: Average Piên meta-olivine pyroxenite.
3: Average Piên meta-norite; sample 96 is excluded.
4: Variations in Piên meta-websterites.
5: Average Piên meta-gabbro.
6: Average Piên basic granulite.
7: St. Paul's rock (Medaris, 1972).
8: Lizard intrusion (Medaris, 1972).
9: Pyrolite (Medaris, 1972; Ringwood, 1975).
10: Average dunite, New Zealand (Challis, 1965).
11: Average harzburgite, New Zealand (Challis, 1965).
12: Average harzburgite, Red Mountain, California (Himmelberg and Coleman, 1968).
13: Average dunite, Red Mountain, California (Himmelberg and Coleman, 1968).
14: Average harzburgitic peridotite, Burro Mountain, California, (Burch, 1968; cf. also Loney et al., 1971).
15: Average dunite, Burro Mountain, California, (Burch, 1968; af. also Loney et al., 1971).
16: Average serpentinite from ultramafic rocks (Lockwood, 1971).
17: Average serpentinite from Mid-Atlantic Ridge (Miyashiro et al., 1969).
18: Synthetic olivine tholeiite, composition selected for high-pressure melting experiments (Green and Ringwood, 1967).
19: Residual liquid after fractionating from olivine tholeiite, col. 18, 30 weight% of crystals at 9 kb., 1230°C (Green and Ringwood, 1967). See text.
20: Snake River olivine tholeiite, selected for high-pressure melting experiments (Thompson, 1974). See text.

[1] Tables 2-6 available from ASIS/NAPS, c/o Microfiche Publications, P.O. Box 3513, Grand Central Station, N.Y., N.Y., 10017.$3.00 prepaid for microfiche, or $5.00 for photocopy (non U.S. orders add $3.00).

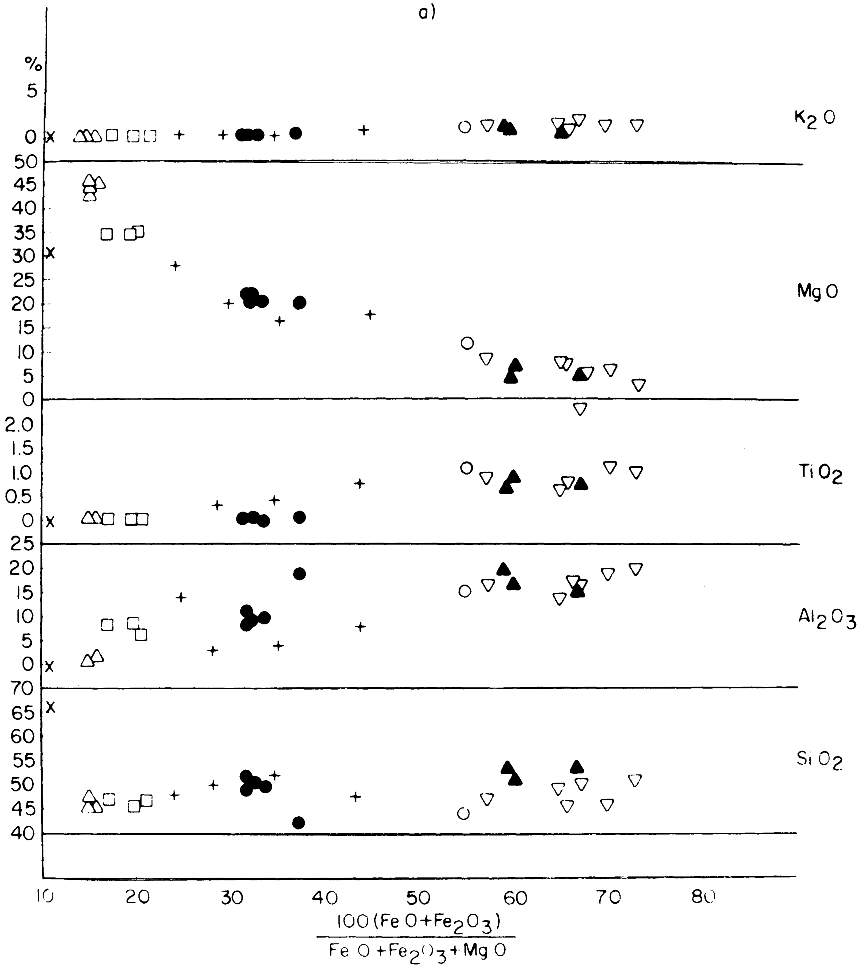
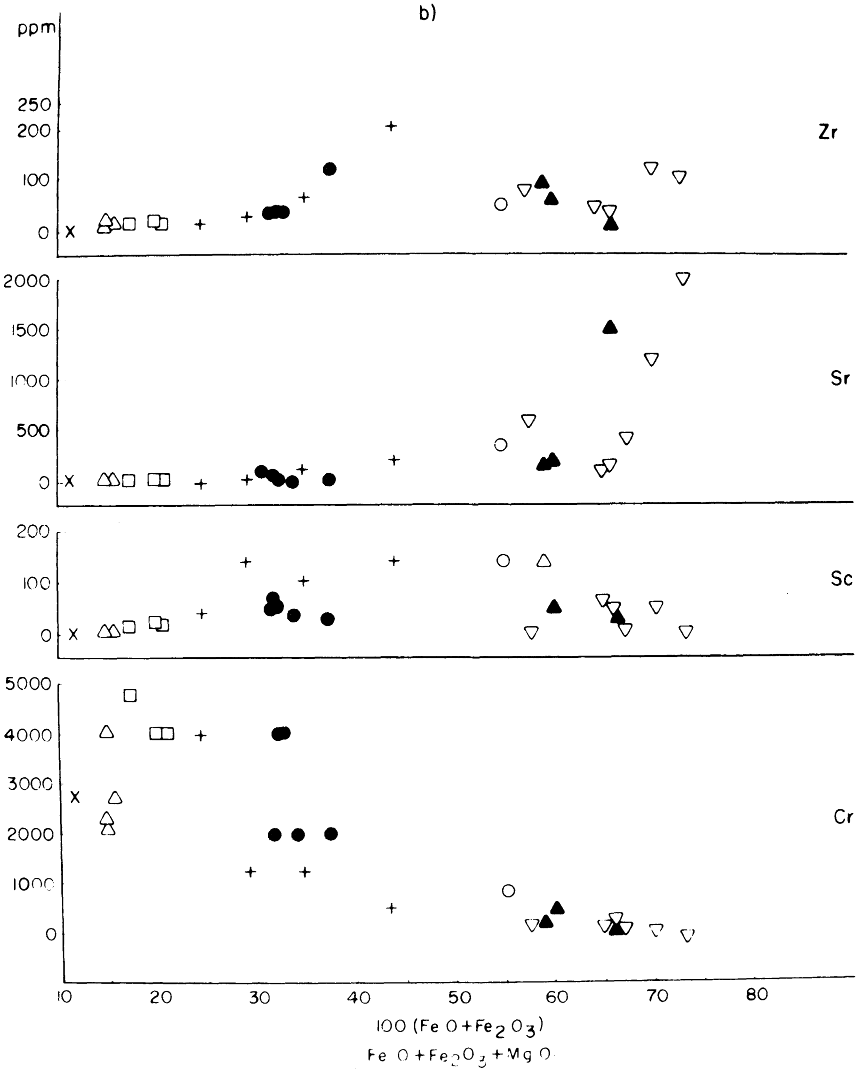


FIG. 2.—Variation diagrams for the Piên mafic and ultramafic rocks. (a) Some major elements, as a function of Simpson's (1954) index. (b) Some trace elements, as a function of Simpson's index. For Piên rock symbols, see fig. 1. Data from Table 1.

Each different petrographic type is chemically rather homogeneous, with the exception of the meta-websterites. In terms of variation diagrams, the following series is established for the ultramafic types: serpentinites—meta-olivine pyroxenites—meta-websterites—meta-norites. Clearly, meta-norites belong chemically to the ultramafic group, while all granulitic lenses within migmatites, together with meta-gabbros and amphibolites, are chemically part of the gabbroic lineage (fig. 2).

Major elements.—The Piên serpentinites are petrochemically similar to other serpentinites (cf. table 1, col. 1, with cols. 16 and 17) and are characterized by low Al_2O_3 (about 1%), high SiO_2 and MgO and very low amounts of alkalies, TiO_2 and CaO. These data are matched by harzburgite analyses from other areas (e.g., California and New Zealand, cf. table 1, cols. 11, 12 and 14) more than by dunitic or lherzolitic sources (e.g., table 1, cols. 10, 13 and 15). The very low CaO content is probably a result of Ca loss (e.g., Miyashiro et al. 1969; Coleman and Keith 1971; Himmelberg and Coleman 1968).



The meta-olivine pyroxenites (table 1) present a relatively high Al_2O_3 content (over 7%), moderate CaO and low TiO_2 and alkali proportions. This indicates a source material with predominance of Al-bearing magnesian silicates (orthopyroxenes, probably also with some Al-spinel) over olivine and especially over clinopyroxenes. The average analysis of these rocks differs markedly from Ringwood's pyrolite and related types (cf. cols. 1, 7, 8 and 9), from compositions of most nodules found in basalts and kimberlites (e.g., Harris et al. 1967; Boettcher et al. 1975) and from analyses of olivine pyroxenites from Alaska-type complexes (Taylor 1967, p. 102).

The present mineralogy of the Piên magnesian schists suggests that orthopyroxene-rich rocks were the most likely source material on account of the Mg, Si and especially Al requirements to be met (see also Schubert 1969). Other possibilities—for example, the replacement of preexisting dolomites or limestones—are ruled out on geological or chemical grounds. Only one analysis of a talc-schist is given (table 1)

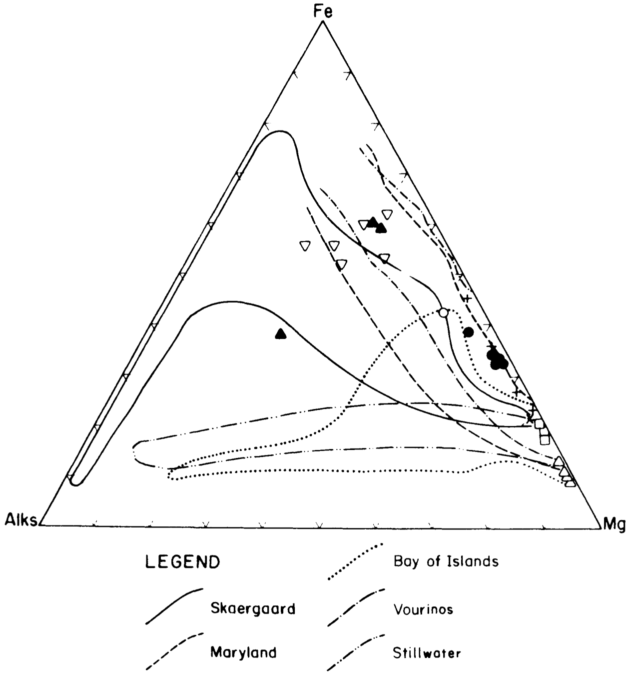


FIG. 3.—Cationic proportions of Fe, Mg and alkalies for the Piên rocks, as compared to trends in several alpine-type (Maryland, Vourinos, Bay of Islands) and stratiform complexes (Stillwater, Skaergaard). Data from Bowes et al., 1970, and Table 1. For Piên rock symbols, see figure 1.

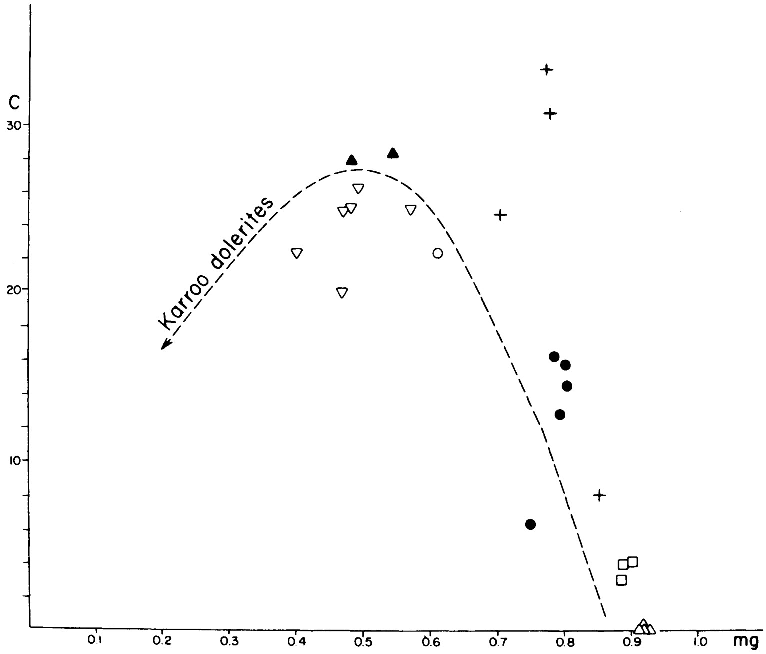


FIG. 4.—Niggli's values *c* vs. *mg* for the Piên rocks. Trend of Karroo dolerites after Evans and Leake (1960). For Piên rocks symbols, see figure 1.

indicating that the rock was selectively enriched at least in SiO_2 during its transformation.

Meta-websterites vary widely in chemical composition (tables 1 and 2) as a reflection of changes in their clinopyroxene/orthopyroxene ratios. Meta-norites, on the contrary, are relatively homogeneous (tables 1 and 2) an exception being sample 96 (sapphirine meta-norite) with its high Al_2O_3 content and undersaturated character.

Rocks from the gabbroic suite, though strongly recrystallized, are undoubtedly of igneous origin (e.g., fig. 4). A trend in Fe-enrichment is significant (fig. 3). On a K_2O vs. Na_2O plot, the gabbroic Piên rocks scatter around a low-potassium line ($K_2O/Na_2O = 0.15$) which separates the area dotted by gabbroic and oversatured rocks from ophiolitic belts (e.g., Dun Montain belt) from the field corresponding to similar rocks associated with island arcs (Coombs et al. 1976). Variability in composition is well reflected in the corresponding norms. Most Piên gabbroic rocks are rich in normative plagioclase (an + ab between 40 and 60%) and either subsaturated or only mildly undersaturated (ol from about 10% to about 0.5%). Their composition approaches that of high-alumina basalts and oceanic tholeiites (Carmichael et al. 1974, p. 376; Ringwood 1975, p. 14) except for higher K_2O and lower CaO contents. Only one Piên metagabbro is oversaturated, and approaches in composition an alkali-poor quartz tholeiite (Ringwood 1975, p. 14). The Piên basic granulites are chemically transitional between mildly undersaturated hy-normative tholeiites and saturated high-alumina basalts (Ringwood 1975).

Trace-elements.—Abundances in minor and trace elements in the Piên ultramafic and mafic rocks reproduce what is know from other occurrences (table 2; see for comparison Goles 1967; Fisher et al. 1969; Wedepohl 1975; and bibliography therein).

Contents of the "compatible" elements Co, Ni, V and Cr in the Piên ultramafic rocks are indistinguishable from those occurring both in alpine-type massifs (e.g., Fisher et al. 1969; Challis 1965) and in layered complexes (e.g., Wager and Brown, 1968, p. 288; Williams and Hallberg, 1973); a comparison of literature data on abundances of some compatible elements (e.g., Ni and Cr) with an incompatible cation (e.g., Ti) shows the same trend for both alpine-type rocks and cumulates (for instance, plots of Ti-normalized Cr- and Ni-contents show a clear linear relationship for all ultramafic and related rocks, irrespective of their geologic setting). This trend is consistent with the petrography of the corresponding rock and, perhaps more than reflecting possible genetic processes, is a result of the known crystallochemical behavior of dispersed minor elements in olivines and pyroxenes.

(*b*) *Mineral composition.*—Tables 3, 4, 5 and 6 (on deposit at ASIS/NAPS) present data on the chemistry of probe-analysed orthopyroxenes, clinopyroxenes, olivines, garnets and spinels. Amphiboles and serpentines were also analysed and will be the subject of an additional paper.

Pyroxenes.—They are very homogeneous and well-exsolved phases, in terms of the corresponding Ca and Mg-Fe contents, and their compositional trend is similar to those shown by pyroxenes from layered and metamorphic complexes (fig. 5). Al_2O_3 content in the Piên pyroxenes from ultramafic rocks is usually high (2.8–4.7% in orthopyroxenes, 3.5–6.7% in clinopyroxenes) and compares well with proportions reported for several high-temperature, high-pressure phases (e.g., Challis 1965; Loney et al. 1971; Medaris 1972; Wilkinson 1975; Capedri et al. 1975). If analysed in terms of their Ca, Mg, Fe, Na and Ti contents, the Piên pyroxenes show a trend of their own, when compared with those shown by pyroxenes from other plutonic and charnockitic environments (Fleet 1974*a*, 1974*b*). In terms of their components jadeite (Jd) and Ca-tschermakite (Ca-Ts), Ca-rich pyroxenes plot within a field subtended by Jd = 0–10 and Ca-Ts = 2–10, well within the area which according to Thompson (1974) defines the composition of most Ca-pyroxenes which do *not* coexist with garnet. The distribution of Al in tetrahedral and octahedral positions is very similar to the ones found in orthopyroxenes from most alpine peridotites (Stroh 1976) and in clinopyroxenes from granulites and inclusions in basalts (Aoki and Kushiro 1968). Na and Ti, although generally low in

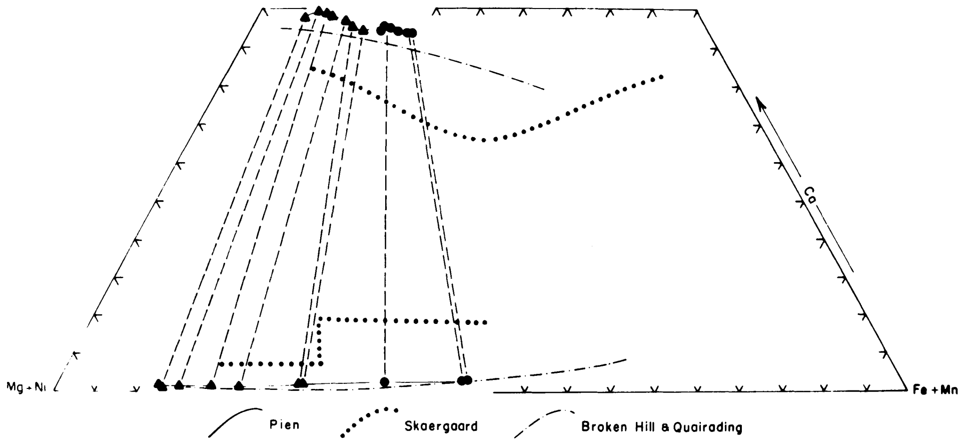


FIG. 5.—Coexisting pyroxenes in Piên rocks represented on a Fe-Mg-Ca cationic diagram. Skaergaard trend from Muir (1955), trend in Broken Hill and Quairanding high-grade metamorphic rocks from Davidson (1968). Triangles: ultramafic rocks; dots: gabbroic lineage.

the Piên pyroxenes, are conspicuously higher than values found in the Red Mountain, New Zealand, pyroxenes in ultramafics (Sinton 1977).

Olivines.—Only three olivines from meta-olivine pyroxenites were analysed (table 5). Crystals are very homogeneous, with a molar Fo content around 90% and thus very similar in composition to olivines from most olivine ultramafic rocks (e.g., Jackson and Thayer 1972; Capedri et al. 1976; Dick, 1977; and bibliography therein).

Spinels.—Four analyses of spinels from meta-olivine pyroxenites, metanorites and serpentinites are presented in table 6. Spinels in magnesium-rich rocks are high in Mg, while the analysed relictic spinel from a serpentinite is a Cr-variety. The cation contents of spinels which coexist with olivine plot close to a narrow Mg-Fe^{2+} equipotential olivine strip at about 800°C (fig. 6; cf. Dick 1977) which suggests an equilibration of spinel with Mg-rich olivines (Fo 85 to Fo 90).

Garnets and sapphirine.—Three analyses of garnets from meta-gabbros and from a sapphirine meta-norite (sample 96) are available (Table 5). All three are almandine- and pyrope-rich with moderate grossular content (12 to 17 mol%) and 2–4% of spessartine. The coexistence of sapphirine and orthopyroxene in meta-norite no. 96 has been discussed separately (Girardi and Ulbrich 1979).

Feldspars.—Dominant feldspars are plagioclases, found as accessory phases in a few meta-websterites, and as main minerals in the rocks of the gabbroic suite. Most are andesines or sodic labradorites, irrespective of bulk rock chemistry, and their Or-content is usually below 0.5% (Girardi 1974). Or-rich alkali feldspars have been detected only in a few acid granulites within migmatites.

DISCUSSION

It is the main purpose of the present investigation to gather information about the crystallization history of the Piên rocks and, in particular, to decide on the ultimate source of the ultramafic and mafic suites. Arguments, as exposed in the literature, are now plentiful to decide among alternative explanations. Obviously, some criteria favoring one or the other genetic hypothesis may be stronger than others. In particular, rather simple arguments derived from field observations and textural criteria are on the whole incontestable and prescribe but one explanation on their sole authority. For instance, the presence of large homogeneous masses of dunites and harzburgites favors a mantle origin, if only because of "the shortcomings of any alternative", as Burch (1968, p. 543) puts it with reference to his discussion on the tectonic emplacement of the Burro Mountain mantle-derived mass in California. In contrast, some chemical or mineralogical criteria—

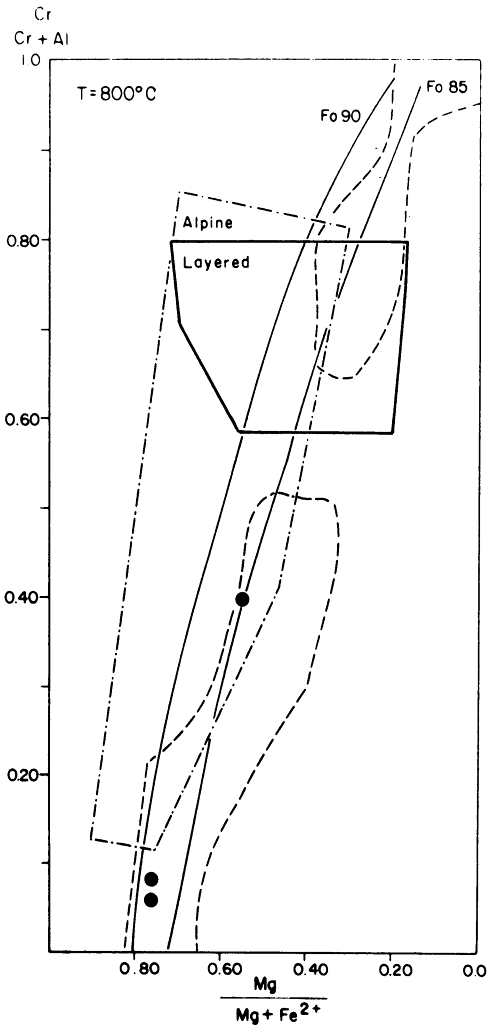


FIG. 6.—A plot of cation proportions in spinel coexisting with olivine. Large dots represent Piên spinels (Table 6). Fields for spinels from alpine-type complexes (dot-dash line), layered intrusions (full line) and regional metamorphic assemblages (dashed line) from Irvine and Findlay (1972) and Evans and Frost (1975). Olivine ideally coexisting with spinel is represented by the equipotential lines for Fo_{90} and Fo_{85}, computed according to a model by Jackson (1969), at a "true" temperature of 800°C (see text).

especially if isolated from field and petrographic information—may only be auxiliary tools, enhancing whatever explanation is preferred, without however supplying decisive petrogenetic arguments, as today's state of the art stands.

Data formerly presented will now be discussed. A picture of the possible origin of the Piên rocks may emerge if we work our way backwards, first establishing the history of the various metamorphic overprints till conditions of the granulite-facies re-equilibration can be established, and arguing from there on as to the source of the Piên massif.

Metamorphic overprints.—As shown in a previous section on Textures, a granulitic (Transamazonic) metamorphism is the oldest detectable event, followed by a second amphibolite-facies Brasiliano metamorphism. The last (Brasiliano-related?) low-temperature event is characterized by the presence of tremolite-actinolite, chlorite, and all three serpentine varieties in appropriate rocks. Temperatures were now probably well below 300°C, as suggested by the chrysotile-lizardite presence in serpentinites (Page 1967; Miyashiro et al. 1969; Evans 1977, p. 408).

Reactions, temperatures and pressures of crystallization during granulitic metamorphism.—Most Piên meta-olivine pyroxenites, meta-websterites and meta-gabbros are a direct product of granulite-facies reactions and thus provide clues as to the original mineral composition. No plagioclase was found in meta-olivine pyroxenites and so relevant reactions in olivine rocks probably only re-equilibrated spinel with aluminous pyroxenes:

Forsterite + Mg-tschermakite
(Mg-Ts, opx) = Enstatite + Spinel (1)

Forsterite + Ca-tschermakite
(Ca-Ts, cpx) = Diopside + Spinel (2)

(Evans 1977). In pyroxene rocks, and especially in meta-websterites, the main reactions re-equilibrated Al-pyroxenes:

Enstatite + Ca-Ts
(cpx) = Diopside + Mg-Ts (opx) (3)

and broke down subcalcic clinopyroxene—if such a phase existed in the original rock—into enstatite and diopside (eventually with some spinel):

$Ca_2Si_2O_6.3Mg_2Si_2O_6$ (subcalcic cpx)
= 2 Diopside + 2 Enstatite (4)

(cf. Thompson 1974; Wilkinson 1975). One meta-websterite presents coexisting spinel and plagioclase, so that a reaction such as:

$$\text{Enstatite} + \text{Anorthite} + \text{Spinel} = \text{Diopside} + \text{Mg-Ts (opx)} \quad (5)$$

may also be relevant for that particular case. Garnet is absence in almost all ultramafic rocks (the most notable exception is a single boulder of a garnet-rich "websterite"—sample GP 260—which will be discussed later on), and is not found as relicts. On the other hand, Ca-rich pyroxenes are relatively low in jadeite and tschermakite components (cf. Thompson 1974, fig. 11). These data suggest that garnet was neither a pre-existing phase before the onset of the oldest (granulitic) re-equilibration, nor a significant product of it. So, all possible reactions are here taken to be essentially garnet-free (see also geobarometric estimates, below).

Meta-norites are plagioclase-bearing rocks, in which both pyroxenes, plagioclase, spinel, and secondary amphiboles are the main minerals. Chemically, they present 40–57% normative hy and about 10% ol in the norm. The most important re-equilibration reactions in these rocks are probably only (3) (4) and (5) since all other possible mineral equilibria involve either the coexistence of anorthite and olivine or the breakdown of grossular-bearing pyrope. Reactions involving sapphirine, pyroxene and spinel in sample 96 (a sapphirine "meta-norite") have been discussed elsewhere (Girardi and Ulbrich 1979).

Data on barometry and thermometry, derived from mineral chemistries in rocks which preserved mosaic granoblastic textures, are presented in table 7. The different 2-pyroxene thermometers yield the most consistent values, especially the one proposed by Wells' (1977) reworking of the Wood and Banno formula. The thermometer by Wood and Banno (1973) apparently exaggerates temperatures for Mg-rich pyroxenes (see discussion in Wells 1977), while figures based on Banno's (1975) proposal are on the high side for Fe-rich members. Spinel-olivine coexistence permits a reasonable temperature estimate if the calibration by Evans and Frost (1975) is used. Pressures were also estimated (table 7).

Evidence is clearly in favor of a crystallization pressure of at least five to seven kb, while temperatures based on the two pyroxene thermometer change from 880°C for some meta-olivine pyroxenites to about 750°C for some granulites and meta-websterites. There is no clear evidence of a regional temperature gradient (see fig. 1), so that these differences in temperatures, more than being controlled by actual processes, may well reflect inconsistencies in present-day thermometric calculations and errors in mineral analyses. The obtained temperatures and pressures are consistent with crystallization under conditions of the intermediate-pressure granulite field (Ringwood 1975, p. 20), characterized by the stable association aluminous diopside-aluminous orthopyroxene-spinel-olivine (and incompatibility of olivine and anorthitic plagioclase) in ultramafic rocks. The corresponding minerals are neither markedly zoned nor do they show signs of unmixing, so that the granulitic re-equilibration was very thorough (Sinton 1977).

A special case is presented by the garnet pyroxenite MP-260. Its garnet presents 17% grossular, 41% pyrope, and 41% almandine (compare with the garnets from the high-pressure ultramafic Oahu and Moncaup occurrences which show about 11–13% grossular, 53–63% pyrope, and 25–33% almandine; cf. Beeson and Jackson 1970; Kornprobst and Conquéré 1972). The orthopyroxene from the Piên sample has 4.86% Al_2O_3, 25.1% MgO and 17.1 FeO, and an appr. Mg-Ts content of 6.8%; clinopyroxenes present 7.15% Al_2O_3, 12.5 MgO, and 7.04% FeO (Ca-Ts is 9.7%). The spinel is a Cr-poor hercynite. Plagioclase is zoned from calcic to sodic labradorite (center to border); it is, as spinel, interstitial and xenomorphic. This rock is unusual in several respects. It presents a five-phase mineral assemblage which is presumably out of equilibrium, and is the only rock in which a visible zoning was detected. Furthermore, it does not fit the general mineralogical pattern presented by all the other ultramafic Piên rocks. While all mineralogical indications point to a recrystallization under a "medium-pressure" regime, the garnet pyroxenite presents a higher-pressure mineralogy (which, according to Herzberg's geobar-

TABLE 7.

Estimated Temperatures and Pressures of Crystallization, Piên Rocks

	MOP 78	MOP 116	MW 223	MW 122	MW 42	MN 74	MN 80	Gr 46	Gr 105	Gr 14	GP 260
Association	Opx-Cpx Ol-Sp	Opx-Ol-Sp	Opx-Cpx Pl	Opx-Cpx	Opx-Cpx	Opx-Cpx Pl	Opx-Cpx Pl-Sp	Opx-Cpx Pl	Opx-Cpx Pl	Opx-Cpx Pl	Ga-Opx-Cpx-Pl-Sp
X(Fe, Opx)	0.12	0.12	0.28	0.14	0.28	0.16	0.21	0.48	0.37	0.48	0.28
X(Fe, Cpx)	0.06	—	0.13	0.07	0.12	0.08	0.10	0.19	0.15	0.20	0.12
% Mg-Ts, Opx	6.2	8.0	6.5	8.0	5.9	10.0	10.0	3.5	5.0	3.0	6.8
% Ca-Ts-Cpx	7.0	—	2.8	6.9	5.4	10.5	7.8	2.0	2.8	2.9	9.7
2-pyroxene (1)	978		823	835	848	921	730	732	767	753	744
2-pyroxene (2)	883		848	723	890	860	651	867	840	910	725
2-pyroxene (3)	884		789	731	824	847	645	733	750	762	781
OBATA (1976)	6^b	5^b	~6	<7	~5						
HERZBERG (1978)	$>7^b$						$>6^c$				10^d

Note: Temperatures in °C, pressures in kilobars. Underlined temperatures are represented in Fig. 1.

MOP: meta-olivine pyroxenite; MW: meta-websterites, MN: meta-norite; Gr: gabbroic granulites; GP: garnet pyroxenite.

[a] maximum values

[b] figures refer to minimum P

[c] appr. minimum P for spinel gabbros

[d] appr. minimum P for garnet pyroxenite

Source: (1) Wood and Banno, 1973; (2) Banno, 1974; (3) Wells, 1977.

ometers, would correspond to a minimum pressure of about 10 kbs.—see table 7). We can only guess at the significance of the presence of this rock within a supposedly overall lower-pressure mineral association (controls of rock composition on garnet crystallization?; tectonic control of emplacement of sample MP 260?)

The Piên ultramafic rocks: residual mantle material or magmatically differentiated mass?—The literature contrasts on geologic grounds alpine-type peridotites against layered major intrusions and concentrically zoned Alaska-type bodies (Wager and Brown 1968; Loney et al. 1971; James 1971; Medaris 1972; Ringwood 1975; Coombs et al. 1976; Dick 1977; Sinton 1977; and bibliography therein). Magmatic differentiation is the controlling phenomenon in the last two geological occurrences, to be called here "differentiated complexes" for short, while most of the alpine-type peridotites—whether ophiolitic or not, whether of the "cold-slab" allofacial (Evans, 1977) variety or intruded as hot masses—have by now been interpreted as transported (residual) mantle (e.g., Medaris 1972; Loney et al. 1971; Ringwood 1975; Coombs et al. 1976; Dick 1977; and many others). Leucocratic rocks associated to alpine peridotites in ophiolitic suites, predominantly basaltic or gabbroic although often presenting a wide range in composition, have been interpreted as genetically related and even complementary fractions of the same melting process (e.g., Coombs et al. 1976; Sinton 1977). There are, however, both chemical and structural reasons to believe that some alpine peridotites may be tectonically transported basal portions of deeper-seated layered complexes (Bowes et al. 1970; see also Capedri et al. 1976, p. 176).

Below are cited contrasts thought to represent differences between mantle material, on the one hand, and differentiated complexes, on the other. Reasons for considering exposed mantle materials as refractory residues of partial melting have been indicated by many authors (e.g., Ringwood 1975, p. 176ff.).

1. Residual mantle is apparently predominantly dunitic, harzburgitic and/or lherzolitic (cf. bibliography); orthopyroxenites, websterites and olivine pyroxenites are apparently absent, or present only in minor amounts. Similarly, alpine peridotite bodies are petrographically rather homogeneous.

2. Mafic rocks associated with residual ultramafic bodies usually appear in proportions that do not exceed a 1:1 ratio or may even be absent (e.g., Ringwood 1975, p. 89, 91; Dick 1977; but see Jackson and Thayer, 1972, p. 292).

3. Bulk chemistry of residual mantle reflects the petrogenetic process. Most "residual" peridotites are chemically strongly undersaturated, alkali- and Ti-depleted rocks, with low Al_2O_3 and CaO contents (usually, well below 1%, CaO sometimes only as traces). But caution should be exercised when applying chemical criteria, because as a single argument, element depletion or enrichment may not be conclusive evidence since it is the crystallochemical behavior of Mg-olivines, for instance, that controls to a great extent the contents of minor and trace elements in peridotitic residues as well as in olivine cumulates.

4. Mineral chemistry in mantle material reflects its chemical and petrographic homogeneity, provided the original material was not chemically transformed (for example, serpentinized and subsequently recrystallized at higher temperatures). Thus, olivines and pyroxenes in mantle peridotites are usually very homogeneous and Mg-rich (olivines about Fo 90, orthopyroxenes about En 90–95, e.g., Medaris 1972; Sinton 1977; Dick 1977; and many others). In some cases, spinel compositions (equilibrated with olivines) may provide a significant argument for distinguishing mantle peridotites from rocks derived from differentiated complexes (e.g., Medaris 1975); moreover, the chemical trend of spinels in mantle material is one that seems to favor an explanation in terms of an isothermal fusion and recristallization process, in contrast to spinel chemistry in stratiform complexes (obviously heterothermal, cf. Dick 1977).

On the strength of previously detailed data in the descriptive section, there is little doubt that the Piên ultramafic suite was originally an association of predominant olivine harzburgites, pyroxenites (mainly websterites and orthopyroxenites) and subordinate olivine pyroxenites and very melanocratic norites, if the present outcrop pattern is taken as indicative of relative proportions (fig. 1). This is a rock association found mainly in differentiated layered complexes. As an additional argument, the associated mafic suite at Piên is exposed in amounts that surpass by a factor of approximately three the outcrops of the ultramafic rocks.

Analysed olivines are all from meta-olivine pyroxenites (table 5) and reflect in their composition the control of rock chemistry and submagmatic crystallization temperatures (Evans 1977, p. 425). Spinels, again, show compositions compatible with those found in high-grade granulites (see fig. 6 and cited bibliography). The spectrum of pyroxene composition, however, is clearly different from what is expected from mantle-rock pyroxenes or their re-equilibrated products (see especially Dick 1977). There is a large spread in Mg/Fe ratios, in particular in the pyroxenes from ultramafic rocks, thus providing a trend that is thought to be characteristic of pyroxene chemistry in differentiated complexes or their recrystallized assemblages (see fig. 5 and discussion above).

The relationship of the Piên mafic rocks to the ultramafic unit.—There are certain indications that both mafic and ultramafic at Piên are genetically related phenomena of a more or less complex magmatic process or processes. One indication is of course the close geological association between the two units. As a second argument, chemical parameters show a continuous variation (see figs. 2, 3 and 4), suggesting that the ultramafic and mafic suites are magmatically complementary parts. There is little else to go by, but on those grounds alone magmatic consanguinity may be a better explanation than any other genetic model. If so, then the ultramafic rocks may be considered the crystal-fractionated portion (or portions) of one or several major units and the gabbroic suite its crystallized magmatic residue.

A simplified representation first proposed by O'Hara (1968) is especially suited to depict possible relationships among mafic and ultramafic rocks (see also Clarke 1970; Cawthorn and Strong 1974). It consists essentially in projecting corrected rock analyses—now expressed in terms of A("Al_2O_3")-S(SiO_2)-M("MgO")-C(CaO)—onto planes that are mineralogically significant to the genesis of mafic and ultramafic rocks; projection points within the ACMS tetrahedron are the ones representing enstatite, diopside and forsterite. Figure 7 is a projection from diopside (CMS_2) onto the plane C_3A-M-S, which contains the points representing enstatite and olivine. Arguments put forward by O'Hara (1968) and Cawthorn and Strong (1974) will now be followed. Any rock composition such as A, if subjected to partial melting, will yield at 30 kb., say, a pseudo-eutetic fraction represented by the corresponding compositional point (fig. 7). Disappearance of certain solids will move the liquid composition down along a cotectic line and subsequently along the BA segment towards A. Conversely, if A represents a liquid coexisting with olivine (and possibly diopside, as the phase representing the projection point), then separation of olivine by crystal fractionation will force the liquid to move towards B, leaving behind a forsterite-rich solid residue.

The clustering of the Piên gabbroic rocks in figure 7 occurs in an area of the diagram in which experimental studies place the compositions of liquid fractions in broadly "basaltic" systems. On the other hand the Piên ultramafic rocks lie in areas compatible with models of crystal fractionation. It is observed, on a purely geometric basis, that the Piên meta-norites (except the sapphirine-bearing sample 96, see discussion above) and meta-websterites (respectively crosses and solid circles in fig. 7) can on the whole be depicted as concentrates governed by simple fractionation of orthopyroxene and diopside. Both rock types project as a well-defined cluster and are placed, at the same time, on a line joining enstatite with possible liquids in low temperature areas at 20 kb or lower pressures (fig. 7). Similarly, the clustering of points showing the Piên meta-harzburgites (serpentinites) and meta-olivine pyroxenites (upright triangles and squares, fig. 7) can be explained graphically as representing accumulations of olivine and orthopyroxene (and diopside, as projection point) in proportions that change according to the necessary petrographic requirements. Thus, the sim-

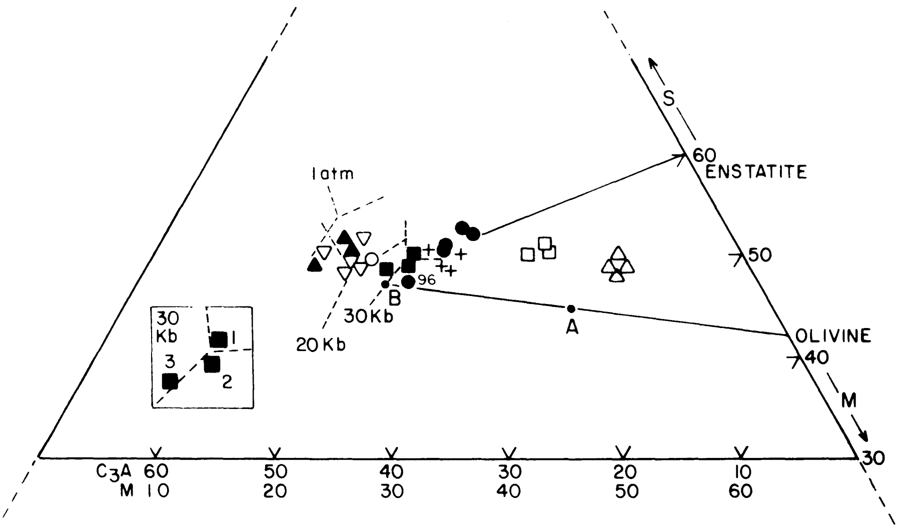


FIG. 7.—A projection of chemical data for the Piên rocks into O'Hara's (1968) ASMC tetrahedron, from diopside (CMS_2) onto the S-M-C_3A plane. High-pressure pseudo-invariant points for natural systems from O'Hara (1968). For symbols of the Piên rocks, see figure 1; the chemically unusual sapphirine meta-norite is identified by "96". Possible primary magmas are represented as solid-squares: (1) is an olivine tholeiite, (2) an alkali-poor olivine tholeiite and (3) an olivine basalt (cf. inset; chemical data from Green and Ringwood, 1967). Possible undepleted and depleted mantle rocks fall mostly within the triangle formed by the 20-kb invariant point and the enstatite and olivine points. For details, see text.

plest model to be adopted would be to relate the mafic and ultramafic suite at Piên through a single magmatic event.

Geometrically, an educated guess can now be made about a possible primary magma that on crystallization would fulfill the necessary requirements: it would fall somewhere between the extended area representing the Piên ultramafic rocks and the cluster portraying gabbroic specimens. Thus, a projection of assorted basaltic compositions (taken from analyses in Ringwood 1975, and Wilkinson 1975) into the ACMS tetrahedron pinpoints as suitable candidates an olivine tholeiite (1 in fig. 7; table 1, 18) an alkali-poor olivine tholeiite (2 in fig. 7) and an olivine basalt (3 in fig. 7), among others.

Can this model of fractionation be substantiated on grounds other than geometrical? Experimental data are available on the crystallization behavior of some liquids of basaltic compositions at high pressures (see summary and references in Ringwood 1975, and Wilkinson, 1975). The first question to be answered asks for a crystal fractionation pattern that may, after subsolidus granulite re-equilibration, yield significant amounts of olivine- and orthopyroxene-rich rocks (see discussion in the corresponding descriptive section).

Mineralogy prior to granulite recrystallization was of course somewhat different from what it is today, if only because subcalcic clinopyroxene is an important liquidus phase under high-pressure crystallization of basaltic magmas (Thompson 1974, see also Green and Ringwood 1967). Subsolidus re-equilibration of these subcalcic clinopyroxenes will trigger breakdown into Al-orthopyroxenes and Al-clinopyroxenes (eventually with some spinel; see reaction 4) and thus increase the modal proportions of subsolidus orthopyroxene. On the other hand, experimental data (Green and Ringwood 1967) show that orthopyroxene and olivine are primary liquidus phases in those basaltic liquids that qualify as a primary magma at Piên. The olivine tholeiite represented in fig. 7 (Green and Ringwood, 1967; Table 1, 18) presents olivine as the first liquidus phase at nine kb (1290°C), shortly followed by Al-orthopyroxene (1250°C), and then by subcalcic

Al-clinopyroxene at 1230°C. At 13.5 kb, Al-orthopyroxene is the primary liquidus phase (Green and Ringwood 1967, p. 139, 150), so that at some intermediate pressures olivine and orthopyroxene will be the main precipitated phases at liquidus or near liquidus temperatures.

Thus, considerations based on experimental data show that the modal proportions of orthopyroxene and olivine in early crystal fractions may be quite high, and the proportion of orthopyroxene in particular will tend to increase as coprecipitated subcalcic clinopyroxene breaks down under subsolidus (granulitic) conditions.

Experimental studies on olivine tholeiites put some pressure constraints on crystallization conditions. Olivine disappears as a liquidus phase above 12 kb in "normal" olivine tholeiites (Green and Ringwood 1967) and above eight kb in the alkali-rich aluminous Snake River olivine tholeiite (Thompson 1974). Orthopyroxene is an important liquidus phase in "normal" olivine tholeiites at pressures between nine and 18 kb, while clinopyroxene is ubiquitous at nine kb and becomes especially important above 10 kb. These results, if applied to the Piên case, suggest that fractionation of olivine and pyroxenes may have taken place at pressures over seven kb (a "minimum" pressure of the Piên granulitic- re-equilibration, see above) and up to about 12 kb, under conditions that could have been that of a large "magma chamber" somewhere at depths of 20–25 Km or thereabouts.

The differentiation trend of Green and Ringwood's olivine tholeiite (table 1, 18) deserves closer examination. At nine kb and after an extraction of 30% solids (12% olivine, 14% subcalcic clinopyroxene, 4% orthopyroxene), the residual liquid is significantly enriched in Al, Ca and Na and visibly depleted in Mg (table 1, 19). The fractionation trend is thus towards an alumina-rich basalt, with over 50% normative plagioclase and half the original ol content. SiO_2 and K_2O contents of Green and Ringwood's residual liquid, in particular, are still too low to match the compositions of the Piên meta-gabbros (and of most published analyses of abyssal tholeiites and high-alumina basalts); this is certainly more a quality inherited from the original composition of the olivine tholeiite (table 1, col. 18) than a reflection of any shortcomings of the differentiation tendencies, since other chemical parameters are already very similar (compare, for instance, Al, Na and Mg contents of columns 5 and 19 in table 1). More important than low SiO_2 and K_2O proportions in the differentiated liquid is the tendency to avoid ne-normative undersaturation. Green and Ringwood (1967, p. 146ff.) argued on the point that enhanced olivine and subsequent orthopyroxene fractionation at pressures somewhat below nine kbs will drive the residual liquid towards compositions closer to published analyses of ol- and hy- normative high-alumina olivine basalts. It is also very likely, on results shown by these experiments, that olivine tholeiites with high normative hy content will eventually fractionate away from ne-normative undersaturated fields and towards anorthite enrichment (Green and Ringwood 1967, p. 148).

It is difficult to envisage, on hydrodynamic grounds alone, large magma chambers controlling crystal differentiation at great depths, a process which requires a delicate density balance. But several authors have presented strong arguments to show that magmatic processes involving crystal fractionation are or were operative at pressures of at least 10 kb, thus located at the base of the crust (or in shallow mantle regions?; e.g., Jackson 1968; Wilkinson 1975). Little can be said about such processes except that they may involve large volumes of crystallizing magma and may not necessarily be similar to the ones generating the shallow-level layered complexes. In the case of Piên the actual geometry of the outcrops indicates a complicated tectonic history (fig. 1) that precludes any attempt at a sound reconstruction of the intrusive and post-intrusive history. Geophysical research, in particular, may be helpful in bringing forward the additional evidence needed to fill structural and stratigraphic gaps.

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