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Metamorphism, Metasomatism And Mineralization At Lagoa Real, Bahia, Brazil

Lydia Maria Lobato

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METAMORPHISM, METASOMATISM AND MINERALIZATION

AT LAGOA REAL, BAHIA, BRAZIL

by

Lydia Maria Lobato

Department of Geology

Submitted in partial fulfillment
of the requirements for the degree of
Doctor of Philosophy

Faculty of Graduate Studies
The University of Western Ontario
London, Ontario
September, 1985

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ABSTRACT

Uranium deposits cumulatively in the 100,000 tonne U_3O_8 range occur within ductile shear zones transecting Archean basement gneisses of the São Francisco Craton, at the Lagoa Real region of south-central Bahia, Brazil. The gneisses, dated at 2.6-3.0 Ga, are at amphibolite and granulite facies and overlie to the west, the Proterozoic Espinhaço metasedimentary sequence along a thrust fault.

Petrography and mineral chemistry show that in the zones of alteration/mineralization, the original k-feldspar + quartz + albite/oligoclase + hastingsite assemblage, is replaced by albite + aegirine-augite + andradite + hematite assemblages, with or without uraninite. This information along with oxygen isotope, whole rock geochemistry and fluid inclusion studies indicate that the alteration process involves removal of Si, K, Rb, Ba and addition of Na under oxidizing conditions. V, Pb, Sr were introduced along with U via interaction with saline CO_2 -rich, isotopically light fluids under varying water/rock ratios and at temperatures of 500-550°C. $^{87}Sr/^{86}Sr$ systematics suggest that it is unlikely that Sr, and by extension uranium, were introduced by fluids originating from the basement.

gneisses'.

Geological constraints and the general alteration pattern are consistent with the release of the mineralizing fluids in response to overloading of the basement rocks onto the sedimentary Espinhaço via a thrust mechanism.

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CHAPTER 1

INTRODUCTION

The study of mechanisms leading to natural uranium accumulations has been dominated by consideration of moderate to low temperature processes. On the other hand, relatively little attention has been paid to the formation of uranium deposits at high temperatures (above $\sim 300^{\circ}\text{C}$), except for pegmatite hosted deposits. Perhaps the best account of the movement of uranium during higher grade metamorphic processes is obtained by the study of hydrothermal uranium deposits. Rich et al (1977) have shown that:

- hydrothermal uranium deposits are usually related to structural deformation;

- even though hydrothermal uranium deposits occur in diverse geological environments and in a wide variety of rock types and ages, the host rocks in most hydrothermal uranium deposits are competent felsic igneous and

metamorphic rocks;

- high uranium concentrations usually correlate well with high concentrations of potassium and silica;

- the uranium deposits have formed at temperatures below 350°C.

It is widely accepted that high grade rocks form near the base of the ancient crust and it is common wisdom that such rocks are depleted in uranium (Heier, 1973, Atal et al, 1978, Fyfe, 1979). Economic uranium deposits are found in low grade metamorphosed sediments as well as in hydrothermal deposits; this is evidence of the tendency of uranium to concentrate upwards in the crust (Heier, 1979). The observations listed above are consistent with this evidence.

In spite of all the research concerning the geochemistry of uranium, the knowledge about its behaviour in metamorphic and/or metasomatic environments is remarkably restricted. Nevertheless, as pointed out by Rogers and Adams (1978) "the mobility of the uranyl ion in surface waters plus the tendency for uranium to occur together with potassium, lead to the presumption that uranium is highly involved in metasomatic activity".

Whereas the association of uranium with K-enrichment has been investigated to some extent, the association of uranium with Na-metasomatism is a relatively new concept in the western geological literature (Adamek and Wilson, 1977, Yerle et Thiry, 1979, White and Martin, 1980, Kish and Cuney, 1981). Such association has been widely described by Russian geologists (e. g. Nikolskii, 1973, Smirnov, 1977, Grechishnicov, 1980, Belevtsev, 1980, Kalyaev, 1980, Tugarinov, 1980, Zhukova, 1980).

Although a small number of high temperature ($\sim 500^{\circ}\text{C}$) uranium deposits have been reported (see Kimberley, 1978), there is very little available information concerning the conditions constraining their formation. In addition, the work by Lemoine (1975, in Rich et al, 1977) which demonstrated that with increasing temperature the solubility of UO_2 in equilibrium with UO_3 probably passes through a maximum near 260°C , with a rapid decrease in the solubility of UO_2 below 200°C and probably also above 300°C , helped support the assumption that most economic uranium deposits are sedimentary or related to relatively low temperature hydrothermal processes.

In the Lagoa Real region of central-southern Bahia State, Brazil, high-grade basement gneisses of the São Francisco Craton show localized zones of uranium enrichment that in places attain ore grade. The deposits

are found within large-scale ductile shear zones associated with extensive high temperature ($\sim 500^{\circ}\text{C}$) metasomatism and retrograde alteration of the basement rocks, producing long tabular bodies of albitites. Lagoa Real is one of the largest uranium deposits in Brazil, with several orebodies comprising a combined ore reserve of the order of 100,000 metric tons (tonnes) of U_3O_8 in ore of average grade 0.15% U_3O_8 , i.e. 1.5 kilos U_3O_8 per metric tonne of ore (Brito et al, 1984). The mineral associations and the uranium mineralization distribution and conditions of formation, makes the Lagoa Real an unique example of highly economical uranium concentration in high grade metamorphic rocks associated with Na-metasomatism.

This study is an attempt to describe the petrography, mineralogy, and geochemistry of the metasomatically altered rocks and their host assemblages, and it is intended to contribute to the understanding of the history of the metasomatic and retrograde metamorphic processes, their relation to the uranium mineralization and their possible link to the tectonic and structural framework of the basement terrain in Bahia State during late Proterozoic times.

CHAPTER 2

REGIONAL GEOLOGY

2.1. Introduction

The area under investigation is located in the central-southern part of the São Francisco Craton, which Almeida (1977) defined as a cratonic nucleus stabilized at the end of the Transamazonian geotectonic cycle (2.7 - 1.8 Ga), surrounded by Late Proterozoic (Brazilian cycle, 700 $\pm$ 0.2 Ma) belts. The craton occupies almost the entire state of Bahia as well as parts of Minas Gerais, Goiás, Pernambuco and Sergipe states. It is limited by the Sergipe fold belt to the northeast, the Riacho do Pontal fold belt to the north, the Rio Preto belt to the northwest and the Brasília and Araçuaí belts to the south and southeast (Mascarenhas, 1983).

A cratonic nucleus has long been recognized in the area by many workers, but even today much controversy exists concerning its boundaries. This nucleus was identified by Guimarães (1951) as "Arqui-Brasil". Pflug

(1965) called "Maciço do São Francisco" the cratonic area east of the Proterozoic Espinhaço range. Bosa (1966) used the term "Craton Franciscano" to designate the area located entirely in the São Francisco river hydrographic basin that was consolidated in the Archean. Almeida (1967) denominated the area as "Plataforma do São Francisco" and "Craton do São Francisco", including both the São Francisco river basin region and the Espinhaço range. Also in 1969, Pflug and others restricted the name "Craton Sanfranciscano" to the region west of the Espinhaço range. Cordani (1973) established a geotectonic model which contemplated the existence of two cratons, separated by a belt of Brazilian isotopic rejuvenation in the Paramirim river valley: the São Francisco Craton to the west of the Paramirim valley and the Salvador Craton to the east. Nevertheless, Almeida's (1977) conception is to this date widely utilized.

In the state of Bahia, the São Francisco Craton comprises high grade metamorphic basement rocks covered by a succession of Precambrian, Paleozoic and Mesozoic metasedimentary and metavolcanic sequences, and Cenozoic deposits. The high grade basement rocks are divided into various complexes: to the east of the Espinhaço range are the migmatitic Caraíba, the granulitic Jequié, and the gneissic-migmatitic Paramirim. To the west, are the Santa

Isabel and Guanambi migmatitic complexes (Jardim de Sá et al, 1976 b).

The Lagoa Real uranium deposit is hosted by basement rocks of the Paramirim complex. The Chapada Diamantina lies to the east and the Espinhaço range (Santo Onofre belt) lies to the west of this basement complex, constituting Middle Proterozoic covers on the São Francisco Craton (Fig. 2.1); these are both part of the Espinhaço Supergroup, which extends to the south into the Minas Gerais state (Jardim de Sá, 1978, 1981). Supracrustal sequences of Archean and Lower Proterozoic ages ("metamorphic complexes") border the Santo Onofre belt to the west (Caetité-Jacaraci-Licínio de Almeida complex) and the Chapada Diamantina to the far east of the area (Brumado complex). These are metavolcanic and metasedimentary sequences, and may represent greenstone belt type sequences according to Mascarenhas (1979, 1981).

Various regional mapping and geological compilations have introduced a complex nomenclature in an attempt to uniformly name the basement rocks and later covers in the region. Among these, the LETOS Project (Moutinho da Costa et al, 1976) initiated in 1972, mapped in 1:250 000 a large portion of the state of Bahia. This work, widely used as the geological base for the region, grouped both basement rocks and metavolcanic/metasedimentary complexes

Fig. 2.1. Simplified geological map of the Lagoa Real region and surroundings, displaying the Archean gneisses, the lower Proterozoic greenstone belts, the middle Proterozoic Espinhaço Supergroup metasedimentary sequence, thrusts and uranium deposits (modified after Schobbenhaus et al, 1976).

under the designation "Pré-Espinhaço", the former being generically named metatexites and the latter, "metamorphic complexes". The Espinhaço Supergroup was divided into Lower, Middle and Upper Espinhaço. "Non-differentiated" or "Undivided Precambrian" is used by Schobbenhaus et al (1976) to designate the gneissic-migmatitic basement terrains. The "metamorphic complexes" of Moutinho da Costa et al (1976) were grouped under "Precambrian-C", and "Precambrian-B Unit" is the Espinhaço Supergroup subdivided in units 1, 2 and 3.

Inda and Barbosa (1978 a, b) and Neves et al (1980) introduced a geochronological approach to the geology of Bahia state, thus creating geotectonic and geochronological domains. A 1:100 000 scale mapping (Moraes et al, 1980) of this region and others, named the basement terrains "Migmatitic-Plutonic Association"; "Metamorphic Association" was used to name the metavolcanic and metasedimentary sequences; the "Espinhaço Association" was divided in accordance to Moutinho da Costa et al, (1976); see Table 2.1).

This work reviews the geology of the region surrounding the Lagoa Real deposit. In view of the much debated nature of the geological knowledge of the region, this review mostly utilizes the works by Moutinho da Costa et al (1976), Schobbenhaus et al (1976) and Moraes et al

(1980), unless otherwise stated. As it is not within the scope of the present work to discuss the various interpretations in the literature, the works by the above authors were chosen because not only are they the latest published, but also because they contain more comprehensive geological descriptions.

2.2. The basement rocks

The Paramirim complex is located between the Chapada Diamantina and the Espinhaço range, extending to the south of the Chapada along the Paramirim and de Contas rivers valleys. Orthogneisses and migmatites of granodioritic to tonalitic, and of granitic compositions predominate, together with diapiric rocks of similar compositions (Jardim de Sá et al, 1976 b, Jardim de Sá, 1978).

Domains of gneisses, heterogeneous and homogeneous migmatites, and granites, have been characterized. Although recognized as separate domains, the metatexites and diatexites (according to Mehnert, 1968) do not have their limits well defined in the region. The metatexites, or heterogeneous migmatites predominate. They are foliated with cataclastic textures, and highly deformed. Migmatites of kinzigitic (granulite facies rocks of pelitic origin) composition occur to the south; these are

banded rocks essentially composed of quartz, orthoclase, biotite, sillimanite, garnet and cordierite intercalated with granitic and pegmatoid bands (Pedreira et al, 1978). The diatexites (homogeneous migmatites, Mehnert, 1968) are homogeneous rocks, with compositions which mainly range from granitic to syenitic. Their neossomes (i.e. newly formed rock portion, Mehnert, 1968) are pegmatoid; their paleossomes (unaltered or slightly modified rock, Mehnert, 1968) is made up of biotite and amphibole, and the predominant texture is porphyroclastic. As a rule, these rocks have been named: biotite-plagioclase-gneisses, hornblende-microcline-gneisses, hornblende-biotite-gneisses. The plagioclase varies from low oligoclase in the more acid varieties to andesine in the tonalites. The metamorphic grade is high amphibolite with frequent zones of retrogression to high greenschist facies; nevertheless, hypersthene-plagioclase-gneisses of enderbitic composition (Streckeisen, 1974) are found sporadically.

Intrusive rocks are also found in the gneissic-migmatitic terrain. These are granitic bodies of variable sizes and ages (see 2.4.) and mafic and ultramafic rocks. The latter forms lenticular bodies of talc schists or isolated bodies of serpentinites, hornblendites, pyroxenites and gabbros.

N-S elongated bodies of metamorphosed volcanoclastic, volcanochemical and sedimentary rocks occur in the Paramirim; these have been generically designated "metamorphic complexes". One of these, the Caetité-Jacarací-Licínio de Almeida complex, forms a long band that extends from north of Caetité to south of Jacarací bordering the eastern margin of the Espinhaço range, west of the Lagoa Real region. In this complex, muscovite-schists, chlorite-schists and magnesium rich rocks abound. The latter includes phlogopite-schists, talc-chlorite-schists and actinolite-dolomite-schists. Quartzites, goudites, marbles and amphibolites are also common. The metamorphic grade is high amphibolite.

Although many metamorphic complexes have been recognized throughout the state and many have been identified as greenstone belts by Mascarenhas (1979, 1981), their real geological significance as well as their chronostratigraphic position is to this date very controversial.

2.3. The Espinhaço Supergroup

The Espinhaço Supergroup outcrops in the central and central-southern parts of the state of Bahia notably in the Chapada Diamantina basin, and in the Espinhaço range

(Santo Onofre belt). The metasediments in the Chapada-Santo Onofre reflect a diverse spectrum of platformal continental, coastal and marine sedimentary facies (Jardim de Sá, 1978). Sedimentological studies in the Santo Onofre belt (in particular along the Santo Onofre river), indicate a succession of marine pelites to the west whereas to the east, fluvial-deltaic sediments are more abundant.

The Espinhaço Supergroup has been subdivided into Lower, Middle and Upper Espinhaço. To the west of the area under study, the Espinhaço range occurs as a long belt from north of Igaporã to Licínio de Almeida. Table 2.1 presents a summary of the proposed stratigraphy for the Espinhaço according to Schobbenhaus (1972), Schobbenhaus et al (1976) and Moutinho da Costa et al (1976).

The Lower Espinhaço is characterized by acid volcanic rocks (rhyolites and dacites) which in places show highly schistose textures. Undeformed varieties are porphyritic, with quartz or feldspar phenocrysts in a quartz-feldspar groundmass. Very deformed rocks comprise quartz-schists and phyllites, and their mineralogy includes muscovite quartz, feldspar, biotite, chlorite, chloritoid, garnet and subordinate kyanite and pyrophyllite. In particular along the extension Caetité-Licínio de Almeida, these rocks show mylonitic and phyllonitic textures. Close to

Licínio de Almeida, schistose volcanic rocks are associated with fuohsite-quartzites and itabirites of the Caetité-Jacaraci-Licínio de Almeida metamorphic complex, and they are also found intercalated with Middle Espinhaço rocks and/or basement rocks along thrust faults.

From south of Licínio de Almeida to Boquira in the north, the Middle Espinhaço has its contacts invariably along reverse faults; to the west, with rocks of the Upper Espinhaço and to the east, either with the Lower Espinhaço or with rocks of the Caetité-Jacaraci-Licínio de Almeida metamorphic complex or migmatites of the Paramirim. In general, either a polymictic conglomerate or a pure quartzite occurs at the base. Also present are sericite-quartzites with monomictic conglomerates, and intercalations of quartz-phyllites, quartz-schists, biotite-schists and feldspar-quartzites.

The Upper Espinhaço contains lithic arenites, feldspar arenites, graywackes, shales and siltites, locally graphitic and/or hematitic or manganese rich. The metamorphic effects produced a large variety of quartzites and phyllites: graphite-phyllites, hematite-phyllites, sericite-feldspar-quartzites, chlorite-quartz-schists. Chloritoid is observed in some metaphyllites.

Pre or post-tectonic diabases (now amphibolites), gabbros and subordinate serpentized ultramafic rocks are found in the Lower and Middle Espinhaço sequences. These form dykes, stocks and sills, and their chemistry points to a tholeiitic continental origin, and transition to alkaline types (Jardim de Sá, 1981). Syn and post-tectonic dykes of trachyte and rhyolite occur between Boquira and Macaúbas (Mc Reath et al, 1981).

The metamorphic grade of the Espinhaço rocks is greenschist facies with sericite, chlorite, graphite, muscovite and biotite. The presence of kyanite in the Middle (and Lower) Espinhaço (to the northwest of the area) remains to be interpreted. Kaul (1970, in Schobbenhaus et al 1976), suggested that it could be allothigenous. According to Jardim de Sá (1978), the presence of kyanite and epidote attest a high greenschist to low amphibolite facies of metamorphism, attained locally.

When comparing the northern portion of the Espinhaço range (north of parallel $14^{\circ}00'$) to the south, some striking differences are noticed in terms of their styles of deformation. Folds are open in the region close to Oliveira dos Brejinhos (in the north), becoming tight with reversal of strata to the south (from Boquira to Caetité). In the north, the Santo Onofre River fault is the most

important tectonic feature, with a complex history of transcurrent and orthogonal movements. To the south, a system of reverse and thrust faults developed along the eastern border of the range from Licínio de Almeida to a few kilometres north of Boquira. Along this extension, the Paramirim "block" was thrust to the west, over rocks of the Espinhaço Supergroup. Such movements printed regional foliations in the Espinhaço rocks, which are similar to those synchronically generated in the basement rocks. As a consequence of this same process, high grade basement rocks are locally altered to low amphibolite (Moutinho da Costa e Inda, 1982), whereas Middle Espinhaço rocks locally show kyanite attesting a low amphibolite or a greenschist (well into the kyanite zone) facies of metamorphism. To the north, the contact Espinhaço/basement is unconformable, and the Espinhaço strata dip 6° to 15° to the east.

2.4. Geochronology

The Precambrian of the São Francisco Craton in the state of Bahia can be subdivided in four major geotectonic cycles: the Jequié (2700 ± 0.2 Ma), the Tranzamazonian (2000 ± 0.2 Ma) and the Brazilian cycle (1000 ± 0.2 Ma) (Figueiredo, 1980). A large number of radiometric

determinations is available for the state, the great majority by Rb/Sr and K/Ar methods. Only the more significant data pertaining to the Lagoa Real region and its surroundings will be discussed.

In the Paramirim complex, a Rb/Sr isochron with $t = 2600$ Ma and initial $^{87}\text{Sr}/^{86}\text{Sr} = 0.703$ was obtained for gneisses, migmatites, granites and metavolcanic rocks, characterizing their formation and/or metamorphism in the Jequié cycle (Jardim de Sá et al, 1976 b). Unfoliated granites yielded ages of 1530 and 1080 Ma, with initial $^{87}\text{Sr}/^{86}\text{Sr}$ variable between 0.705 and 0.720 (Jardim de Sá et al, 1976 b). K/Ar ages in biotites, amphiboles and total rock (gneisses) are: 1830 ± 60 Ma and 1140 ± 35 Ma (Távora et al, 1967; see Fig. 2.2 and Table 2.2). East of Boquirã, K/Ar ages of 1100 and 815 Ma in biotites of granites and migmatites from the base of the Santo Onofre were encountered (Jardim de Sá et al, 1976b). Tranzamazonian cycle ages by Rb/Sr of 1900 and 2000 Ma are difficult to interpret: they could represent age of formation of these rocks or the age of reworking of Archean material (Inda e Barbosa, 1978 a, b). Brazilian cycle ages obtained by Rb/Sr are also reported in gneisses and pegmatites south of parallel $14^{\circ}30'$ (700 - 455 Ma); according to Inda e Barbosa (1978 a), these ages reflect the tectonic movements experienced by the "basement rocks

close to the Espinhaço/basement border, in the Brazilian cycle. Intrusive granites farther to the east (north of Paramirim) and to the north (close to the boundary basement/Chapada Diamantina) yield K/Ar ages of 580-475 Ma.

For the Espinhaço Supergroup, Rb/Sr ages published by Inda e Barbosa (1978 a, b) are as follows: 1100 Ma for rhyolites and tuffs at the base of the Chapada Diamantina; 1150 Ma for siltites from the Chapada; 580 Ma for phyllites and metamorphosed shales from the Espinhaço range, west of Macaúbas. Volcanic rocks of the "Boqueira Formation", lowest sequence at the base of the Espinhaço close to Macaúbas (Moutinho da Costa et al, 1976), show a range of Rb/Sr ages (1150 and 750 Ma) with points which are irregularly distributed on the isochrons (Jardim de Sá et al, 1976 a); K/Ar determinations produced ages of 580 to 520 Ma (Távora et al, 1967). Based on these values, it has been concluded that the volcanic rocks were formed at 1000 Ma and later altered/metamorphosed at 750 Ma (Jardim de Sá et al, 1976 a). In spite of the stratigraphy proposed by many authors for the Espinhaço Supergroup (see 2.3.), this issue remains very controversial. Stratigraphical correlations along the Espinhaço range as well as between lithologies from the Chapada Diamantina and the Espinhaço range, present a number of problems which makes

geochronological interpretations very difficult. In spite of that, many authors consider the period 1200-1300 Ma as the main tectonic and metamorphic event for the Espinhaço Supergroup, with post-tectonic events extending to 1000-950 Ma (Jardim de Sá et al, 1976 a, b; Jardim de Sá, 1978; Schobbenhaus et al, 1976; Inda e Barbosa, 1978 a).

A number of K/Ar age determinations was published by Távora et al (1967) and condensed by Schobbenhaus et al (1976). These can be found in Table 2.2 and Fig. 2.2.

Cordani (1982) undertook Rb/Sr age determinations of gneissic rocks close to the Lagoa Real ore deposits. Some of his results and interpretations are presented. An isochron defined by migmatites and augen-gneisses close to the town of Lagoa Real and S. Timóteo (see map, Fig. 5.1) yield an age of 2860 Ma with an initial $^{87}\text{Sr}/^{86}\text{Sr}$ ratio of 0.705. Samples collected close to Lagoa Real yield an age of 1300 Ma with higher initial ratios, suggesting reworking of the Archean rocks. Intrusive bodies close to the augen-gneisses border located west of S. Timóteo, define an isochron with $t = 1750$ Ma and initial $^{87}\text{Sr}/^{86}\text{Sr}$ ratio of 0.705. Microcline-gneisses from anomalies 9 and 13 yield an age of 1500 Ma, with an initial ratio of 0.723, and may represent the time of main regional metamorphism. On the other hand, microcline-gneisses from anomaly 3 align along another

Fig. 2.2. General geology of the region surrounding the Lagoa Real Province, with location of samples for K-Ar age dating (after Schöbenhaus et al, 1976). Results are displayed on Table 2.2.

Table 2.2. K-Ar age dates obtained for the region
surrounding the Lagoa Real Province
(see Fig. 2.2 for sample location).

Sample No.	Rock type	Mineral	Age (Ma)
3	pegmatite	muscovite	495 $\pm$ 15
5	gneiss	muscovite	505 $\pm$ 15
6	quartzite	muscovite	505 $\pm$ 15
12	gneiss	biotite	540 $\pm$ 20
13	amphibolite	hornblende	550 $\pm$ 20
21	amphibolite	hornblende	700 $\pm$ 35
23	gneiss	biotite	750 $\pm$ 25
24	gneiss	biotite	820 $\pm$ 25
27	migmatite	biotite	1090 $\pm$ 35
30	gneiss	biotite	1830 $\pm$ 60

isochron with $t = 1000$ Ma and initial $^{87}\text{Sr}/^{86}\text{Sr} = 0.72$, and may be interpreted as the minimum age of metasomatic alteration (?).

Cordani (1983) presented new data from Rb/Sr determinations done on samples hosting the U ore bodies. Microcline-gneisses from drill holes in anomaly 13 again yield an age of 1520 ± 25 Ma, with an initial $^{87}\text{Sr}/^{86}\text{Sr}$ of 0.7223; this age was again interpreted as being correspondent to the main metamorphic event responsible for the formation of these rocks. Samples of amphibolites intercalated with these microcline-gneisses in anomaly 13, yield an age of 624 ± 70 Ma, which is probably related to the period of Na-metasomatism.

In conclusion, Cordani (1983) suggested that the regional microcline-gneisses bordering the ore deposits, which originated from a magmatic precursor formed at 1750 Ma, were metamorphosed at 1500 Ma ("Espinhaço cycle") and then reworked in the Brazilian cycle; the uraniferous mineralization is probably associated with this last event.

U-Pb age determinations were carried out by Cooper (1979) in samples of common mineralized albitites from anomalies 1 and 3. He obtained an age of 820 Ma for the formation of uraninite.

CHAPTER 3

PETROGRAPHY

3.1. Introduction

Information in this chapter, as well as descriptive sections throughout this thesis, are based on the study and/or analyses of about 120 samples. These are the result of a few traverses undertaken across albitite bodies in anomalies 3, 5, 6 and 7, and traverses in the Archean-Lower Proterozoic gneissic terrain (see sample location map, Fig. 5.1). From anomaly 05, 27 samples from drill hole no. BR 01 were used for mineral and rock chemistry (Chapters 4 and 5).

This section presents observations based on petrography, but it should be noted that much of the detailed mineral identification was done by electron microprobe (Chapter 4).

3.2. Regional gneisses

In the vicinity of ore bodies, most of the regional gneisses studied are highly deformed coarse augen-gneisses and some banded gneisses (Plate 1.A). They are pinkish-grey to bluish-grey and brown in colour, massive, with predominance of feldspars and quartz, mafic minerals occurring as patchy concentrations. Some show foliation which is defined by the alignment of coarse mafic mineral concentrations.

The most common texture is porphyroclastic where megascopically visible lenticular porphyroclasts of K-feldspar, plagioclase, quartz and/or mineral agglomerates are separated by a groundmass of rather finely ground and less recrystallized material. The porphyroclasts make up more than 50% of the rocks; in most cases these rocks can be texturally classified as protomylonites and more rarely as microbreccias and blastomylonites (Spry, 1969, Higgins, 1971). The groundmass texture is very fine to medium mylonitic and/or granoblastic, depending on the degree of recrystallization; it is made up of quartz and feldspars.

Perthitic K-feldspar is by far the most abundant mineral. It ranges in size from 0.2mm and finer in the groundmass to about 10mm in the porphyroclasts. K-

PLATE 1

- A: Sample of regional gneiss; coarse grained gneissic rock made up of feldspars + quartz showing patches of mafic minerals (amphibole + biotite);
- B: Well foliated "transitional" gneiss, showing fine layers of feldspar + quartz intercalated with layers of amphibole + biotite;
- C: Sample of albitite showing fine to medium grained "sugary" albitic concentrations and fine to medium grained mafic mineral concentrations (pyroxene + garnet); an incipient foliation is noticeable;
- D: Same as C, with foliation well developed;
- E: Granoblastic texture in "transitional" rock. Note that microcline grains are well recrystallized;
- F: Albitic rim surrounding original plagioclase in "transitional" rock.

PLATE 1

feldspar porphyroclasts are cloudy, weakly altered to sericite and highly strained. Along their margins a certain amount of tectonic crushing and/or recrystallization is common. Medium sized grains can be hypidioblastic, but the majority of porphyroclasts are rounded or lenticular, and intergranular crushing with stress corrosion cracks (Barnett and Kerrich, 1980) are widespread. In a variety of samples, perthitic K-feldspar also occurs as aureoles or large xenoblastic rims surrounding hypidioblastic plagioclase in optical continuity; in places, only streaks of partly absorbed plagioclase crystals are seen (patchy-perthites, Spry, 1969), with typically sutured or corroded borders, suggesting substitution of plagioclase by K-feldspar. X-ray diffraction studies indicate that K-feldspar is probably orthoclase.

Plagioclase is hypidioblastic and ranges from 0.2mm and finer in the groundmass to about 6mm in the porphyroclasts. It is highly deformed with undulose extinction and twin lamellae that are bent and interrupted; intergranular crushing is also observed and plagioclase borders are usually serrated, ground and/or recrystallized. Anthiperthitic grains are frequently observed. Normal zoning (i.e. progressively more sodic towards the rim) is found in most unrecrystallized grains.

Plagioclase is always slightly altered to sericite (which predominates in the core of grains) and rarely to calcite and epidote. Myrmekitic intergrowths along the margins of feldspars are seen everywhere. Plagioclase in the regional gneisses is oligoclase or albite-oligoclase.

Quartz occurs in different shapes and sizes. In some very cataclastic rocks (microbreccias), medium to coarse (up to 5mm) quartz is xenoblastic/rounded with strong undulose extinction and comminution along the borders. It can also appear xenoblastic, highly strained with sutured boundaries partly recrystallized. Quartz is however most commonly seen in aggregates, as strained elongate ribbons (0.5 to 2mm in size), and as fine recrystallized grains in the matrix producing a mortar texture. Globular quartz is common as inclusions in other minerals.

Amphibole (hastingsite and also Fe-pargasite) and biotite appear as lenticular or ribbon-like aggregates of fine to medium-grained flakes, fibres or prisms, defining an incipient foliation (fluxion structure). They are also present as clusters of somewhat prismatic crystals surrounding porphyroclasts. Amphibole is medium to coarse grained and evidence of its replacement by biotite is widespread; it is deep greenish-blue to deep olive green and very dark green. A large number of grains are poikiloblastic showing minute inclusions of quartz and

sphene. Amphibole replacement by biotite is observed along cleavage lines and borders, accompanied by iron release in the form of limonitic patches. Biotite flakes ($\pm 0.8\text{mm}$) are light brown to dark brownish green, kinked or bent, though in general they are largely recrystallized; streaks of opaque mineral (probably magnetite) and very fine sphene altered to leucoxene occur along cleavage planes. Biotite can sometimes alter to muscovite, chlorite or epidote. Sphene, rare calcite and opaques are commonly associated with mafic concentrations.

Sphene is the most abundant accessory mineral and as already mentioned, it is found associated with amphibole and biotite either as minute granular crystals or as fine hypidioblastic grains. It is also very common as fine to medium grained xenoblastic aureoles surrounding ilmenite and/or magnetite.

Other accessory minerals are: ilmenite, magnetite, zircon, apatite, allanite, fluorite and rare pyrite and chalcopryrite. They are invariably associated with the mafic concentrations.

Granulite facies gneisses are found sporadically as mentioned in Chapter 2. They are green to greenish-grey rocks, medium grained, massive. In thin section they show medium to fine grained granoblastic texture with

pronounced orientation of pyroxenes and feldspars. Their mineralogy includes: plagioclase, pyroxenes, perthitic K-feldspar, quartz, biotite and accessory magnetite, ilmenite and pyrite. Plagioclase is andesine in hypidioblastic crystals slightly altered to sericite. Pyroxenes are hypersthene and augite, granoblastic and usually intergrown, showing restricted transformation to reddish-brown biotite along their borders. Perthitic orthoclase is hypidioblastic and granoblastic. Quartz is rounded, fine grained. The few samples that have been studied can be classified as enderbites (Streckeisen, 1974)..

3.3. "Transitional" rocks

If one examines the sections based on drill cores across various albitite bodies, one finds a series of foliated k-feldspar+quartz rich bands which have been generically named "microclinites" or "albite-microcline-quartz-metasomatites" (Geisel et al, 1980). These rocks are always intercalated with different types of albitites and show a number of petrographic as well as chemical aspects that seem to indicate their intermediate nature between the regional gneisses so far considered, and the albitites sensu stricto. For this reason, they have been

considered as transitional and their petrographic characteristics are enlightening as far as mineral transformations are concerned.

The "transitional" rocks are usually well foliated, fine to medium grained, pinkish-grey to very light grey and pink in colour with predominance of feldspars and quartz, mafic minerals occurring in fine bands whose alignment defines the foliation (see Plate 1.B). Folding of the foliation is sometimes observed.

The predominant texture is fine to medium granoblastic, and polygonal granoblastic in places, with recrystallization being dominant over cataclasis (Plate 1.E). Porphyroclasts of K-feldspar, quartz and plagioclase if present, are generally smaller than 4mm and show recrystallization. The foliation is given by the preferential alignment of mafic minerals, recrystallized feldspars and quartz, and medium to coarse grained ribbon quartz.

The proportions of each of the present minerals is variable: in some of these gneisses, K-feldspar is the predominant mineral together with quartz, plagioclase and amphibole+biotite, being mineralogically and chemically similar to the regional gneisses. A number of samples however, show a decrease in the proportion of K-feldspar

and sometimes quartz, and in these, plagioclase predominates together with amphibole and/or biotite. Independently of whether K-feldspar or plagioclase predominates, garnet and/or pyroxene replacing amphibole, have been observed in some samples. A few of these, where plagioclase may reach about 70%, have been named "albitites". Even though they are composed of plagioclase + pyroxene and/or garnet, microprobe analyses have revealed that plagioclase is not pure albite and that pyroxene and garnet have different compositions from those encountered in albitites sensu stricto (see Chapter 4). In the present study, such gneissic rocks which are mineralogically and chemically very similar to the regional gneisses will be referred to as Type A "transitional" rocks. Those that present more advanced stage of alteration, with amphibole being replaced by garnet and/or pyroxene will be referred to as Type B "transitional" rocks. Their petrography will be described together.

Recrystallization of microcline is almost complete giving place to fine (0.3-0.8mm), clear granoblastic, at times polygonized, perthitic microcline crystals forming agglomerations everywhere. Porphyroclasts are rarely found: they range from 2 to 4 mm, are round and xenoblastic, highly deformed showing recrystallization.

Plagioclase is mostly seen as clear fine to medium (0.2 to 1.2 mm) hypidioblastic, granoblastic and polygonal granoblastic grains, sometimes showing normal zoning. Unrecrystallized plagioclase is xenoblastic, highly deformed and altered to sericite, calcite and more rarely to epidote (generally in central parts of grains); in a few samples it is found coarse grained, constituting porphyroclasts. Though rarely observed, microcline can occur engulfed by plagioclase suggesting substitution of the former by the latter. A very important feature common to a large number of plagioclase grains, is the presence of rims of untwinned albite surrounding them, in optical continuity with the centre of the grain (overgrowth). The plagioclase in these rocks is oligoclase, but in the majority of samples studied, grains of albite are found as well (see Plate 1.F).

Amphibole (mostly hastingsite) and biotite look and occur in the same way as they do in the regional gneisses. In a restricted number of samples (type B "transitional" rocks), amphibole is seen being replaced by garnet and/or pyroxene along cleavage lines and borders in association with accessory fluorite (Plate 2.A).

Garnet is rarely present and if so, it appears in an almond-like colour, fine xenoblastic, surrounding, interstitial and included in amphibole (in type B

PLATE 2

- A: Amphibole (hastingsite) being replaced by a grossular/
andradite garnet in type B "transitional" rock;
amphibole: in black
- B: Zoned grain of garnet replacing hastingsitic amphibole
in type B "transitional" rock. Note that light coloured
central part grades into darker borders (see section
4.2. c for exact compositions);
- C: Hedenbergitic pyroxene + fluorite replacing original
hastingsitic amphibole in type B "transitional" rock;
amphibole: in black
- D: Polygonal granoblastic texture in albitite;
- E: Zoned grain of aegirine-augite in albitite, showing
incipient alteration to secondary amphibole (black
prismatic inclusions); see section 4.3. b for exact
compositions);
- F: Fine xenoblastic andradite replacing aegirine-augite
in albitite; fine grained black crystals are magnetite.

PLATE 2

"transitional" rocks). It is also medium to coarse grained, poikiloblastic at times, again associated with amphibole, opaques, sphene and commonly altered to epidote. In one sample a few medium grained zoned crystals of garnet were encountered: they show pink almandine rich cores which grade into light brownish grossular/andradite rich borders (see Plates 2A and 2B).

Like garnet, pyroxene is rare. It is a pale green to green pyroxene occurring hypidioblastic, fine to medium grained altering to hornblende (a little) and frequently to epidote. When zoned, it shows almost colourless to very pale green cores that grade into slightly greener borders. It can be spotted by fine magnetite which is usually altered to sphene. Textural relations seem to suggest that pyroxene and garnet are replacing original hastingsite (see Plate 2.C).

Epidote is fine granular, colourless to very light green always associated with mafic minerals like amphibole, biotite, garnet and pyroxene. In some type B "transitional" rocks, pyroxene and/or garnet can appear being replaced by epidote and in these samples epidote may occur in as much as ~10%.

Sphene is the most abundant accessory mineral and its distribution as well as occurrence is the same as in the

regional rocks. Other accessory minerals are: magnetite, fluorite, zircon, apatite, allanite and rare pyrite, chalcopyrite and pyrrhotite.

3.4. Metasomatic Rocks

In the present study, metasomatic rocks are those which have undergone total or almost total chemical and mineralogical changes when compared to either the regional or the "transitional" rocks. As mentioned in Section 3.3., Type B "transitional" rocks show partial replacement of the original hastingsitic amphibole as well as a new generation of albitic plagioclase in the form of rims around original albite-oligoclase or oligoclase. As the process proceeds, the replacements and/or transformations of the original mineralogy tends to be completed, the final product being partially or completely albitized rocks. Such albitization process is accompanied by other types of reactions which will be discussed in Chapter 6.

For the purpose of simplifying their petrographic description, the metasomatic rocks will be divided into: albitites (Ca-rich albitites and mineralized albitites discussed here) and "epidosites".

3.4.1. Albitites

The albitites constitute a range of mineralogical types and attempting to divide them into separate units is mostly complicated by the transitional nature of their contacts. Nevertheless, a number of textural features are common to essentially all rock types. Albitites are white to light grey in colour, fine to medium and less coarse grained gneissic rocks where white to grey albite predominates along with patches or bands of mafic minerals (Plates 1C, 1D); some samples have a light green colour (green albite) which may be due to the presence of uraninite in the feldspar. In hand specimen the foliation is usually inconspicuous but at the outcrop a general N-S orientation is obvious. Very finely foliated albitites are found in places. Augen-like structures similar to those found in the regional gneisses are commonly seen except that each augen is made up of "sugary" albite (see Plate 2.D).

The most common and typical texture is "open" perfect polygonal granoblastic, fine to medium grained texture (Plate 2.D), with albite commonly showing an overall preferential alignment of its crystals. Less perfect polygonal granoblastic textures are also encountered. Porphyroclastic textures are very rare, with porphyroclasts of albite ranging from 1 to 6mm displaying

remarkable examples of subgrain nucleation (Vernon, 1976); they show bent twin lamellae and undulose extinction.

In general, the albitites could be divided into two main groups:

- albite rich rocks or albite-gneisses, where albite occurs in less than about 50%;
- albitites sensu stricto, where albite exceeds 50%. Zones of pure albitites are known in the region and in these, apart from accessory minerals, albite reaches almost 100%.

Among the albite rich rocks one encounters: garnet-pyroxene-albitites, garnet-pyroxene-calcite-albitites, garnet-pyroxene-magnetite-albitites, pyroxene-amphibole-albitites, amphibole-biotite-albitites, wollastonite-pyroxene-garnet-albitites and biotite-albitites. Among the albitites pyroxene-garnet-albitites and magnetite-albitites are probably the most common. Table 3.1 displays a series of mineral assemblages with modal compositions.

Polygonal equigranular albite occurs in clear 0.2 to 1.5mm crystals commonly untwinned and it can range from about 20 to 95-100% in a variety of bands, zones or agglomerations that are intercalated with the mafic minerals concentrations; albite grains are usually

Table 3.1 - Examples of mineral assemblages in the regional, "transitional" and metamorphic rocks, with percentages.

Rock type	Sample no.	ol	alc	qr	am	bl	pyr	que	ep	cal	mol	spn	mg	lim	hem	all	chl	sil	ep	or	vesu
Regional gneisses	4433	10	40	30	10	10															
	4437	10	40	40	5	5															
	4440	25	15	25	10	5															
Type A "transitional"	14, 0a-																				
	14, 10a	25	50	10	5	5						5									
	19, 00-																				
	19, 10a	50	10	20	1	10															
	107, 00-																				
Type B "transitional"	107, 10a	10	25	10	2																
	107, 11-																				
	107, 10a	22	10	10	2		1					2	1								
	124, 95-																				
	129, 00a	10	1	10	10		1	40	2	1											
Albitites	146	10	30	35			20	5													
	146, 00-																				
	146, 10a	62					5	25				1	5								
	147, 60-																				
	147, 72a	95																			
	145, 40-																				
	145, 20a	28	2				1	10		5											
	145, 62-																				
Car-rich albitite	145, 20a	62	1			5	15					2									
	145, 1	56					15	15		5	10										
	145, 1																				
Gabbro	14, 4a	15		15			25		25												
	75, 12-																				
	75, 30a	50		10			20		20												

ol: plagioclase; alc: microcline; qr: quartz; am: amphibole; bl: biotite; pyr: pyroxene; que: quartz; ep: epidote; cal: calcite; mol: wollastonite; spn: spinel; mg: magnetite; lim: ilmenite; hem: hematite; all: allanite; chl: chlorite; sil: sillimanite; flu: fluorite; ep: epidote; chl: chlorite; sil: sillimanite; vesu: vesuvianite; or: orthopyroxene; py: pyroxene.

a: accessory mineral

aligned defining an incipient foliation.

The mafic minerals tend to form concentrations and segregations which in some samples are aligned parallel to the general foliation. Among these, pyroxene and garnet are the most common and abundant. Pyroxene is green to yellow-green and pale green, and sometimes dark green; in thin section it can be classified as aegirine-augite. It occurs fine to coarse grained, granoblastic and hypidioblastic, in most samples altered to amphibole (hornblende) along cleavage lines and borders at various degrees; it can show inclusions of and be intergrown with sphene, opaques and less with calcite; in some samples it alters to epidote. Pyroxene can appear zoned, with a very light green core and darker green borders (Plate 2.E). In one sample studied (HBG 956), pyroxene alters not only to amphibole but also to aggregates of prehnite.

Andradite garnet is dark brown granoblastic, fine to coarse grained invariably intergrown with, surrounding and surrounded by pyroxene. It is sometimes altered to calcite and most times contains inclusions of sphene and opaques.

In almost all samples dominated by either garnet or pyroxene, these minerals occur together always intergrown. The textural evidences suggest that andradite is

formed after pyroxene (Plate 2.F).

Amphibole from pyroxene is fine grained, prismatic hornblende, bluish-green to brownish-green and also very dark green. It can occasionally appear coarse, as a major constituent (up to about 25%). Though rare, the alteration of amphibole to biotite and/or chlorite, and calcite is sometimes observed. Such late alteration is in places very advanced and both biotite and/or chlorite can reach about 20% of the rock. Biotite is dark green to very light greenish-brown appearing in aggregates of medium and fine tabular grains. In other samples, biotite is brownish-red to light yellow in association with calcite and opaques; in rare cases, biotite occurs in dark brownish-red flakes as an alteration product after magnetite. Chlorite is green to brownish-green in flake-like aggregates associated with biotite, epidote and calcite.

Epidote (pistacite) is at times seen as fine granular crystals after pyroxene, amphibole and garnet in amounts up to 1-2%. Very fine zoned grains of epidote occur as accessory in a number of samples, bearing no relation to any other minerals; its core is light brown with a light greenish-yellow border.

Granoblastic calcite is sparsely found, generally associated with mafic concentrations (mostly with garnet). It has been observed forming vein-like bands also associated with mafic minerals in up to about 15% of the rock. Late generation calcite veins are also encountered.

Fine interstitial, granoblastic and venular microcline surrounds albite crystals and is commonly seen being substituted by the latter (Plate 3.A); it rarely exceeds 3-5% of the rock. In some rare samples, microcline is observed to appear up to 25-30%, in clear polygonized fine to medium grains; these rocks can contain appreciable amounts of quartz (about 10-15%).

Quartz rarely amounts to more than 3%. It appears fine xenoblastic interstitial to feldspar and as globular inclusions in it. Quartz veinlets parallel and cross-cutting the main foliation are found in some albitites; it is xenoblastic fine to coarse grained with strong undulose extinction and may be of late formation. In samples where garnet seems to form after pyroxene, magnetite crystals included in either mineral are commonly surrounded by quartz.

Prehnite was observed in one mineralized sample replacing pyroxene, and in another one where it forms a micro-vein parallel to the foliation. It replaces

PLATE 3

- A: Fine interstitial microcline being replaced by albite
in a pyroxene-andradite-albite;
- B: Prismatic wollastonite crystal in a polygonal
granoblastic Ca-rich albite;
- C: Uraninite crystals occurring along andradite border
in a pyroxene-andradite-albite;
- D: Uraninite crystals occurring along albite borders and
also as inclusions in the latter;
- E: Idioblastic to hypidioblastic magnetite exhibiting
alteration to hematite, mainly along the borders;
magnetite: grey
hematite: white
- F: Idioblastic uraninite grains occurring associated
with hematite;
uraninite: dull grey
hematite: white

PLATE 3

pyroxene along borders and cleavages, usually in contact with plagioclase "corroding" its borders. Prehnite occurs in colourless grains forming tabular groups and barrel shaped aggregates, with an imperfect columnar "bow-tie" appearance. It contains no inclusions of uraninite.

Accessory minerals are listed in section 3.4.2.

a) Ca-rich albitites

A number of albite rich rocks with abnormally high CaO, as well as fairly high MgO values in whole rock (see Chapter 5), appear as part of the ore zone in anomalies 7 and 13. They are texturally and mineralogically much like all albitites except for the fact that both pyroxene and garnet have slightly different compositions, and because of the presence of wollastonite in some of the samples; vesuvianite has also been encountered. Garnet, pyroxene, wollastonite, calcite, sphene and vesuvianite form large patchy concentrations and segregations of intergrown crystals weakly defining the foliation. Garnet is very light brown to yellowish-brown and pyroxene is very light green to colourless (diopsidic).

If present, wollastonite forms perfect prismatic and sometimes fibrous crystals fine to medium grained, intergrown with the mafics and also with calcite (Plate

3.B); it is slightly altered to fibrous pectolite along the borders. Calcite occurs both as polygonal fine to medium grained crystals, and as fine xenoblastic inclusions in garnet suggesting its secondary formation.

Vesuvianite was observed in two samples only (AN 7 LL45-1, AN 7 LL45-14). Vesuvianite is medium grained, hypidioblastic and round. In the first sample, it appears almost colourless to very pale yellow, weakly birefringent and surrounded by garnet; it contains minute inclusions of very altered sphene (leucoxene ?). In the second sample, it is light yellow, weakly birefringent to isotropic associated with and surrounded by either garnet or pyroxene; again, it contains minute inclusions of altered sphene, as well as inclusions of quartz, garnet, pyroxene and perhaps uraninite (?). Some grains of vesuvianite seem metamict and resemble allanite. In this last sample, wollastonite is absent.

b) Mineralized albitites

In the present study, mineralized albitites are thus referred to if they contain uraninite. Mineralized albitites occur as parallel bands averaging 6m thick within the albitized zones, with no significant variations down dip to depths of 800m (Raposo e Matos, 1982). Their textural features are like those of all other common

albitites. Mineralized albitites always show either one or more of the following: aegirine-augitic pyroxene, andraditic garnet, amphibole, and magnetite (always altered to hematite), and perhaps epidote, biotite and chlorite. Pure albitites are found, not to be mineralized. By far the most common types of mineralized albitites are the pyroxene-garnet-albitites and magnetite-hematite-albitites.

Uraninite occurs in round crystals $\pm 5-25 \mu\text{m}$ across associated with mafic concentrations, mainly garnet and also pyroxene, but also interstitial or included in albite and hematite (see Plates 3.C, 3.D, 3.F). It forms trails of minute crystals mostly inside garnet, but also inside pyroxene, albite (along crystal borders and as inclusions along fine cracks), amphibole and sphene, and at times inside microcline; inclusion of uraninite inside biotite is uncertain, because in many cases uraninite seems to be associated with neighbouring crystals like albite and/or amphibole. In CaO rich albitites, uraninite is constantly included and associated with garnet, and less with pyroxene, sphene and perhaps (?) vesuvianite, but never with calcite or wollastonite. Uraninite is at times very altered to β - uranophane which constitutes minute yellow needles or fibres, forming vein-like crusts that cross all directions.

Accessory minerals are the same as for all other albitites. It is however very common to find metamict sphene in mineralized albitites, associated with mafic minerals (mainly garnet but also with pyroxene). See 3.4.2. for complete list of accessory minerals.

3.4.2. Epidosites

Two types of epidote rich rocks are restrictively found in the area, the first (type 1) being massive, fine grained, dark brown in colour (usually mineralized), and the second (type 2) is a fine grained, light green coloured rock (unmineralized). Both types form bands or layers intercalated and inside oligoclase rich zones (oligoclasite zones ?); they can vary from a few centimetres to 1 or 2 metres in thickness parallel to the main foliation wedging out at depths of about 100m (K. Fuzikawa, personal communication).

They are polygonal granoblastic rocks and their mineralogy includes: epidote (replacing pyroxene, 50-80%), plagioclase (20-30%), pyroxene (aegirine-augite, 2-15%), quartz (5-35%), hornblende (~1%, alteration from pyroxene) and accessory microcline, sphene, zircon, apatite and opaques; sericite from plagioclase and micaceous minerals from pyroxene can also be found. Uraninite is at times observed to appear in type 1 epidosites.

In type 1 epidiosites, pyroxene is light green to colourless. It is always seen as fine xenoblastic remnant grains surrounded by "masses" of epidote. At closer inspection, epidote "masses" constitute groups of fine grained round and granoblastic epidote crystals, pseudomorphing pyroxene. Epidote has a dull brownish-yellow colour, no pleochroism and very low interference colours (which usually decrease outward in each grain); core of grains may exhibit typical pyroxene birefringence.

In type 2 epidiosites (one sample only; RA 87 - 75.32-75.38m), pyroxene is found fine to medium grained, light green to bluish-green granoblastic and polygonal granoblastic. Pyroxene is being replaced by epidote and less by hornblende, but in this case such replacement does not constitute perfect pseudomorphism. Epidotes form fine granoblastic bright lemon yellowish-green crystals; they replace pyroxene along crystal boundaries and cleavage planes (commonly seen as minute inclusions in the latter). Epidote is weakly pleochroic and shows very high birefringence.

Both plagioclase and quartz are granoblastic, fine to medium grained. Microprobe analyses of plagioclase have shown that its composition is oligoclase and not pure albite.

In all types of albitites as well as epidiosites studied, accessory minerals may include: magnetite and sphene (each may sometimes reach about 5%), fluorite, allanite, zircon, apatite, pyrite and muscovite (both rare). The most abundant are sphene and magnetite. Sphene is fine to medium grained either occurring as isolated hypidioblastic and xenoblastic crystals or fine xenoblastic, associated with all other mafic minerals in general. Magnetite occurs both idioblastic, very fine to fine associated with and included in garnet and pyroxene (where it appears surrounded by a quartz pellicle), and idioblastic fine and medium grained disseminated in the rock. Magnetite is commonly being substituted and surrounded by sphene. Martitization of magnetite is very common in both types of magnetites; in some samples it is restricted to the borders of the grains and in some others such transformation can be complete or almost complete (as such, only observed in mineralized rocks; see Plates 3.E, 3.F).

3.5. Summary

Amphibolite facies regional gneisses have been recrystallized and Na-metasomatized, giving place to albitites or albite-rich rocks that are locally enriched

or mineralized in uranium.

The regional rocks are highly deformed coarse to very coarse augen-gneisses showing incipient foliation. Their textures are mostly porphyroclastic and mylonitic/cataclastic partly recrystallized. Porphyroclasts of K-feldspar and plagioclase can reach up to 10mm, very strained and crushed along the borders. Their mineralogy includes: perthitic K-feldspar (probably orthoclase, 45-70%), quartz (15-30%), albite-oligoclase and oligoclase (20-30%), amphibole (hastingsite altering to biotite; both amount to 10-20%), with accessory sphene, ilmenite, magnetite, zircon, apatite, allanite, fluorite and rare sulfides.

As one approaches the zones of alteration, a number of fine to medium grained finely foliated rocks are encountered constituting bands that are intercalated with levels of albitized rock. Because some of these present distinct intermediate mineralogical and chemical characteristics between the regional gneisses and the albitites, they are referred to as "transitional" rocks. Most of the studied samples have a number of mineralogical as well as chemical similarities to the regional gneisses, but their dominant texture is granoblastic and polygonal granoblastic with recrystallization dominant over cataclasis. They contain: microcline, quartz, amphibole

(hastingsite altering to biotite), with accessory sphene, magnetite, zircon, apatite, allanite, fluorite and rare sulfides; these have been named type A "transitional" rocks. An important feature in almost all "transitional" rocks is the presence of rims of albite surrounding a number of plagioclase grains in optical continuity. Some of the rocks studied, may contain garnet and/or pyroxene, which seem to be replacing original hastingsitic amphibole; these have been denominated type B "transitional" rocks. In such cases, garnet is grossular/andradite together with Fe-salitic or hedenbergitic pyroxene.

Once all original mineral assemblages have been modified, the resulting products (metasomatic rocks) constitute albitites which are locally mineralized, and a group of epidote rich rocks (epidosites). The albitites are fine to medium grained gneissic rocks showing incipient and at times pronounced foliation. They are typically polygonal granoblastic with albite many times displaying a preferred alignment of its crystals. Their mineralogy includes: albite, pyroxene (aegirine-augite altering to hornblende, epidote and less to calcite), garnet (andradite from pyroxene; it may alter to epidote or calcite), hornblende (in very few samples altering to biotite or chlorite and calcite), biotite, chlorite,

calcite (also as late veins), microcline (interstitial, up to 5%, and more rarely in recrystallized portions, in as much as ~15%), quartz (up to 3% and as late veins and agglomerations). Accessories are: sphene, magnetite (at times altering to hematite), fluorite, allanite, zircon, apatite, pyrite.

A group of albitites rich in both CaO and MgO in whole rock (being referred to as Ca-rich albitites), have garnet that is grossular/andradite and pyroxene that is richer in MgO, with the eventual occurrence of wollastonite and vesuvianite; these albitites contain abundant calcite (~20%). Almost pure albitites are found as well.

Epidosites occur very locally in the region; they are greenish-brown rocks, fine to medium granoblastic, containing about 50-80% epidote. In mineralized epidosites (type 1), brown epidote (showing very low birefringence) pseudomorphs colourless pyroxene; uraninite is at times present, but more commonly U is found in the epidote structure. In type 2 epidosites (unmineralized), lemon green epidote (common pistacite) replaces green pyroxene along borders and cleavage lines. Both types of epidosites contain up to 35% quartz and ~25% plagioclase (which is high oligoclase).

Mineralized albitites (albitites containing uraninite), are basically like all other albitites except that they always contain either one or more of the following: aegirine-augitic pyroxene, andraditic garnet, magnetite (always altered to hematite), amphibole and perhaps epidote. The most common types of mineralized albitites are: aegirine-augite-andradite-albitites and magnetite-hematite-albitites. Uraninite occurs as round crystals, 5-25 μ m across associated with mafic minerals, mostly garnet, pyroxene and martitized magnetite and also included in and interstitial to albite and hematite. In Ca rich-albitites, uraninite is invariably associated with garnet, and less with pyroxene, albite, sphene and perhaps vesuvianite (not so clear), but never with calcite or wollastonite.

CHAPTER 4

MINERAL CHEMISTRY

Comprehensive tables of representative electron microprobe analyses of the minerals discussed in this chapter are presented in Appendix I ; selected analyses are presented within the text. Iron was determined as total Fe (as FeO).

4.1. Regional Gneisses

a) Plagioclase

Plagioclase in the regional gneisses predominates as porphyroclasts but it is also found fine to medium grained, in the matrix. Their composition is either albite with significant calcium content (An ~3-5 mole%) or oligoclase with anorthite content between 11-18%. Centres of zoned crystals, mostly in the matrix, are composed of oligoclase with An component of ~11% and albitic borders

(An ~ 28). Results are displayed on Table 4.1.

b) Amphibole

Amphibole occurs in lenticular or ribbon-like aggregates commonly altered to biotite. Selected analyses can be found in Table 4.2. Amphibole belongs to the calcic amphiboles group (Leake, 1978, Hawthorne, 1981) being part of the pargasite-hastingsite compositional series (Deer et al, 1976) of the hornblendes. Amphiboles of this series are commonly found in metamorphic rocks, like skarns, some gneisses, amphibolites and granulites (Thomas, 1982).

Using Papike et al's (1974) characterization which estimates Fe^{+3} , the best end-member names are: hastingsite and some Fe-pargasite with high Fe^t/Mg ratios ranging from 15 to 60. Fig. 4.1 displays Fe^t/Mg against Fe^t/Fe^t+Mg ionic ratios (Fe^t = total iron).

c) Biotite

Biotite forms light brown to dark brownish-green flakes after amphibole. Chemical analyses are set down in Table 4.3. Biotites are very rich in FeO and some are extremely poor in MgO with Fe^t/Mg ratios ranging from 9.5 to 36.39 and $Mg/Mg+Fe^t$ varying from 0.03 to 0.2 (see Fig. 4.2). They could be termed iron-biotites (Foster, 1960);

Table 4.1. Selected analyses of plagioclases in regional rocks.

Sample no.	1 AM8	2 AM37	3 AM37	4 AM40	5 AM40
SiO ₂ (wt%)	64.08	67.34	67.79	65.68	68.05
Al ₂ O ₃	22.27	19.93	19.93	21.52	19.63
CaO	3.48	0.77	0.47	2.61	0.44
Na ₂ O	10.11	11.60	11.18	10.82	11.55
K ₂ O	0.08	0.10	0.09	0.07	0.05
Total	100.02	99.74	99.46	100.70	99.72
Ions on the basis of 32 oxygens					
Si	11.322	11.837	11.905	11.501	11.931
Al	4.637	4.128	4.124	4.441	4.056
Ca	0.659	0.145	0.088	0.490	0.083
Na	3.463	3.953	3.807	3.673	3.926
K	0.018	0.022	0.020	0.016	0.011
An%	15.91	3.52	2.26	11.72	2.06
Ab%	83.65	95.94	97.23	87.91	97.67
Or%	0.44	0.34	0.51	0.37	0.28

1: Oligoclase in groundmass; 2: Albite grain, centre;
 3: Albite grain, border; 4: Zoned plagioclase, centre;
 5: Same as 4, border.

Table 4.2. Representative analyses of amphiboles from regional gneisses.

Sample no.	1 AM30	2 AM8	3 AM33	4 AM34	5 AM37
SiO ₂ (wt%)	39.67	37.54	37.92	38.54	37.09
Al ₂ O ₃	10.14	12.17	11.78	12.04	11.99
FeO*	32.18	31.43	31.34	30.59	32.17
FeO*	29.28	29.12	29.32	28.23	28.57
Fe ₂ O ₃ *	3.22	2.57	2.24	2.62	4.00
MgO	1.25	0.27	0.59	1.08	0.41
MnO	0.51	0.47	0.38	0.59	0.36
TiO ₂	1.14	0.94	0.95	0.36	0.44
CaO	10.52	10.49	10.52	10.54	10.63
Na ₂ O	1.78	1.64	1.71	1.56	1.61
K ₂ O	1.77	2.03	2.11	2.37	2.32
Total	99.28	97.23	97.52	97.93	97.42
Ions on the basis of 23 oxygens					
Si	6.357	6.159	6.202	6.246	6.102
Al	1.643	1.841	1.798	1.754	1.898
Al	0.272	0.513	0.473	0.547	0.427
Fe ²⁺	3.925	3.996	4.010	3.927	3.931
Fe ³⁺	0.388	0.317	0.276	0.319	0.496
Mg	0.298	0.066	0.144	0.261	1.005
Mn	0.069	0.065	0.053	0.081	0.050
Ti	0.137	0.116	0.117	0.044	0.054
Exc. Oct.	0.090	0.073	0.073	0.079	0.059
Ca	1.906	1.942	1.943	1.930	1.974
Na-Mg	0.103	0.094	0.083	0.091	0.067
Na-A	0.449	0.434	0.459	0.399	0.446
K	0.362	0.425	0.440	0.490	0.447
Fe ²⁺ /Mg	14.68	66.30	30.17	16.20	44.52
Fe ²⁺ /Fe ²⁺ +Mg	0.94	0.98	0.97	0.94	0.98

2,3,4: Ferro-margasite; 1,5: Hastingsite.

*Calculated according to Papike et al. (1974).

Fe²⁺: total iron.

Table 4.3. Representative analyses of biotites from regional gneisses.

Sample no.	1 AM30	2 AM8	3 AM33	4 AM34	5 AM37
SiO ₂ (wt%)	34.72	33.70	34.14	35.48	34.09
TiO ₂	4.27	2.49	2.62	2.08	1.87
Al ₂ O ₃	12.15	15.62	13.04	14.14	14.69
Cr ₂ O ₃	0.35	0.00	0.00	0.07	0.00
FeO*	35.46	34.64	35.34	33.65	34.41
MnO	0.47	0.38	0.40	0.53	0.64
MgO	1.55	0.54	0.97	2.02	1.14
CaO	0.02	0.00	0.00	0.08	0.04
Na ₂ O	0.06	0.10	0.05	0.00	0.08
K ₂ O	9.09	8.83	9.79	8.66	9.24
Total	98.14	96.30	96.35	96.71	96.20
Ions on the basis of 22 oxygens					
Si	5.616	5.509	5.651	5.723	5.597
Al	2.316	2.491	2.349	2.277	2.403
Al	0.000	0.517	0.194	0.411	0.439
Ti	0.519	0.306	0.326	0.252	0.231
Mg	0.374	0.132	0.239	0.486	0.219
Fe	4.797	4.735	4.892	4.539	4.724
Mn	0.064	0.053	0.056	0.072	0.089
Cr	0.045	0.000	0.000	0.009	0.000
Na	0.019	0.032	0.016	0.000	0.025
Ca	0.003	0.000	0.000	0.014	0.007
K	1.875	1.841	2.067	1.782	1.935
Fet/Mg	13.01	36.39	20.68	9.49	17.25
Fet/Fet+Mg	0.93	0.97	0.95	0.90	0.945

*FeO calculated as total iron.

Fet: total iron.

1, 2, 3, 4, 5: Biotite from amphibole.

Fig. 4.1. Fe^t/Mg versus $\text{Fe}^t/(\text{Fe}^t + \text{Mg})$ ionic ratios of amphiboles in regional and "transitional" rocks. Only a few analyses of amphiboles replacing pyroxene in metasomatic rocks, have been included for comparison. Fe^t = total iron.

Symbols for this figure as well as for all figures displaying either results from mineral chemistry and rock chemistry (this Chapter and Chapter 5), are as follows:

- △ regional rocks
- type A "transitional" rocks
- type B "transitional" rocks
- albitites (unmineralized)
- ★ albitites (mineralized)
- ✦ Ca-rich albitites (unmineralized)
- ✧ Ca-rich albitites (mineralized)
- type 1 epidiosites (mineralized)
- type 2 epidiosites (unmineralized)

Fig. 4.2. Fe^t/Mg versus $\text{Fe}^t/(\text{Mg} + \text{Fe}^t)$ ionic ratios of biotites in regional, "transitional" rocks and common unmineralized albitites. For symbols, see Fig. 4.1.

they have a range of Mg (atomic) of less than 1.00 as expected in iron-biotites. In fact, sample AM 8 could probably be named lepidomelane according to Foster (1960); its Mg ionic value is less than 0.15 and Al is very low.

Based on a compilation of a large number of biotite analyses from metamorphic rocks, Guidotti (1984) has concluded that $Mg/Mg+Fe^t$ of biotites in common metamorphic parageneses is almost always >0.1 , and this could be a function of bulk composition controls. Foster (1960) reports Mg poor biotites ($MgO < 1.5\%$) as being typical of rocks like pegmatites, some nepheline syenites (pegmatitic) and greisens.

In spite of such considerations, it can be suggested that the compositions of the biotites here studied reflect those of the amphiboles from which they alter. This can be demonstrated in Fig. 4.3 where Fe^t/Mg ratios of both minerals are plotted.

d) Ilmenite

Fine grained ilmenite, commonly surrounded by sphene was analysed in sample AM 30. It contains 46.93% FeO and 2.30% MnO (Table 4.4).

Fig. 4.3. Fe^t/Mg ionic ratios of amphiboles and their alteration product, biotite, in regional and type A "transitional" rocks. For symbols, see Fig. 4.1.

Table 4.4. Analyses of oxide minerals in regional and metamorphic rocks.

	1	2	3	4	5	6	7
	4430	BR01 179.95- 180.00m	BR01 185.10- 185.20m	Idem	Idem	BR01 188.75- 188.85m	Idem
SiO ₂	0.16						
TiO ₂	50.09	9.71	9.40	1.11	11.40	0.00	9.66
Al ₂ O ₃	0.00	0.10	0.05	0.98	0.35	0.06	0.13
Fe ⁺	46.93	90.10	91.03	86.45	97.99	99.75	87.42
Fe ₂ O ₃				96.0%		99.71	
MnO	2.30	0.02	0.09	0.02	0.07	0.06	0.00
MgO	0.01	0.0%	0.05	0.07	0.08	0.0%	0.02
CaO	0.23						
Total	99.72	100.02	99.66	98.22	99.91	99.97	96.29

1: ilmenite in regional rock
4,6: hematite in albitites

2,3,5,7: magnetite in albitites

*sum for hematite considers Fe₂O₃ as total iron

4.2. "Transitional" rocks

a) Plagioclase

Plagioclase in "transitional" rocks is oligoclase or albite-oligoclase with anorthite contents ranging from ~8.5 to 19.5%; in some samples plagioclase is more albitic but their calcium content is still significant (An ~4-8%). Zoned crystals are only sparsely found showing oligoclase rich cores (An ~15%) and albitic borders (An ~5%). Moreover, a number of grains of recrystallized plagioclase display untwinned borders or rims whose composition is albite (see Table 4.5); see Fig. 4.10.

b) Amphibole and biotite

As pointed out in section 3.3., both amphibole and biotite occur much the same way as in the regional gneisses, except in a few samples where amphibole is being replaced by garnet and/or pyroxene (Type B "transitional" rocks).

In Type A "transitional" rocks, amphibole is again hastingsite and Fe-pargasite (see 4.1.1. b, for discussion) with high Fe^{+2}/Mg atomic ratios ranging from ~44 to 88 (see Fig. 4.1 and Table 4.6).

Table 4.5- Selected analyses of plagioclases in "transitional rocks".

Sample no.	1	2	3	4	5	6	7	8	9
	AM2	AM2	BR01* 107.11- 107.14m	M039 52.63- 52.67m	Idem	BR01* 19.04- 19.10m	Idem	BR01 29.07- 29.15m	Idem
SiO ₂ (wt%)	67.64	66.71	66.64	65.54	67.95	64.43	68.10	64.02	67.29
Al ₂ O ₃	19.95	19.90	20.23	21.77	19.68	21.75	20.13	22.44	20.21
CaO	1.74	1.79	1.02	2.54	0.48	3.00	0.94	3.21	0.89
Na ₂ O	10.24	10.73	11.41	10.05	11.04	10.18	11.27	10.70	11.48
K ₂ O	0.30	0.07	0.12	0.23	0.16	0.20	0.12	0.05	0.29
Total	99.97	99.20	99.42	100.13	99.31	99.56	100.46	100.42	100.14

Ions on the basis of 32 oxygens

Si	11.958	11.796	11.763	11.512	11.946	11.421	11.863	11.293	11.793
Al	4.121	4.146	4.208	4.488	4.054	4.543	4.132	4.660	4.174
Ca	0.327	0.339	0.193	0.478	0.090	0.570	0.157	0.606	0.167
Na	3.480	3.679	3.905	3.423	3.763	3.499	3.806	3.656	3.902
K	0.067	0.016	0.027	0.052	0.036	0.045	0.027	0.011	0.063
Am%	9.44	9.41	4.68	12.10	2.32	13.85	3.93	14.18	4.05
Ab%	59.83	91.20	94.68	86.60	96.75	85.05	95.40	85.55	94.44
Or%	1.73	0.39	0.65	1.30	0.92	1.40	0.67	0.26	1.52

1: Oligoclase; 2: Albite-oligoclase porphyroclast; 3: Albite; 4: Zoned crystal, centre; 5: Zoned crystal, border; 6: Rimmed plagioclase, centre; 7: Rimmed plagioclase, untwinned rim; 8: Rimmed plagioclase, centre; 9: Rimmed plagioclase, untwinned rim

Type B

Table 4.6-Representative analyses of amphiboles from "transitional rocks".

Sample no.	1 BR01 208.92-209.00m	2 MD39 52.63-52.67m	3 RR01 18.0m-18.10m	4 + M03 RAA413	5 + M03 129.95-129.00m	6 + RAA36A
SiO ₂ (wt)	36.80	37.93	37.71	37.41	37.96	38.02
Al ₂ O ₃	11.63	11.78	12.02	10.63	11.22	10.96
FeO*	33.25	32.82	32.26	32.61	30.58	25.82
FeO*	27.52	27.48	29.87	27.79	27.57	21.59
FeO*	6.36	5.93	3.76	5.35	3.35	5.42
MgO	0.38	0.35	0.24	0.64	0.77	3.14
MnO	0.36	0.62	0.73	0.59	0.93	1.60
TiO ₂	0.57	0.59	0.76	0.46	0.49	0.31
CaO	9.99	10.71	10.30	10.69	10.95	10.61
Na ₂ O	1.63	1.05	1.70	1.37	1.29	1.45
K ₂ O	2.03	2.10	2.13	2.23	2.04	1.07
Total	97.28	98.50	98.23	97.17	96.56	94.87

Ions on the basis of 23 oxygens

Si	6.059	6.141	6.141	6.179	6.256	6.231
Al	1.940	1.859	1.859	1.821	1.744	1.769
Al	0.317	0.381	0.449	0.249	0.436	0.349
Fe ²⁺	3.791	3.721	3.932	3.839	3.799	2.959
Fe ³⁺	0.798	0.722	0.461	0.665	0.415	0.717
Mg	0.093	0.084	0.058	0.157	0.189	0.767
Mn	0.050	0.085	0.101	0.082	0.129	0.222
Ti	0.071	0.072	0.093	0.057	0.06	0.047
Exc. Oct.	0.110	0.066	0.094	0.050	0.031	0.060
Ca	1.763	1.854	1.797	1.492	1.934	1.970
Na+K	0.127	0.076	0.104	0.058	0.035	0.069
Na-A	0.394	0.254	0.429	0.341	0.377	0.455
K	0.426	0.434	0.442	0.469	0.429	0.224
Fe ²⁺ /Mg	49.63	53.62	77.15	29.11	22.57	5.08
Fe ²⁺ /Fe ²⁺ +Mg	0.99	0.99	0.99	0.97	0.96	0.94

1,2,3,4,6: Hastingsite; 5: Ferro-pargasite.

*Calculated according to Papike et al. (1976).

Fe²⁺: total iron.

*: amphibole in these samples is being replaced by garnet and/or pyroxene ("type A").

In Type B "transitional" rocks, amphibole is usually xenoblastic, showing very irregular borders. Their Fe^t/Mg atomic ratios are much lower and vary from 3 to 29 (with average ~ 10 , see Fig. 4.1). Even though Fe^{+3} calculated for amphiboles represents an indirect approximation using Papike et al's (1974) work (see Robinson et al, 1982), a plot of $\text{Fe}^{+3}/\text{Fe}^t$ against $\text{Fe}^t/(\text{Fe}^t + \text{Mg})$ was done to evaluate the chemical changes experienced by hastingsite due to the reaction amphibole $\rightarrow$ pyrox + garnet. It is clear from Figures 4.1 and 4.4 that not only Fe^t/Mg ratios of amphiboles drop to as low as 3, but also $\text{Fe}^{+3}/\text{Fe}^t$ tends to be higher than those obtained for amphiboles in regional and type A "transitional rocks". Similar results were obtained by Gilbert et al (1982), when studying reactions involving oxidation of hornblende.

Biotites in "transitional" rocks are also very rich in FeO and poor in MgO (iron-biotites, see 4.1.1. c for discussion), with Fe^t/Mg atomic ratios ~ 35 to 55. In a number of samples they could be classified as lepidomelanes (MO39 52.63-52.67m; BR01 29.07-29.15; BR01 18.04-18.10m). Again their chemistry seems to reflect that of the amphiboles from which they derive (see Fig. 4.3 and Table 4.7).

c) Garnet

Fig. 4.4. $\text{Fe}^{+3}/\text{Fe}^{+2}$ versus $\text{Fe}^{\text{t}}/\text{Fe}^{\text{t}}+\text{Mg}$ ionic ratios of amphiboles from regional and "transitional" rocks. Fe^{+3} is a rough estimate done according to Papike et al (1974); see text. For symbols, see Fig. 4.1.

Table 4.7- Selected analyses of biotite from "transitional rocks".

Sample no.	1 M039 52.63- 52.67m	2 BR01 29.07- 29.15m	3 BR01 18.04- 18.10m	4 AN9 HAZ903	5 AN6 HBU894
SiO ₂ (wt%)	33.05	32.46	32.33	33.86	36.66
TiO ₂	1.32	1.85	2.38	0.14	2.30
Al ₂ O ₃	15.81	14.67	16.06	17.04	14.55
Cr ₂ O ₃	0.09	0.03	0.02	0.00	0.01
FeO*	35.46	38.70	37.29	25.49	24.79
MnO	0.39	0.46	0.51	1.84	0.36
MgO	0.36	0.54	0.60	5.90	7.96
CaO	0.02	0.05	0.08	0.04	0.00
Na ₂ O	0.00	0.00	0.05	0.04	0.11
K ₂ O	9.89	7.85	6.66	3.38	10.27
Total	96.39	96.61	95.98	87.63	97.02
Ions on the basis of 22 oxygens					
Si	5.467	5.398	5.325	5.651	5.677
Al	2.533	2.602	2.674	2.349	2.323
Al	0.548	0.272	0.443	1.002	0.331
Ti	0.164	0.231	0.295	0.018	0.268
Mg	0.089	0.134	0.147	1.443	1.837
Fe	4.905	5.382	5.137	3.558	3.210
Mn	0.055	0.065	0.071	0.260	0.047
Cr	0.012	0.004	0.003	0.000	0.001
Na	0.000	0.000	0.016	0.013	0.033
Ca	0.004	0.009	0.014	0.07	0.000
K	2.087	1.665	1.399	0.720	2.028
Fet/Mg	55.88	40.69	35.35	2.65	1.77
Fet/Fet+Mg	0.98	0.98	0.97	0.73	0.64

*FeO; Fet: total iron.

1,5: Biotite as only mafic mineral.

4: Biotite starting to alter to chlorite (in type A).

2,3: Biotite from amphibole.

Microprobe analyses (see Table 4.8) show that in Type B "transitional" rocks, garnet is mostly composed of grossular and andradite end-members. Grossular ranges from 38.24 to 55.80% and andradite from 37.86 to 58.36%; spessartine ranges from 5.41 to 2.79%.

In only one sample studied (AN 3-RAA 368), does garnet show clear zoning (see Plate 2.B) with the centre being light yellowish-brown and becoming darker brownish towards the outer portions of the grains. Compositions are as follows:

centre: 52.31% almandine, 30.37% grossular, 16.41% spessartine

intermediate zone: 61.08% grossular, 19.14% almandine, 18.35% spessartine

border: 87.20% andradite, 10.23% grossular, 2.08% spessartine (see diagrams, Figs. 4.5 and 4.6).

Structural formulas were calculated taking into consideration that $M^{4+} = 3/2$, $M^{3+} = M^{2+}$ (Barriga, 1983).

End-member molecules were subsequently computed following the procedures of Rickwood (1968).

d) Pyroxene

Table 4.8-Representative analyses of garnets from type A "transitional rocks".

Anal. no.	1	2	3	4	Average
SiO ₂ (wt%)	95.92	96.12	96.06	96.08	96.02
TiO ₂	0.19	0.11	0.09	0.15	0.13
Al ₂ O ₃	9.05	9.31	9.08	9.71	9.29
Fe ₂ O ₃ ^c	25.03	26.39	25.25	23.98	26.64
Fe ₂ O ₃ ^a					16.38
FeO ^a					7.43
MnO	1.53	1.51	1.53	1.54	1.53
MgO	0.06	0.00	0.03	0.00	0.02
CaO	29.60	29.90	29.34	29.96	29.70
Total	100.28	100.34	100.41	100.31	100.33

Number of ions on the basis of 24 oxygens

Si	5.961
Ti	0.016
Al	1.617
Fe ³⁺	2.040
Fe ²⁺	1.028
Mn	0.214
Mg	0.004
Ca	5.266
Andrad %	55.79
Pyrope %	0.09
Spess %	3.91
Gross %	40.21
Almand %	

^aCalculated according to Garriqs (1943)

Fe₂O₃^c: total iron as Fe₂O₃

Sample M3-MV 791

Fig. 4.5. Chemical variation of garnets in type B "transitional" rocks, common and Ca-rich albitites, expressed as end-members andradite, grossular and spessartine (see text). For symbols, see Fig. 4.1.

Fig 4.6. Chemical variation of garnets in type B "transitional" rocks, common and Ca-rich albitites expressed as end-members andradite, almandine and grossular (see text). For symbols, see Fig. 4.1.

2

MICROCOPY RESOLUTION TEST CHART
NATIONAL BUREAU OF STANDARDS
STANDARD REFERENCE MATERIAL 1010a
(ANSI and ISO TEST CHART No. 2)

Pyroxenes in Type B "transitional" rocks are Ca-pyroxenes belonging to the hedenbergite-diopside series (Cameron and Papike, 1982), very rich in FeO which varies from 18.12 to 27.70%; MgO varies from 0.21 to 6.07% (see Table 4.9). They are hedenbergites and ferrosalites (see Figs. 4.7 and 4.8). They are rather poor in Na₂O, and Fe₂O₃ is sometimes absent; however, these tend to increase in borders of zoned grains from 0.39 to 0.80% Na₂O and from 0.31 to 3.50% Fe₂O₃ (only observed in sample BR01 128.95-129.00m; Fig. 4.9).

e) Chlorite

In very few samples, amphibole and/or biotite appear being partly altered to chlorite. Chlorite chemical analyses are found in Table 4.10.

f) Amphibole from pyroxene

Three type B "transitional" rocks show restricted pyroxene replacement by hornblende. These are hastingsite and Fe-pargasite, with Fe^t/Mg ratios from 1.95 to 8.89 (Table 4.11 and Fig. 4.1).

g) Carbonate

Table - 9 - Analyses of clinopyroxenes replacing hastingsite in type A "transitional rocks".

Sample no.	1 5807 129.95- 129.00%	2	3	4	5	6 8901 107.00- 107.70%	7	8	9
	129.00%	Idem	Idem	Idem	Idem	107.70%	Idem	Idem	Idem
SiO ₂ (wt)	49.98	49.78	49.84	49.07	47.25	51.71	50.01	50.47	49.51
TiO ₂	0.10	0.15	0.11	0.14	0.14	0.10	0.14	0.06	0.01
Al ₂ O ₃	0.44	1.54	1.25	0.57	1.56	1.35	1.96	1.33	1.59
FeO*	20.20	23.46	23.78	19.51	26.00	18.12	19.88	18.38	20.14
FeO ²⁺	19.92	23.46	23.78	17.46	20.95	18.70	19.78	18.37	16.99
FeO ³⁺	0.31	0.00	0.00	1.46	3.50	2.69	0.11	0.01	3.49
MnO	1.04	1.13	1.17	1.02	1.14	0.71	0.43	0.44	0.45
MgO	4.75	2.41	2.14	5.16	2.39	6.07	5.33	5.63	5.93
CaO	22.29	20.46	20.21	21.91	20.60	18.12	20.43	21.26	21.18
Na ₂ O	0.39	0.64	0.60	0.57	0.90	2.00	0.87	0.96	0.88
Cr ₂ O ₃	0.00	0.02	0.00	0.00	0.00	0.14	0.14	0.00	0.02
Total	99.55	99.65	99.90	97.97	98.08	98.32	99.19	99.67	99.73

Number of ions on the basis of 6 oxygens

Si	1.998	2.000	2.080	1.974	1.936	2.000	1.990	2.003	1.964
Al	0.012	0.000	0.000	0.024	0.064	0.000	0.020	0.000	0.056
Al	0.027	0.075	0.060	0.003	0.011	0.062	0.071	0.062	0.018
Fe ²⁺	0.663	0.798	0.908	0.601	0.714	0.514	0.655	0.610	0.554
Fe ³⁺	0.00	0.000	0.000	0.056	0.108	0.079	0.003	0.000	0.103
Mg	0.292	0.144	0.130	0.310	0.146	0.354	0.314	0.333	0.347
Mn	0.035	0.038	0.047	0.035	0.040	0.024	0.014	0.016	0.015
Ti	0.003	0.05	0.003	0.004	0.004	0.003	0.004	0.002	0.001
Cr	0.000	0.001	0.000	0.000	0.000	0.004	0.004	0.000	0.001
Exc. Oct.	0.019	0.051	0.044	0.010	0.023	0.040	0.067	0.023	0.042
Ca ²⁺	0.951	0.991	0.990	0.946	0.913	0.76	0.967	0.976	0.991
Na	0.030	0.051	0.047	0.045	0.064	0.152	0.067	0.066	0.067
En	14.97	7.94	7.13	14.44	9.23	21.75	17.13	19.03	19.32
Fs	34.99	43.47	44.46	32.39	40.24	31.57	35.67	33.01	31.07
Wo	50.15	49.57	49.41	50.92	51.49	46.68	47.20	49.95	49.61
Annite ⁴	-	-	-	4.40	-	15.14	-	-	12.23
Jadeite ⁴	1.74	4.00	4.12	-	-	-	9.30	4.42	-
Calcic ⁴	-	-	-	-	12.34	-	-	-	-
Name	Fe-salite	Magn-hastingsite	Magn-hastingsite	Fe-salite	Magn-hastingsite	Fe-salite	Fe-salite	Fe-salite	Fe-salite

1: centre of grain; 2,3: same, border; 4: centre of grain; 5: same, border

*Calculated according to Peacock et al. (1976).
FeO*: total iron as FeO.

Fig. 4.7. Chemical variation of clinopyroxenes in type B "transitional" and metasomatic rocks, expressed as end-members acmite, diopside and hedenbergite (see text). For symbols; see Fig. 4.1.

Fig. 4.8. Distribution of clinopyroxenes analyses in the system $\text{CaMgSi}_2\text{O}_6$ - $\text{CaFeSi}_2\text{O}_6$ - $\text{Mg}_2\text{Si}_2\text{O}_6$ - $\text{Fe}_2\text{Si}_2\text{O}_6$, for type B "transitional" rocks, mineralized and unmineralized Ca-rich albitites and type 1 and 2 epidiosites (see text). For symbols, see Fig. 4.1.

Table 1.10- Analyses of chlorites from "transitional rocks" and albitites.

Sample no.	1 160.00-160.10m	2 160.10-160.20m	3 160.20-160.30m	4 160.30-160.40m	5 160.40-160.50m	6 160.50-160.60m	7 160.60-160.70m	8 160.70-160.80m
SiO ₂ (wt)	51.00	50.31	50.18	50.05	50.29	50.40	50.61	50.13
TiO ₂	0.89	0.08	0.07	0.05	0.10	0.05	0.60	0.13
Al ₂ O ₃	16.57	19.16	18.57	17.07	16.15	16.65	16.35	16.55
Cr ₂ O ₃	0.11	0.01	0.03	0.06	0.25	0.76	0.05	0.06
FeO	23.33	30.06	30.09	35.36	35.26	21.20	19.77	22.05
MnO	2.07	1.89	1.64	0.37	0.61	0.25	0.13	0.31
MgO	11.57	6.36	6.66	6.97	6.53	18.07	16.61	16.08
CaO	3.17	0.00	0.00	0.00	0.00	0.05	0.05	0.03
Na ₂ O	0.08	0.02	0.00	0.00	0.00	0.00	0.03	0.00
K ₂ O	1.96	0.36	0.66	0.06	0.51	13.7	2.12	0.62
Total	97.25	98.21	98.52	98.00	96.90	98.26	97.12	96.06

Loss on the basis of 22 oxygens

Si	6.351	5.683	5.666	5.606	5.769	6.663	6.922	6.738
Al	1.667	2.517	2.356	2.196	2.231	1.357	1.078	1.262
Al	2.680	2.576	2.556	2.270	2.183	2.173	2.606	2.336
Ti	0.161	0.016	0.012	0.008	0.017	0.008	0.062	0.052
Cr	0.018	0.002	0.005	0.007	0.065	0.023	0.008	0.007
Fe	6.126	6.621	6.393	6.606	6.706	3.626	3.605	3.867
Mn	0.371	0.763	0.691	0.070	0.679	0.063	0.056	0.055
Mg	3.665	2.165	2.199	2.979	2.890	9.508	9.057	9.082
Ca	0.038	0.000	0.000	0.000	0.000	0.011	0.011	0.007
Na	0.033	0.009	0.000	0.000	0.000	0.000	0.012	0.000
K	0.396	0.096	0.137	0.017	0.146	0.157	0.157	0.166
Fe ²⁺ /mg	1.23	1.36	1.28	2.26	2.367	0.67	0.69	0.78
Fe ²⁺ /Fe ³⁺ mg	0.95	0.77	0.77	0.69	0.70	0.40	0.61	0.66

Fe²⁺/Fe³⁺ total iron.

1: Green chlorite replacing amphibole.

2: "Chloritized" biotite.

3, 4-6: Chlorite from amphibole.

6, 7: Chlorite from biotite.

1-3: "transitional rocks".

4-6: albitites.

Table 1. Analyses of amphiboles which alter from pyroxenes in type A "transitional rocks".

Sample no.	1 BR01 128.95- 129.00m	2 Idam	3 Idam	4 BR01 107.11- 107.14m	5 Idam
SiO ₂ (wt)	38.13	38.98	38.45	43.25	43.32
Al ₂ O ₃	11.08	10.54	11.71	10.16	10.39
FeO ^a	24.31	26.49	24.69	19.24	19.33
FeO ^c	29.60	29.16	28.77	22.82	21.57
Fe ₂ O ₃ ^a	5.88	2.97	4.33	3.98	2.49
MgO	2.18	1.90	2.04	6.75	6.20
MnO	0.97	0.92	0.98	0.64	0.29
TiO	0.30	0.37	0.21	0.14	0.09
CaO	10.98	10.74	10.88	9.75	9.28
Na ₂ O	1.39	1.42	1.18	1.69	1.33
K ₂ O	1.67	1.94	1.96	1.52	1.46
Total	96.89	96.29	96.45	97.12	96.18

Number of ions on the basis of 23 oxygens

Si	6.197	6.385	6.243	6.600	6.935
Al	1.803	1.615	1.757	1.320	1.165
Al	0.319	0.420	0.489	0.529	0.767
Fe ²⁺	3.304	3.629	3.354	2.485	2.550
Fe ³⁺	0.719	0.366	0.553	0.463	0.296
Mg	0.529	0.464	0.494	1.554	1.454
Mn	0.133	0.128	0.135	0.094	0.039
Ti	0.037	0.046	0.026	0.016	0.011
Exc. Oct.	0.041	0.052	0.050	0.131	0.121
Ca	1.912	1.945	1.993	1.613	1.569
Na+K	0.047	0.059	0.057	0.255	0.310
Na+K	0.391	0.392	0.314	0.251	0.097
K	0.346	0.405	0.406	0.299	0.294
Fe/Mg	7.87	8.99	8.14	1.95	1.94
Fe ²⁺ /Fe ²⁺ +Mg	0.99	0.90	0.99	0.64	0.64

1,3: Hastingsite

2,4,5: Fe-pargasite

^aCalculated according to Papike et al. (1974).
Fe²⁺: total iron.

When present in "transitional" rocks, carbonate is associated with mafic minerals. It has been analysed in one sample (see Table 4.12); it is calcite with ~1.5% FeO.

4.3. Metasomatic rocks

a) Plagioclase

In all types of albitites studied, plagioclase is either pure albite or albite with An content from ~3-7% (see Table 4.13). In both types of epidiosites (only three samples were studied), plagioclase is oligoclase with An content ~18% (see Table 4.13 and Fig. 4.10).

b) Pyroxene

Representative analyses of pyroxenes are given in Table 4.14. Pyroxenes belong to the diopside-hedenbergite series (Ca-pyroxenes), with moderate to high acmitic contents varying from ~8-35%, with one sample (AN 3 RAA 331) reaching as much as 80%. For this reason, they should be considered under the Calcium-Sodium Pyroxene Group (Deer et al, 1978), and most of them can be named aegirine-augites. Figs. 4.7 and 4.9 display the distribution of pyroxenes in terms of acmite, hedenbergite and

Table 4.12-Analyses of carbonates from type A "Transitional rocks", albitites and Ca-rich albitites.

Sample no.	1 BR01 29.07- 29.15m	2 Idem	3 BR01 188.00- 188.13m	4 Idem	5 AM17	6 Idem	7 AM7 1145-2	8 Idem
MgO	0.00	0.00	0.00	0.00	0.16	0.00	0.00	0.00
CaO	54.28	54.04	55.89	55.32	55.00	54.90	56.00	56.20
FeO	1.44	1.58	0.58	0.30	0.63	0.70	0.00	0.00
MnO	0.54	0.54	0.02	0.00	0.05	0.02	0.00	0.00
Total	56.26	56.16	56.48	55.62	55.84	55.62	56.00	56.20
MgCO ₃					0.34			
CaCO ₃	96.89	96.46	99.75	98.75	98.18	98.00	99.96	100.32
FeCO ₃	2.32	2.55	0.94	0.48	1.02	1.13		
MnCO ₃	0.87	0.87				0.03		
	100.09	99.88	100.71	99.23	99.61	99.16	99.96	100.32

1,2: carbonate associated with amphibole in type A "transitional rock".

3-6: mineralized garnet-pyrox-albitites; carbonate is associated with garnet

7,8: carbonate in mineralized garnet-pyrox-albitite (Ca-rich).

Fig. 4.9. Chemical variation of clinopyroxenes (only some analyses), expressed as end-members acmite, diopside and hedenbergite, for metasomatic rocks, except dashed line which is from a type B "transitional" rock. Arrows indicate trends from core-rim zoned grains (see text). For symbols, see Fig. 4.1.

Fig. 4.10. Chemical variation of plagioclase expressed as end-members albite and anorthite, including analyses from regional, "transitional" and metasomatic rocks. Each point represents the average of various analysed grains. For symbols, see Fig. 4.1.

Fig. 4.11. Distribution of clinopyroxenes analyses in the system $\text{CaMgSi}_2\text{O}_6$ - $\text{CaFeSi}_2\text{O}_6$ - $\text{Mg}_2\text{Si}_2\text{O}_6$ - $\text{Fe}_2\text{Si}_2\text{O}_6$, for common unmineralized and mineralized albitites (see text). For symbols, see Fig. 4.1.

diopside end-members. As pointed out by Deer et al (1978), the name aegirine-augite is used for those members of $\text{Ca}(\text{Mg}, \text{Fe}^{+2})\text{Si}_2\text{O}_6$ - $\text{NaFe}^{+3}\text{Si}_2\text{O}_6$ (Diopside-Hedenbergite-Aegirine) series containing between $\text{Fe}^{+3} = 0.2$ and $\text{Fe}^{+3} = 0.8$ (ionic); aegirines contain $\text{Fe}^{+3} = 0.8$ and higher (i.e. sample AN 3, RAA 331 would be the only one). For all other pyroxenes with less than $\text{Fe}^{+3} = 0.2$, their names should be derived from the diopside (Di) - hedenbergite (Hd) - clinoferrosilite (FS) - clinoenstatite (En) quadrilateral (Figs. 4.8 and 4.11).

On the quadrilateral, almost all analyses plot above the join CaMg-CaFe where Ca+Mg+Fe are normalized to 100, because of the high acmite content of most analysed grains. This does not indicate abnormal Ca-rich pyroxenes; aegirines from alkaline rocks plot in a similar fashion (Duda and Schmincke, 1978). It becomes clear from these diagrams that other than aegirine-augites, the majority of pyroxenes in metasomatic rocks fall in the fields of ferrosalite, salite and diopside. In particular, pyroxenes from Ca-rich albitites and epidiosites fall in the salite and diopside fields.

A rather striking feature is the chemical zoning of pyroxenes (see Plate 2.E). Even though this is not always clear in thin section, microprobe analyses have shown that zoning exists in many grains, in all types of metasomatic

Fig. 4.12. Chemical variation of clinopyroxenes as a function of Fe^{+3} versus Fe^t/Mg . Arrows indicate trends from core-rim zoned grains. All grains are from metasomatic rocks, except those linked by a dashed line which are from a type B "transitional" rock. For symbols, see Fig. 4.1.

rocks. Light green cores grade into dark green rims reflecting the replacement $\text{NaFe}^{+3} \rightleftharpoons \text{Ca}(\text{Mg}, \text{Fe}^{+2})$, so common in the system diopside-hedenbergite-acmite (Nolan, 1969). From Fig. 4.9 it is obvious that there is a general increase in ferric iron towards the rims. Fig. 4.12 is a plot of Fe^{+3} versus $\text{Fe}^{\text{t}} + \text{Mg}$ from analyses done on cores and rims, showing the evident Fe^{+3} enrichment.

c) Garnet

Table 4.15 displays selected analyses of garnets in albitites. Garnet all types of albitites occur replacing pyroxene and their composition is essentially andradite with varying amounts of the grossular end-member (see Figs. 4.5 and 4.6). With the exception of the Ca-rich albitites, where the grossular end-member varies from 26 to 37%, in all other albitites grossular end-member never exceeds 12%; in sample AN 3 RAA 331 garnet is almost pure andradite with grossular 1%. Both Mn and Mg are fairly low and spessartine end-member does not exceed 2.5%, (Fig. 4.5) whereas pyrope is always less than 0.5%.

Andradite garnet is a principal mineral in many contact metasomatic skarns (Taylor and Libou, 1978) where it occurs replacing hedenbergitic pyroxene (Sato, 1980).

Table 4.15-Representative analyses of garnets from albitites

Sample no.	1	2	3	4	5	6	7	8	9	10
	BRD1 106.70- 106.80a			BRD1 105.60- 105.73a				R43 RAA 387		
		Idem	Average		Idem	Idem	Average		Idem	Idem
SiO ₂ (wt)	35.34	35.27	35.20	35.34	35.45	35.38	35.26	36.43	35.88	35.90
TiO ₂	0.36	0.38	0.47	0.53	0.52	0.52	0.52	0.97	0.29	0.35
Al ₂ O ₃	2.46	2.52	2.89	2.30	2.67	2.77	2.33	5.48	2.81	0.81
Fe ₂ O ₃ ¹	28.88	29.59	29.21	30.69	29.03	29.98	29.91	25.06	28.00	31.36
Fe ₂ O ₃ ²			27.59				26.62	22.63	26.76	29.01
FeO ³			1.46				3.34	2.37	1.12	2.11
MnO	0.52	0.63	0.75	0.62	0.63	0.55	0.59	0.28	0.88	1.08
MgO	0.00	0.00	0.00	0.00	0.05	0.02	0.03	0.36	0.0	0.30
CaO	33.26	31.91	32.58	30.53	31.69	31.30	31.17	32.26	32.36	30.98
Total	100.89	100.57	100.70	99.61	99.99	99.81	99.81	100.63	99.98	100.50

Number of ions on the basis of 24 oxygens

Si	5.982	5.967	5.976	6.002	6.084
Ti	0.059	0.066	0.119	0.036	0.084
Al	0.687	0.665	1.099	0.958	0.361
Fe ³	3.665	3.365	2.769	3.368	3.679
Fe ²	0.202	0.888	0.325	0.157	0.298
Mn	0.105	0.088	0.031	0.119	0.388
Mg		0.007	0.039		0.025
Ca	5.793	5.652	5.669	5.768	5.545
Andrad %	87.09	87.86	71.32	85.15	95.81
Pyrope %		0.13	0.68		0.68
Sporss %	1.86	1.67	0.54	2.13	2.54
Gross %	11.08	10.33	26.41	12.72	1.18
Alm. %					

*Calculated according to Barrois (1983)

Fe₂O₃¹: total iron as Fe₂O₃

1: mineralized garnet-albitite

8: garnet-albitite

8: Ca-rich albitite

9,10: mineralized pyroxene-garnet-albitite

d) Amphiboles

Analyses of amphiboles can be found in Table 4.16. In albitites, hornblende occurs as an alteration product after pyroxene along borders and cleavage lines. Like amphiboles described in earlier sections, these are also calcic amphiboles and the best end-member names are: . hastingsite (most cases), Mg-hastingsite and rare Fe-pargasite, with $\text{Fe}^{\text{t}}/\text{Mg}$ ratios from 0.61 to 3.29. The chemistry of amphibole expressed as a function of $\text{Fe}^{\text{t}}/\text{Mg}$ ratios, suggests that it reflects that of the pyroxenes from which they alter (Fig. 4.13).

e) Biotite

In albitites, biotites which are rare, have been found to occur in various ways. It may occur as the only mafic mineral, thus obscuring its relation to the alteration process (Table 4.17, analyses 1-3 and 6-8). In one other case (analyses 4-5), it appears with epidote and it is altering to chlorite. Sample AN13-HAV 777 (analyses 11-13) is a mineralized albitite and in this, biotite is replacing green amphibole. Very local alteration of magnetite to biotite is observed in one sample, and in this case TiO_2 in biotite is substantial (up to 4.5% TiO_2).

Table 1. Selected analyses of amphiboles which appear as secondary product after pyroxene in metasomatic rocks.

	1	2	3	4	5	6	7	8	9	10	11
	BR01 185.27- 185.37m	idem	idem	BR01 185.62- 185.70m	BR01 186.00- 186.10m	idem	BR01 185.10- 185.20m	idem	AM7 LL45-12	AM3 RAA336	idem
SiO ₂ (wt%)	43.50	44.54	43.40	39.96	41.89	45.79	42.82	39.92	40.95	43.61	44.78
Al ₂ O ₃	7.57	7.975	8.05	10.92	8.41	6.71	8.96	11.38	11.32	9.46	7.41
FeO ^t	21.34	20.57	19.95	26.74	26.57	22.23	24.63	25.98	21.02	17.03	14.71
FeO ^o	17.94	17.02	17.66	19.80	22.44	19.79	20.23	22.38	18.23	13.02	12.36
Fe ₂ O ₃ [*]	3.78	3.94	2.54	7.7	4.15	2.72	4.99	4.00	3.04	4.46	2.61
MgO	9.49	10.32	9.77	5.79	6.37	8.92	6.77	4.84	8.57	12.25	13.50
MnO	0.38	0.40	0.27	0.42	0.48	0.31	0.48	0.43	0.31	0.24	0.18
TiO	0.38	0.41	0.41	0.09	0.30	0.29	0.21	0.12	0.20	0.07	0.01
CaO	10.13	10.11	10.43	11.02	10.77	11.03	11.40	11.07	10.73	11.64	11.61
Na ₂ O	2.50	2.77	2.54	1.61	3.12	1.66	1.33	1.63	2.46	2.25	1.63
K ₂ O	1.61	1.48	1.70	1.84	0.71	1.02	1.38	2.05	2.32	1.55	1.14
Total	97.28	98.74	96.81	99.18	99.04	95.13	98.47	97.82	98.18	98.55	97.27

Number of ions on the basis of 23 oxygens

Si	6.715	6.728	6.706	6.203	6.519	6.975	6.604	6.301	6.307	6.509	6.941
Al	1.285	1.272	1.294	1.797	1.481	1.024	1.395	1.698	1.693	1.490	1.059
Al	0.092	0.108	0.173	0.201	0.061	0.181	0.234	0.419	0.362	0.175	0.237
Fe ²⁺	2.316	2.150	2.252	2.571	2.972	2.521	2.609	2.955	2.354	1.625	1.533
Fe ³⁺	0.439	0.448	0.296	0.900	0.466	0.311	0.568	0.475	0.353	0.501	0.292
Mg	2.183	2.323	2.249	1.339	1.677	2.002	1.556	1.139	1.967	2.725	2.945
Mn	0.049	0.051	0.035	0.055	0.063	0.040	0.063	0.057	0.040	0.030	0.023
Ti	0.044	0.047	0.048	0.010	0.035	0.033	0.024	0.014	0.024	0.009	0.001
Exc. Oct.	0.124	0.127	0.084	0.078	0.095	0.088	0.054	0.059	0.100	0.064	0.072
Ca	1.675	1.636	1.727	1.833	1.796	1.800	1.984	1.872	1.771	1.862	1.844
Na-Mn	0.200	0.237	0.189	0.089	0.109	0.111	0.062	0.064	0.129	0.074	0.082
Na-A	0.548	0.575	0.584	0.395	0.832	0.379	0.336	0.431	0.405	0.577	0.346
K	0.317	0.285	0.335	0.368	0.141	0.194	0.272	0.413	0.456	0.295	0.223
Fe ^t /Mg	1.29	1.14	1.16	2.63	2.38	1.43	2.08	3.06	1.39	0.79	0.62
Fe ^t /Fe ^t _{Mg}	0.56	0.53	0.54	0.72	0.70	0.59	0.67	0.75	0.58	0.44	0.34

1,3-8: Hastingsite

2,10,11: Mg-hastingsite

9: Fe-pargasite

1,4,5: garnet-pyrox-albite

7: mineralized garnet-pyrox-albite

9: mineralized garnet-pyrox-

10: mineralized epidote

albite (Ca-rich)

*Calculated according to Papine et al. (1974).

Fe^t: total iron.

Table 4.17-Analyses of biotites in albitites.

Sample no.	1 MOJ 43.97- 44.03m	2 1dm	3 1dm	4 4N3 HAV327	5 1dm	6 4N13 HOU269	7 1dm	8 1dm	9 4N6 HAV695	10 1dm	11 4N13 HAV777	12 1dm	13 1dm
SiO ₂ (wt)	35.66	35.12	35.67	35.25	35.96	36.27	36.80	35.80	35.09	36.68	39.06	39.26	38.68
TiO ₂	3.22	2.26	3.25	1.89	1.79	0.71	0.63	0.69	4.59	3.88	1.23	1.37	1.24
Al ₂ O ₃	14.02	13.38	14.18	13.42	13.14	13.97	14.38	14.08	12.48	12.77	13.86	14.48	14.05
Cr ₂ O ₃	0.00	0.00	0.00	0.03	0.13	0.00	0.01	0.00	0.07	0.00	0.00	0.00	0.00
FeO	30.81	29.66	29.80	28.67	27.51	18.51	18.13	19.09	28.24	29.20	11.68	11.09	11.86
MnO	0.14	0.22	0.27	0.42	0.40	0.15	0.12	0.15	0.93	0.81	0.22	0.24	0.25
MgO	3.52	4.30	3.76	6.52	6.28	13.70	16.02	14.88	3.10	2.75	17.80	18.08	16.78
CaO	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
Na ₂ O	0.10	0.27	0.03	0.00	0.00	0.24	0.14	0.20	0.00	0.11	0.07	0.06	0.11
K ₂ O	9.28	9.86	9.05	9.74	9.82	8.95	6.08	6.24	5.65	9.09	10.36	10.02	10.14
Total	96.75	95.07	96.01	95.76	95.01	96.50	90.31	91.13	93.19	93.29	96.26	96.60	93.11

Total on the basis of 22 oxygens

Si	5.681	5.713	5.692	5.652	5.776	5.835	5.502	5.627	5.769	5.741	5.819	5.796	5.837
Al	2.319	2.287	2.308	2.348	2.226	2.145	2.498	2.373	2.231	2.259	2.181	2.206	2.163
Al	0.312	0.277	0.359	0.188	0.262	0.346	0.181	0.235	0.187	0.233	0.253	0.313	0.335
Ti	0.386	0.276	0.390	0.228	0.216	0.081	0.075	0.082	0.567	0.483	0.138	0.152	0.142
Mg	0.836	1.043	0.894	1.558	1.504	3.114	3.775	3.477	0.760	0.679	3.954	3.977	3.774
Fe	4.105	4.035	3.977	3.818	3.688	2.348	2.387	2.509	3.883	4.043	1.456	1.369	1.497
Mn	0.019	0.030	0.036	0.057	0.054	0.019	0.016	0.020	0.130	0.114	0.028	0.030	0.032
Cr	0.000	0.000	0.000	0.004	0.017	0.000	0.001	0.000	0.009	0.000	0.000	0.000	0.000
Na	0.031	0.085	0.009	0.000	0.000	0.071	0.043	0.061	0.013	0.035	0.020	0.017	0.032
Ca	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.007	0.000	0.000	0.000	0.000	0.000
K	1.886	2.046	1.842	1.982	2.012	1.741	1.226	1.251	1.814	1.919	1.970	1.984	1.852
Fe ²⁺ /Mg	4.934	3.90	4.49	2.49	2.49	0.76	0.64	0.73	5.28	6.12	0.37	0.35	0.40
Fe ²⁺ /Fe ²⁺ +Mg	0.83	0.80	0.82	0.71	0.71	0.43	0.39	0.42	0.44	0.86	0.27	0.26	0.29

*FeO calculated as total iron (Fe²⁺)

1-3: quartz-biotite-albitite (only mafic mineral)

4-5: chlorite-biotite-epidote-albite-quartz

6-8: biotite-albite (only mafic)

9-10: biotite from magnetite in mineralized magnetite-albitite

11-13: biotite from granoblasts in mineralized biotite-magnetite-albitite

Fig 4.13. Fe^t/Mg ionic ratios of pyroxenes and their alteration product, amphibole, in metamorphic rocks. For symbols, see Fig. 4.1.

f) Chlorite

Analyses of chlorite replacing amphibole and in one other instance, biotite, are found in Table 4.10. Such alteration is very rare.

g) Magnetite and hematite

Analyses of both oxides are given in Table 4.4. Magnetite can contain up to 9% TiO_2 and hematite up to 1%.

h) Carbonate

Carbonate from andradite, observed in a few samples, and carbonate in Ca-rich albitites have been analysed by microprobe. They are essentially pure calcite (Table 4.12).

i) Epidotes

Epidotes in epidosites replace pyroxene, usually constituting perfect pseudomorphism. They are common epidotes (see Table 4.18), and their Ps varies from 26 to 28 (Ps is the mole fraction of component $\text{Ca}_2\text{Fe}_3\text{Si}_3\text{O}_{12}$, pistacite, in the clinozoizite-epidote series, calculated from the ratio $\text{Fe}^{+3}/\text{Fe}^{+3}+\text{Al}$; see Holdaway, 1972). Epidotes in epidosites contain a maximum of 1.2% UO_2 . Small accessory epidote crystals have also been analysed.

Table 6.18 Selected analyses of epidotes in metamorphic rocks.

Sample	1 495 11.64	2 1.00	3 1.00	4 1.00	5 1047-705	6 1.00	7 1.00	8 1.00	9 8047 75.12 75.10	10 1.00	11 80405	12 1.00
$\text{SiO}_2(\text{wt})$	57.32	58.27	58.36	57.68	57.11	57.00	57.29	57.00	57.07	57.45	51.96	57.17
TiO_2	0.06	0.06	0.06	0.06	0.06	0.06	0.06	0.00	0.00	0.00	0.27	0.00
Al_2O_3	26.46	25.49	26.85	26.50	26.09	26.06	25.66	25.50	22.11	22.34	16.52	21.55
Fe_2O_3^a	11.25	11.02	11.23	10.35	10.55	10.75	10.39	10.38	16.05	16.30	16.75	11.75
FeO^b	0.02	0.01	0.00	0.01	0.00	0.00	0.05	0.06	0.00	0.06	0.91	0.36
MgO	0.00	0.00	0.02	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.11	0.00
CaO	21.66	21.00	21.09	21.11	21.49	21.39	21.27	21.20	21.36	21.11	12.51	21.39
Na_2O	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.10	0.00	0.00	0.99	0.91
K_2O	0.08	0.06	0.09	0.06	0.02	0.06	0.01	0.02	0.21	0.12	7.89	0.00
MnO	0.05	0.07	0.10	0.10	0.06	0.06	0.26	1.20	0.07	0.01	0.07	0.07
Li_2O	0.00	0.06	0.03	0.00	0.00	0.00	0.02	0.05	0.00	0.07	1.03	0.00
Cr_2O_3	0.02	0.06	0.02	0.06	0.06	0.00	0.06	0.01	0.00	0.07	6.16	0.01
Total	99.81	100.08	99.81	100.16	99.90	99.95	99.60	99.60	99.01	99.15	92.01	99.05

Number of ions on O10 basis of 25 oxygens

Si	5.896	5.907	5.999				5.998
Ti	0.007	0.006	0.007				0.006
Al	6.551	6.436	6.575				6.389
Fe^{+3}	1.575	1.512	1.556				1.695
Fe	0.001	0.000	0.000				0.007
Mg	0.000	0.000	0.000				0.000
Ca	4.091	3.989	3.965				3.901
Na	0.000	0.000	0.000				0.000

1-g: epidote (plagioclase) reacting pyroxene in type 1 epidote

2-h: epidote (plagioclase) reacting pyroxene in type 2 epidote

11: accessory epidote showing zoning in albitic centre of grain (light colour)

12: same as 11, light green border

^atotal iron as Fe_2O_3 ^bon (plagioclase) not determined; value reported as total oxide

Table 4.19- Analyses of "other minerals" in albitites.

Sample no.	AN7 LL45-14	Idem	AN7 LL45-1	Idem	AN13 MFC956	AN13 RAA001	Idem
mineral	Vesuvianite	Idem	wollastonite	pectolite from wollastonite	prehnite from pyroxene	uraninite	Idem
SiO ₂ (wt.%)	35.94	35.11	51.08	52.96	44.89	16.57	10.38
TiO ₂	2.67	2.59	0.03	0.00		0.49	0.39
Al ₂ O ₃	12.07	11.83	0.01	0.40	23.50	2.88	1.27
FeO ^c	6.11	6.08	0.00	0.00		8.06	6.89
MnO	0.14	0.01	0.18	0.05		0.45	0.59
MgO	1.70	1.40	0.04	0.01		0.31	0.23
CaO	32.04	32.60	48.36	33.43	27.22	11.71	10.05
Mg ₂ O	0.43	0.55	0.04	8.91	0.18	0.28	0.37
K ₂ O	0.02	0.05	0.02	0.04	0.02		
Cr ₂ O ₃	0.07	0.00	0.00	0.06			
Y ₂ O ₃						0.28	0.44
UO ₂ ^a						55.25	66.21
La ₂ O ₃						0.60	0.79
Ce ₂ O ₃						0.82	1.09
	91.17	90.22	99.76	95.87	95.81	97.71	98.71

^aoxidation state not determined; values reported as total oxide

j) Other minerals

Table 4.19 shows analyses of the following:

- wollastonite;
- pectolite, which replaces wollastonite along borders;

- prehnite, which replaces pyroxene;
- vesuvianite: vesuvianite has been analysed in only one sample. According to Deer et al (1982), it corresponds to the variety wilwite because it is almost isotropic, sometimes isotropic, and optically positive. The increase in isotropism is related to increase in (OH); this probably explains why the totals of the analyses are so low. In addition, the metamict (?) character of the grains also suggest they could contain radioactive impurities (see Heinrich, 1958);

- uraninite: these results probably do not reflect the overall composition because elements like Th, Pb and Zr were not analysed;

- K-feldspar: analyses of K-feldspar in regional, "transitional" and metasomatic rocks are found in Table 4.20.

Table 4. 20 Analyses of microcline from regional, transitional and metamorphic rocks.

Sample no.	1	2	3	4	5	6	7	8	9	10	11	12
	AP00	AP01		AP02	AP03	AP04	100000	AP05	AP06		AP07	
	100.00	100.00	100.00	100.00	100.00	100.00	100.00	100.00	100.00	100.00	100.00	100.00
SiO ₂ (wt)	65.29	66.69	66.70	65.60	65.50	65.55	65.11	66.87	65.97	66.70	66.60	66.11
Al ₂ O ₃	12.93	14.30	19.07	16.45	17.96	18.02	18.33	18.59	17.82	17.82	17.68	17.73
CaO	0.01	0.00	0.00	0.01	0.02	0.01	0.02	0.06	0.02	0.02	0.02	0.00
MgO	0.22	0.46	0.67	0.65	0.79	0.63	0.67	0.69	0.54	0.62	0.51	0.36
K ₂ O	15.59	15.86	16.00	15.06	15.93	16.09	16.00	16.16	15.79	15.82	16.22	15.98
Total	99.06	99.30	100.52	99.65	99.52	99.30	100.17	100.11	99.68	99.54	99.02	100.20
Si	12.113	11.992	11.976	12.051	12.105	12.020	12.077	11.957	12.187	12.127	12.066	12.151
Al	3.920	4.026	4.179	4.002	3.900	3.926	3.982	4.038	3.867	3.875	3.891	3.839
Ca	0.004	0.000	0.000	0.006	0.006	0.002	0.004	0.009	0.004	0.006	0.002	0.000
Mg	0.079	0.231	0.239	0.232	0.106	0.227	0.269	0.279	0.249	0.222	0.195	0.128
K	3.689	3.754	3.750	3.531	3.761	3.621	3.762	3.799	3.591	3.630	3.668	3.785
Na	0.16	0.00	0.00	0.16	0.10	0.05	0.10	0.20	0.10	0.10	0.05	0.00
Or	2.10	3.38	5.98	6.15	2.69	5.61	6.15	5.97	5.85	5.75	6.56	5.31
Ors	97.75	98.22	98.02	93.69	97.21	98.36	93.76	91.89	98.65	98.16	95.39	96.69

1: Regional gneiss; 2, 3: Type A transitional rocks; 4: Type B transitional rocks; 5, 6, 7: Microcline-albitites; 8, 9, 10, 11, 12: Pyroxene-albite (almandine).

CHAPTER 5

LITHOGEOCHEMISTRY

A series of diagrams were made to illustrate and compare some of the chemical results obtained for regional, "transitional" and metasomatic rocks. They will be discussed at length in section 5.3.1. c. The general chemical characteristics of regional and "transitional" rocks are discussed in the next two sections.

5.1. Regional gneisses

5.1.1. Introduction

Chemical analyses of regional gneisses are presented in Table 5.1 . This combines sampling done to the NNE and SSE of the U anomalous area (see Location Map, Fig.

5.1), corresponding to the Caetité-Maniaçú and Caetité-Brumado traverses.

Table 5.1- Major and selected trace element abundances in regional gneisses.

Sample no.	1 #45	2 #46	3 #47	4 #48	5 #49	6 #32	7 #33	8 #34	9 #35	10 #37	11 #38
SiO ₂ (wt)	72.52	69.49	73.41	67.90	73.90	76.96	72.70	76.13	73.10	73.68	73.40
TiO ₂	0.79	0.25	0.17	0.30	0.01	0.27	0.38	0.33	0.50	0.30	0.24
Al ₂ O ₃	12.99	14.93	12.78	14.78	13.50	11.64	12.00	11.40	11.43	11.97	11.90
Fe ₂ O ₃	2.95	0.39	2.91	0.92	0.08	0.44	4.25	0.45	4.95	0.46	0.38
FeO		2.67		3.07	0.57	2.32		2.46		2.07	1.47
MnO	0.03	0.04	0.04	0.06	0.02	0.04	0.05	0.04	0.06	0.04	0.03
MgO	0.05	0.04	ND	ND	0.22	0.05	0.14	0.04	0.08	0.07	0.25
CaO	1.93	1.93	0.64	2.02	0.77	0.69	0.94	0.63	0.44	0.95	1.11
Mg ₂ O	3.23	3.34	2.42	3.31	3.41	3.14	3.08	3.58	3.12	3.59	4.99
K ₂ O	4.97	5.70	6.42	5.99	5.49	4.96	4.63	4.54	4.60	4.45	4.76
*P ₂ O ₅	ND	0.50	ND	0.02	0.03	0.31	0.06	ND	0.03	ND	0.05
LOI	0.20	0.50	0.40	0.30	0.31	0.30	0.39	0.40	0.20	0.60	0.50
Total	98.96	99.81	99.11	94.95	94.37	99.40	94.66	99.07	94.91	99.21	99.69
Nb(ppm)	39	35	38	29	-	19	34	20	31	16	11
Zr	550	397	349	453	26	367	490	356	694	371	293
Y	37	45	34	25	4	43	45	162	37	35	39
Sr	149	181	107	229	38	67	54	56	99	63	77
Rb	125	155	174	148	115	201	167	141	127	177	154
Pb	21	33	30	29	35	29	28	19	26	24	23
Zn	49	87	99	86	25	78	88	49	47	75	62
Cu	0	9	9	10	9	9	10	9	11	10	9
Ni	3	2	4	3	3	6	6	7	7	5	5
Co	14	32	52	17	36	59	32	40	43	32	44
Cr	5	6	6	5	3	7	7	5	6	4	4
Ba	1039	1758	1152	2253	224	605	732	445	795	745	691
V	2	7	6	3	3	5	6	4	-	7	8
S	70	62	54	49	48	50	52	54	57	54	43
Ca	19	26	24	27	21	20	21	15	14	18	16
Sc											
Ta											
Hf											
In											
U	3.52	2.96	4.87	3.01	3.48	3.89	4.13	2.84	3.50	7.21	3.43
residue		0.11		0.21	0.11	0.14		0.24		0.27	0.15

Table 1
Cont.

Sample no.	12	13	14	15	16	17	18	19	20	21
	4479	4480	4482	4483	4484	4487	4487	4481	4482	4480
SiO ₂ (wt)	73.39	73.67	73.90	73.93	74.39	64.60	67.94	68.90	69.70	55.92
TiO ₂	0.40	0.33	0.47	0.32	0.27	0.40	0.64	1.61	1.09	0.68
Al ₂ O ₃	1.75	12.75	12.30	12.02	1.99	14.00	13.67	12.40	15.40	16.40
Fe ₂ O ₃	1.04	0.41	0.53	0.64	0.43	1.52	1.64	16.90	10.40	0.94
FeO	2.69	2.71	3.73	2.89	2.50	3.62	4.32			5.91
MnO	0.05	0.05	0.06	0.04	0.04	0.07	0.11	0.24	0.20	0.11
MgO	0.06	ND	0.29	0.12	0.04	0.14	ND	5.03	3.54	5.30
CaO	0.97	0.52	1.08	0.41	0.67	2.04	2.45	9.04	14.10	7.23
Na ₂ O	3.12	3.31	3.35	3.34	3.35	3.45	3.25	2.57	1.51	4.44
K ₂ O	4.45	5.25	4.21	4.94	4.49	5.04	4.36	0.46	0.67	0.99
P ₂ O ₅	0.01	ND	0.04	0.01	ND	0.11	0.15	0.15	0.12	0.13
LOI	2.30	0.40	0.19	0.30	0.99	0.46	0.09	0.70	1.16	0.10
Total	94.44	99.40	94.40	99.07	99.44	94.95	99.53	94.34	94.69	94.42
Nb (ppm)	19	34	22	16		32		12	13	-
Zr	465	342	556	376		347		113	146	117
Y	193	64	42	25		53		19	44	3
Sr	73	54	46	64		244		165	232	416
Rb	144	200	125	139		117		-	-	4
Pb	26	26	20	24		24		12	19	15
Zn	44	41	41	75		74		122	116	73
Cu	10	10	10	9		10		19	19	12
Mn	-	5	6	5		5		22	52	33
Co	32	13	35	43		33		24	36	39
Cr	5	5	7	5		5		24	90	60
Ba	736	602	753	727		1741		247	138	496
V	-	12	3	3		14		275	143	43
S	64	47	43	44		52		75	46	74
Ca	17	19	17	19		20		14	21	19
Sc										
Ta										
Hf										
Th										
U	2.72	3.55	2.70	3.64		1.52	4.2	1.02	24.90	0.12
Fe/ΣFe	0.26	0.21	0.11	0.19	0.14	0.13	0.22			0.12

*Total Iron

1-18: amphibole-K-feldspar-gneisses

19-20: amphibolites

21: hypersthene-augite-gneiss (granulite)

L.O.I.: (loss of ignition) is mainly H₂O as indicated by petrography
ND: not detected

The regional gneisses hosting the Lagoa Real rocks, are part of the "Paramirim Complex" (Jardim de Sá, 1978) forming the high-grade Archean and Lower Proterozoic terrain east of the Espinhaço range (see Chapter 2). These are amphibolite facies orthogneisses of granodioritic, tonalitic and granitic compositions, with a few localized inclusions of granulite facies rocks. Adjacent to the project area, the rocks are coarse grained, porphyroclastic (augen) gneisses, relatively homogeneous, containing quartz, perthitic k-feldspar, oligoclase or albite, amphibole (hastingsite) and biotite. Samples of typical amphibolites and one sample of granulite have also been analysed.

General characteristics of the Paramirim Complex rocks, seem to indicate that they are rather similar to other basement terrains around the world. They are dominated by tonalitic grey gneisses (the augen gneisses), with equivalent granulite facies assemblage containing quartz, plagioclase, hypersthene and augite. Good comparisons are the Archean gneiss terrains of Scotland and East Greenland (Tarney et al, 1979).

5.1.2. Chemical Results

Table 5.2 displays the average composition of 18 analyses of regional rocks as well as their ranges. They

Table 3 2-Value ranges and averages (bracketed) of major and selected trace elements in regional and "transitional" rocks.

	Regional rocks		"Transitional rocks"		"Transitional rocks"	
	Min	Max	Min	Max	Min	Max
SiO ₂ (wt)	69.95-76.96(72.65)	from 16 analyses	65.02-73.96(69.71)	from 10 analyses	60.60-71.09(65.76)	from 5 analyses
TiO ₂	0.01-0.06(0.02)	"	0.03-0.36(0.11)	"	0.22-0.50(0.33)	"
Al ₂ O ₃	11.63-16.93(12.62)	"	13.62-16.75(14.61)	"	13.63-15.60(14.56)	"
FeO _T (t)	0.71-6.93(3.69)	"	0.76-7.96(4.67)	"	2.16-7.96(4.99)	"
FeO _T	0.03-7.52(1.76)	from 16 analyses	0.23-8.31(1.76)	from 6 analyses	2.63	analyses only
FeO	0.57-6.62(2.67)	"	1.03-6.24(2.89)	"	4.66	"
Mg	0.03-0.11(0.05)	from 16 analyses	10-0.07(0.03)	from 10 analyses	10-0.32(0.10)	from 5 analyses
MgO	10-0.29(0.09)	"	10-0.37(0.07)	"	10-0.33(0.15)	"
CaO	0.52-7.63(1.15)	"	0.76-7.09(2.22)	"	0.57-7.57(4.23)	"
MgO	2.42-6.99(3.36)	"	2.96-7.62(4.19)	"	3.30-5.99(4.19)	"
SiO ₂	4.21-6.42(5.03)	"	5.36-6.40(4.40)	"	0.20-5.83(3.87)	"
P ₂ O ₅	10-0.50(0.07)	"	10-0.06(0.02)	"	10-0.09(0.04)	"
L.O.I.	0.03-0.99(0.30)	"	0-0.93(0.33)	"	0-0.67(0.25)	"
Na ₂ O _{eq}	10-39(25)	from 16 analyses	7-37(37)	from 10 analyses	27-60(46)	from 5 analyses
Zr	36-68(46)	"	116-106(100)	"	105-100(100)	"
V	6-19(11)	"	2-60(36)	"	22-10(52)	"
Sc	10-26(18)	"	121-21(100)	"	76-10(127)	"
Rb	115-201(150)	"	0-200(123)	"	0-10(10)	"
Pb	19-34(26)	"	19-61(27)	from 9 analyses	5-19(15)	from 3 analyses
Zn	23-69(77)	"	10-62(69)	"	16-50(29)	"
Cu	9-10(10)	"	4-12(10)	"	9-11(10)	"
Hg	2-7(5)	"	3-10(5)	"	4-6(5)	"
Co	17-99(37)	"	17-16(55)	"	11-36(45)	"
Ce	4-7(5)	"	3-10(7)	from 10 analyses	5-10(7)	from 5 analyses
La	229-225(199)	"	196-23(160)	from 9 analyses	71-176(100)	from 1 analysis
V	0-10(5)	"	0-11(5)	"	1-5(5)	"
S	43-104(57)	"	51-10(6)	"	17-10(4)	"
Co	15-27(20)	"	15-10(25)	"	22-31(25)	"
Se	10	"	2-7(4)	from 5 analyses	1 and 4	2 analyses only
Te	10	"	4-10(11)	"	9	"
Mo	10	"	9-11(10)	"	10 and 10	"
Bi	10	"	15-27(20)	"	11 and 11	"
U	1.52-9.20(3.86)	from 17 analyses	2.92-7.62(4.99)	from 10 analyses	2.63-10.04(5.86)	from 5 analyses

* With the exception of 2 samples of amphibolites and 1 biotite-granulite.

L.O.I.: (loss of ignition) is mainly H₂O as indicated by microprobe.FeO_T(t): total iron as FeO_T.

10: not analyzed 10: not detected

are quite uniform geochemically over the area examined. The majority is in the acid ($\text{SiO}_2 > 70\%$) and some are in the sub-acid ($65\% < \text{SiO}_2 < 70\%$) ranges. They are very rich in K and some in Na, being in good agreement with a granitic composition (Joplin, 1971), and similar to parts of other "typical" basement terrains around the world (O'Nions and Pankhurst, 1979).

• Their Na/K ratios vary from 0.34 to a maximum of 0.93, but most are around ~0.5; K% varies from 3.62 to 4.97, with an average of 4.17 (from 18 samples), slightly higher than the mean content of granitic rocks (Haack, 1983). Total iron is variable with $\text{Fe}^{+3}/\Sigma \text{Fe}$ ranging from 0.11 to 0.27 (average from 13 samples: 0.17). CaO is fairly low and its variation is mainly related to varying plagioclase compositions and concentration.

Their K/Rb ratios vary from 204 to a maximum of 396, with an average from 16 samples of 280. This ratio is a little higher, but very close to the average of 230 for normal upper crustal rocks (from Taylor, 1964 in Heier, 1973); there is no tendency of increasing K/Rb ratios with decreasing K, as it is commonly observed (Shaw, 1968 in Tarney and Windley, 1977). The one granulite sample (AM 10) however, shows a K/Rb ratio of 1027, which is in good agreement with Rb depletion, characteristic of some granulite terrains like the Lewisian (Scotland) and East

Greenland (Tarney and Windley, 1977).

Figueiredo (1980) has shown that high-grade gneisses of the Caraiba Complex in the northeast of Bahia State, also show high K/Rb ratios (405 and 653); an average value of 626 was found by Sighinolfi (1971) for the medium to high pressure Itabuna granulites (in southeastern Bahia). Geological as well as geochronological and geochemical information are insufficient to draw conclusions on the temporal relationship between coexisting amphibolite and granulite facies rocks in the Paramirim Complex. The chemical characterization of contemporaneous segments of Archean lower (granulite facies) and middle (amphibolite facies) crust (Sheraton et al 1973, Weaver and Tarney, 1981, Weaver and Tarney, 1983), have shown that K/Rb ratios of the amphibolite facies gneisses approximate that of 230, whereas the granulites exhibit values close to 1000. Where the granulites show retrograde effects, the K/Rb ratios of altered counterparts (also amphibolite facies rocks) are not significantly reduced (Beach and Tarney, 1978). The fact remains that more studies are necessary in order to conclude whether the amphibolite facies rocks of the Paramirim are a product of retrogressive metamorphism of early granulite complexes, or contemporaneous as part of a prograde sequence.

Sr is variable and is positively correlated with Ca, suggesting that most of the Sr is contained in plagioclase. Ba is relatively high and always $Ba > Sr$, as it has been observed in tonalites of the Lewisian (Tarney et al, 1979). K and Ba have a rough positive correlation and K/Ba ratios vary from 2.2 to 72. Rb/Sr values are high and vary from 0.47 to 3.45 (average from 16 samples: 1.80); the sample of granulite has a value of 0.02 like that for the lower continental crust (Hurley, 1968).

The average U content in the regional gneisses is 3.86 ppm, in conformity with published mean contents for granitic rocks of 2.2-15ppm (see Langmuir, 1978) and 5.4 (Haack, 1982 in Haack, 1983); it is about 1.4 times higher than the average crustal abundance of 2.6ppm (Dyck, 1978). The U/Pb average ratio is 0.15.

5.2. "Transitional" rocks

5.2.1. Introduction

Type A "transitional" rocks represent the least modified suite of rocks in the zone of alteration, in both mineralogical and chemical terms, being principally characterized by highly sheared textures. Type B "transitional" rocks show a number of changes in their

mineralogy. For this reason, this group of rocks is extremely important in attesting to the initial stages of alteration. A comparative chemical study of these with the regional rocks will be presented. Chemical analyses can be found in Tables 5.3 and 5.4, and their averages and ranges in Table 5.2.

5.2.2. Type A

The overall chemistry of type A "transitional" rocks is rather similar to that of the regional rocks, though more variable. Compared with the chemical signature of the regional gneisses, in type A "transitional" rocks:

- SiO_2 is also high, and the samples are in the acid and sub-acid ranges as well, but the average SiO_2 content is slightly lower (68.71%);

- K_2O is practically the same; Na_2O values are slightly higher;

- Na/K ratios vary from low (0.3) to as much as 24.90. The average ratio from 10 samples is 3.99, that is 8 times that of the regional rocks. Such difference can probably be linked to the neoformation of albite as rims surrounding original plagioclase;

Table 5.3- Major and selected trace element abundances in type A "transitional rocks".

Sample no.	1 BR01 17.02- 17.30m	2 BR01 18.04- 18.10m	3 BR01 19.09- 19.14m	4 BR01 29.07- 29.15m	5 BR01 107.00- 107.10m	6 BR01 160.00- 160.18m	7 BR01 172.00- 172.08m	8 BR01 204.92- 209.00m	9 442	10 447 52.63- 52.67m
SiO ₂ (wt)	65.66	70.29	71.99	65.52	65.02	70.52	64.00	64.30	73.96	70.00
TiO ₂	0.32	0.26	0.24	0.56	0.35	0.16	0.42	0.38	0.08	0.31
Al ₂ O ₃	16.16	15.14	14.64	14.67	13.42	14.76	13.91	15.14	14.26	14.10
Fe ₂ O ₃	5.96	3.23	0.21	1.71	7.99	1.92	4.51	1.25	0.79	3.05
FeO	3.37		2.39	5.24		1.08	1.74	3.51		
MnO	0.07	ND	ND	0.10	ND	0.03	0.07	0.07	ND	0.05
MgO	ND	ND	0.04	0.04	0.37	ND	0.02	0.06	0.10	0.09
CaO	2.07	1.35	1.43	3.26	7.09	0.99	1.33	2.44	0.74	1.54
Na ₂ O	3.67	3.23	5.23	3.35	4.46	3.51	3.38	4.28	7.42	2.96
K ₂ O	6.10	6.47	2.96	4.74	0.40	6.27	5.67	5.02	72.29	6.70
P ₂ O ₅	0.04	ND	ND	0.06	ND	ND	0.04	ND	ND	0.05
LOI	0.40	0.30	0.50	0.20	-	0.10	0.20	0.20	0.50	0.91
Total	99.19	100.46	99.79	100.03	99.09	99.26	99.38	99.12	94.44	100.00
Spec. Gravity (g/cm ³)	2.66			2.76		2.63	2.70			
Nb(ppm)	36	11	4	72	77	7	43	53	24	40
Zr	627	337	340	1066	757	246	760	702	119	540
Y	31	14	12	60	52	4	66	39	2	40
Sr	209	206	151	190	203	167	179	216	121	200
Rb	142	153	73	101	171	134	135	117	-	200
Pb	27	28	19	29	41	23	29	26	24	
Zn	106	69	62	162	20	23	58	112	10	
Cu	10	9	12	11	12	9	10	9	4	
Ni	4	5	4	4	7	4	10	4	3	
Co	54	62	103	116	24	17	25	45	49	
Cr	-	6	7	9	7	4	7	7	3	10
Ba	2114	2314	974	1648	1793	1976	1926	1429	194	
V	10	6	3	-	4	11	1	4	10	
S	58	40	160	112	67	60	53	103	51	
Ca	28	22	24	27	15	15	20	14	23	
Sc	-	1	1	7	4	2	4	4		
Ta	12	44	15	10	14	11	11	11		
Hf	19	12	11	33	20	9	25	22		
Th	21	13	20	19	13	24	27	24		
U	3.35	3.02	2.92	3.40	3.45	4.59	7.42	4.15	7.59	5.21
Fe ³ /ΣFe	0.20		0.07	0.23		0.60	0.70	0.74		

*total iron

L.O.I.: (loss of ignition) is mainly H₂O as indicated by
petrography

ND: not detected

1-9, 10: amphibole-microcline-quartzites

9: quartz-oligoclase (oligoclase)

Table 5.4- Major and selected trace element abundances
in type B "transitional rocks".

Sample no.	1 6801 107.11- 107.14m	2 9901 128.95- 129.00m	3 AN3 RAA368	4 AN3 RAA413	5 AN6
SiO ₂ (wt%)	71.09	64.43	69.20	60.40	68.70
TiO ₂	0.24	0.30	0.37	0.50	0.22
Al ₂ O ₃	14.26	13.45	14.40	15.40	15.20
Fe ₂ O ₃	3.19*	2.43	3.99*	7.63*	2.18*
FeO		4.46			
H ₂ O	ND	0.32	0.02	0.13	0.01
H ₂ O	0.17	ND	0.06	0.21	0.33
CaO	0.57	2.57	1.78	6.43	4.90
Na ₂ O	5.59	4.93	3.30	3.15	3.99
K ₂ O	4.27	0.20	5.43	5.47	3.57
P ₂ O ₅	ND	ND	0.07	0.09	0.05
LOI	-	0.30	0.47	0.16	0.31
Total	99.38	98.68	99.60	99.80	99.36
Nb(ppm)	27	38	60	50	54
Zr	345	530	990	1040	517
Y	22	31	60	90	56
Sr	144	74	270	750	395
Rb	99	-	150	110	92
Pb	19	9			18
Zn	14	59			15
Cu	9	11			9
Mn	5	6			4
Co	47	54			33
Cr	5	6	<10	<10	6
Ra	1704	71			1046
V	1	5			4
S	42	144			37
Ge	22	23			11
Sc	3	4			
Ta	9	9			
Hf	10	14			
Th	13	11			
U	2.43	4.31	10.8	4.36	7.22
Fe ³⁺ /ΣFe		0.33			

*total iron L.O.I.: (loss of ignition) is mainly
H₂O as indicated by petrography
ND: not detected

1,4r amphibole-pyroxene-microcline-quartz
2r amphibole-pyroxene-garnet-microcline-quartz
3,4r amphibole-garnet-microcline-quartz

- total iron values are essentially the same, but $\text{Fe}^{+3}/\Sigma\text{Fe}$ ratios are variable; the average $\text{Fe}^{+3}/\Sigma\text{Fe}$ ratio is 0.34 which is two times that of the regional rocks, though values as low as 0.07 are found; however, such minor variation cannot be considered representative;

- although an increase in the average CaO value is noted, it is due to the high CaO content in one sample (analysis no. 5; CaO: 7.09%). All other samples have CaO values which are close to those encountered in regional gneisses, and if this value of 7.09% is excluded from the average calculation, a value of 1.68% is obtained, similar to that of regional rocks.

No major differences were found in the trace element abundances. To be noted is the somewhat lower Rb/Sr ratio, with an average of 0.64. U concentrations are slightly higher but the magnitude of this enrichment is so small, that speculations on possible causes are unwarranted. Th contents average 20 ppm, similar to the mean value for granitic rocks (Haack, 1983). Th/U ratios range from 3.45 to 6.85 with an average of 5.13 (from 6 analyses), also close to ratios found for granitic rocks. U/Pb average ratio is 0.17.

5.2.3. Type B

Only five analyses of type B "transitional" rocks are available (Table 5.4). Nonetheless, a few observations can be drawn; relative to regional and type A "transitional" rocks:

- somewhat lower SiO_2 values are found (average: 66.76%), and these are lowest in samples where amphibole replacement is at its most advanced stage (analyses nos. 2 and 4, Table 5.4);
- total iron values are rather variable, as a consequence of varying mafic minerals concentrations;
- K_2O is lower compared to both regional and type A "transitional" rocks, and Na_2O contents are roughly the same as in type A; Na/K ratios vary from 0.50 to 24.65, with an average of 5.56 (~1.5 times higher than that of type A "transitional" rocks, and ~10 times that of regional rocks);
- CaO contents are variable and are primarily related to high garnet proportions (analyses nos. 2 and 4) and high oligoclase proportions (analysis no. 5).

Among the trace elements, Rb and Sr show the most significant differences. Rb is substantially lower and Sr higher than those values obtained for regional and type A

"transitional" rocks. As a consequence, Rb/Sr ratios are low, with an average of 0.32 which is almost 6 times lower than that of regional rocks. Differences in U concentrations are again too small to be evaluated. Th was determined in two samples only, and Th/U ratios are 4.90 and 2.55 respectively, still within crustal averages. U/Pb ratios average 0.36.

5.3. Metasomatic rocks

5.3.1. Albitites

a) Chemical results

Table 5.5 shows results of the chemical analyses obtained for common and Ca-rich albitites; Table 5.6 displays averages and ranges. Based on the chemical analyses alone, the noticeable metasomatic changes in relation to both regional and "transitional" rocks are:

- a decrease in SiO_2 content, with most samples containing ~60%;

- a decrease in K_2O content;

- an increase in Na_2O ; Na/K average ratios of albitites and mineralized albitites are 23 and 43,

Table 3
(Cont.)

	20 8804	21 8805	22 8806	23 8807	24 8808	25 8809	26 8810	27 8811	28 8812	29 8813	30 8814	31 8815	32 8816	33 8817	34 8818	35 8819	36 8820	37 8821	38 8822	39 8823	40 8824
Temp. °C	167.75	167.75	168.15	168.55	168.95	169.35	169.75	170.15	170.55	170.95	171.35	171.75	172.15	172.55	172.95	173.35	173.75	174.15	174.55	174.95	175.35
Wt. (g)	61.65	61.65	61.65	61.65	61.65	61.65	61.65	61.65	61.65	61.65	61.65	61.65	61.65	61.65	61.65	61.65	61.65	61.65	61.65	61.65	61.65
Wt. (g)	2.78	2.78	2.78	2.78	2.78	2.78	2.78	2.78	2.78	2.78	2.78	2.78	2.78	2.78	2.78	2.78	2.78	2.78	2.78	2.78	2.78
Wt. (g)	15.75	15.75	15.75	15.75	15.75	15.75	15.75	15.75	15.75	15.75	15.75	15.75	15.75	15.75	15.75	15.75	15.75	15.75	15.75	15.75	15.75
Wt. (g)	2.78	2.78	2.78	2.78	2.78	2.78	2.78	2.78	2.78	2.78	2.78	2.78	2.78	2.78	2.78	2.78	2.78	2.78	2.78	2.78	2.78
Wt. (g)	1.62	1.62	1.62	1.62	1.62	1.62	1.62	1.62	1.62	1.62	1.62	1.62	1.62	1.62	1.62	1.62	1.62	1.62	1.62	1.62	1.62
Wt. (g)	0.09	0.09	0.09	0.09	0.09	0.09	0.09	0.09	0.09	0.09	0.09	0.09	0.09	0.09	0.09	0.09	0.09	0.09	0.09	0.09	0.09
Wt. (g)	2.99	2.99	2.99	2.99	2.99	2.99	2.99	2.99	2.99	2.99	2.99	2.99	2.99	2.99	2.99	2.99	2.99	2.99	2.99	2.99	2.99
Wt. (g)	1.62	1.62	1.62	1.62	1.62	1.62	1.62	1.62	1.62	1.62	1.62	1.62	1.62	1.62	1.62	1.62	1.62	1.62	1.62	1.62	1.62
Wt. (g)	9.16	9.16	9.16	9.16	9.16	9.16	9.16	9.16	9.16	9.16	9.16	9.16	9.16	9.16	9.16	9.16	9.16	9.16	9.16	9.16	9.16
Wt. (g)	9.16	9.16	9.16	9.16	9.16	9.16	9.16	9.16	9.16	9.16	9.16	9.16	9.16	9.16	9.16	9.16	9.16	9.16	9.16	9.16	9.16
Wt. (g)	0.02	0.02	0.02	0.02	0.02	0.02	0.02	0.02	0.02	0.02	0.02	0.02	0.02	0.02	0.02	0.02	0.02	0.02	0.02	0.02	0.02
Wt. (g)	0.20	0.20	0.20	0.20	0.20	0.20	0.20	0.20	0.20	0.20	0.20	0.20	0.20	0.20	0.20	0.20	0.20	0.20	0.20	0.20	0.20
Total	99.60	99.60	99.60	99.60	99.60	99.60	99.60	99.60	99.60	99.60	99.60	99.60	99.60	99.60	99.60	99.60	99.60	99.60	99.60	99.60	99.60
Total Gravity																					
(g/cm ³)	2.80	2.80	2.80	2.80	2.80	2.80	2.80	2.80	2.80	2.80	2.80	2.80	2.80	2.80	2.80	2.80	2.80	2.80	2.80	2.80	2.80
 
Wt. (g)	9.16	9.16	9.16	9.16	9.16	9.16	9.16	9.16	9.16	9.16	9.16	9.16	9.16	9.16	9.16	9.16	9.16	9.16	9.16	9.16	9.16
Wt. (g)	0.02	0.02	0.02	0.02	0.02	0.02	0.02	0.02	0.02	0.02	0.02	0.02	0.02	0.02	0.02	0.02	0.02	0.02	0.02	0.02	0.02
Wt. (g)	0.20	0.20	0.20	0.20	0.20	0.20	0.20	0.20	0.20	0.20	0.20	0.20	0.20	0.20	0.20	0.20	0.20	0.20	0.20	0.20	0.20
Total	99.60	99.60	99.60	99.60	99.60	99.60	99.60	99.60	99.60	99.60	99.60	99.60	99.60	99.60	99.60	99.60	99.60	99.60	99.60	99.60	99.60
Total Gravity																					
(g/cm ³)	2.80	2.80	2.80	2.80	2.80	2.80	2.80	2.80	2.80	2.80	2.80	2.80	2.80	2.80	2.80	2.80	2.80	2.80	2.80	2.80	2.80

Total Iron (g) not detected

20-25: mineralized garnet-syncline-schistosity

11-12: garnet-syncline-schistosity (Co-rich)

11-12: mineralized garnet-syncline-schistosity (Co-rich)

Table 5.5
(Cont.)

Sample no.	39 440 LL45-9	40 440 LL45-12	41 440 LL45-13	42 440 LL45-14	43 R487 75.32- 75.38m	44 440 LL44-A	45 440 LL44-A	46 1041-206	47 440 RAA336	48 440 RAA336
SiO ₂ (wt)	55.29	55.56	61.03	60.67	61.70	61.49	60.00	60.12	60.50	55.60
TiO ₂	0.40	1.00	0.51	0.24	0.32	0.27	0.31	0.22	0.46	0.21
Al ₂ O ₃	13.55	10.36	14.57	17.10	13.10	16.62	16.78	18.51	14.90	9.12
Fe ₂ O ₃	6.91	6.69	3.67	2.98	4.93	5.47	5.91	4.18	7.24	10.90
FeO	0.91	3.92	0.66	0.49		0.41	0.70	0.48		
MnO	0.08	0.12	0.09	0.06	0.05	0.06	0.03	0.06	0.13	0.37
MgO	1.00	4.01	1.89	0.64	0.43	ND	ND	ND	0.50	4.47
CaO	13.74	11.05	9.52	7.73	13.10	10.22	11.44	13.90	11.20	12.30
Na ₂ O	7.64	6.21	9.93	9.79	1.20	3.04	2.29	ND	4.46	5.20
K ₂ O	0.15	0.17	0.16	0.17	0.18	ND	ND	ND	0.40	0.40
P ₂ O ₅	0.10	0.24	0.09	0.04	0.07	0.03	0.04	0.07	0.06	0.06
LOI	0.37	0.47	0.20	0.79	1.23	2.46	2.49	2.38	2.08	0.39
Total	100.09	100.23	100.69	100.49	100.50	100.10	100.07	100.05	100.10	99.32
Nb(ppm)	90	222	126	74	60	72	76	71	60	9
Zr	577	2216	951	665	530	477	524	658	990	414
Y	32	94	77	40	90	103	49	74	160	29
Sr	155	102	237	142	790	918	940	1066	410	215
Rb	30	65	36	38	10	26	20	20	40	-
Pb	369	743	462	339		223	239	209		12
Zn	62	161	46	59		26	23	22		100
Cu	10	11	10	9		10	10	10		10
Hf	5	9	5	6		7	6	6		5
Co	6	4	10	23		16	20	14		17
Cr	10	10	9	6	<10	4	9	10	10	9
Ba	207	404	144	118		942	444	433		90
V	347	195	68	254		131	153	135		434
S	30	45	32	30		25	35	34		64
Ga	11	24	25	31						22
Sc										4
La										5
Pr										12
Th										21.1
U	2800	6200	4100	1760	9.80	1640	1710	1410	429	16
Fe ²⁺ /ΣFe	0.88	0.60	0.83	0.84		0.92	0.88	0.94		

*total iron ND: not detected

39-42: mineralized garnet-pyroxene-albitites (Ca-rich)

43: epidote (type 2)

44-47: mineralized epidotes (type 1)

48: pyrox-amph-albitite (pyr - amph 70%)

respectively, and 55 for Ca-rich albitites;

- an irregular but noticeable increase in both Fe_2O_3 and CaO ; $\text{Fe}^{+3} / \Sigma \text{Fe}$ ratios vary from 0.34 to 0.97, with an average of 0.7 for each one of the groups of albitites;

- a very small increase in MgO and TiO_2 ;

- among the trace elements, there is a tendency towards slightly higher Nb, Zr and Y; Sr, Pb, V and naturally U are much higher whereas Rb and Ba are definitely lower.

b) Chemical mass balance studies

Examination of Table 5.5 shows that there are significant differences in the chemical compositions between the albitites and their "parent rocks", i. e., the regional and "transitional" rocks. Mass balance studies (Lobato et al, 1982), were carried out with the purpose of comparing a number of type A "transitional rocks" with common mineralized and unmineralized albitites. Considering the petrographic evidence for metasomatism, four samples of type A "transitional" rocks (Table 5.3 analyses nos. 1, 4, 6 and 7) were averaged and used as "parent rock"; nine samples of albitites were used as daughter products (Table 5.5 analyses nos. 5, 7, 8, 17, 19, 21, 22, 23).

Gresens (1967) suggested incorporating specific gravity data into two-way mass balance calculations, such that fixing either volume change or the behaviour of one component during a reaction provides a unique solution. Costa's (1980) approach to Gresen's work was used in the present study. The magnitude of mass transfer (ΔX) of a component "i" in a n-component system, where rock A is being altered to rock B can be related to the initial and final compositions of the rock by:

$$\Delta X_i = b(X_i^B) - a(X_i^A), i=1,2,3,\dots,n$$

where, X_i^A and X_i^B are the weight fraction of component "i" in rocks A (parent) and B (daughter); "a" is the initial weight of rock A, usually considered as 100g, from the standard whole rock analyses; "b" is the weight of the daughter rock produced by the alteration of initial weight of rock A. The volumes occupied by "a" grams of A and "b" grams of B can be related by the ratio of final and initial volumes:

$$K_v = V_B/V_A = b \cdot \rho_A / a \cdot \rho_B$$

where, K_v is the volume factor, or ratio between final and initial volumes; V is volume; and ρ is specific gravity (measured, see Tables 5.3 and 5.5). Following Gresens' (1967) work, a $K_v = 1$ means isovolumetric metasomatism and K_v s of 0.5 and 1.5, for instance, express 50% volume reduction and 50% volume increase, respectively, from the parent to the altered product. Combining the two equa-

tions above, the obtained expression permits the quantitative evaluation of mass transfer:

$$\Delta X_i = a[K_V \rho B / \rho A (X_i^B) - (X_i^A)]$$

In this expression both ΔX_i and K_V are unknown, and a solution is possible only if either K_V is considered to be 1 (isovolumetric alteration) or by assigning an immobile behaviour to one component; in this case $\Delta X_i = 0$.

With that in mind, K_V s (volume factors) were calculated by assuming $\Delta X_i = 0$ for each one of the elements being considered. A lack of convergence of K_V values for all elements except Al_2O_3 , prompted the use of the average K_V value for that element; a value of 0.9 was found, signifying 10% volume reduction. Gains and losses (per 100g of parent rock) were then calculated (see Table 5.7). These were used to make a series of plots of components versus progressive alteration as indicated by increasing Na_2O content, in an attempt to assess the behaviour of the various components during the alteration. The results obtained in terms of gains and losses are very variable for most elements, with no obvious definite trends along the alteration profile. Nevertheless, Na_2O shows consistent gain throughout the alteration, whereas SiO_2 , K_2O , Rb and Ba appear to have been removed.

Table 5.7-Major oxides for parent rock and compositional changes (gains and losses) for the transformation of type A "transitional" rocks to unmineralized and mineralized albitites (in grams per 100 g parent-rock). Transformations arranged according to increasing Na_2O gain (see 5.3 and Fig. 5.2).

Parent ^a Rock	PR01 185.80- 185.90m	PR01 185.10- 185.20m	PR01 186.70- 186.80m	PR01 187.60- 187.72m	PR01 188.30- 188.35m	PR01 ^b 188.90- 189.00m	PR01 189.63- 189.70m	PR01 179.95- 180.00m	PR01 189.13m
SiO_2	-11.05	-10.60	-7.58	-7.76	-8.54	-8.95	-6.12	-7.84	-6.41
TiO_2	+0.15	0.00	+0.05	-0.04	-0.08	+0.19	-0.09	-0.06	-0.07
Al_2O_3	+1.29	+0.05	+1.29	+1.25	+0.82	+0.95	+1.62	+2.21	+1.95
Fe_2O_3	+2.13	+0.41	-0.41	+0.08	+0.93	-1.78	+0.07	+0.01	-0.66
MnO	-0.04	+0.04	+0.03	+0.02	+0.04	+0.58	+0.05	-0.05	0.00
MgO	+0.03	+0.22	+0.09	0.00	+0.09	+0.46	0.00	+0.03	+0.01
CaO	-0.77	-4.20	+3.87	+3.79	-4.67	+1.66	-4.31	-1.11	+2.54
Na_2O	+3.25	-4.85	+5.92	+6.02	+6.25	6.43	+6.43	6.46	+6.88
K_2O	-2.06	-5.32	-5.33	-5.44	-5.41	-5.33	-5.27	-5.31	-5.42
Spec. Grav. vol	0.92	0.9	0.93	0.935	0.941	0.959	0.98	0.988	0.98

^a: average from samples PR01 17.02-17.10m, 29.07-29.15m, 160.00-160.10m, 172.00-172.00m

^b: volume factors

Figure 5.2 shows gains and losses (in grams per 100g of parent) for K_2O , Na_2O , SiO_2 , Al_2O_3 , CaO and Fe_2O_3 versus Na_2O contents. K_2O is lost in roughly the same amount that Na_2O is gained (average Na_2O gain: 5.83g; average K_2O loss: 5.00g). There is an average Al_2O_3 gain of 1.3g, and an average SiO_2 loss of 8.3g. Overall silica depletion is probably higher because SiO_2 values in regional rocks tend to be higher than those in type A "transitional" rocks.

In conclusion, mass balance results indicate that, in good agreement with petrographic indications, microcline $\rightarrow$ albite reaction was dominant during the metasomatic process, coupled with loss of silica resulting in a 10% volume decrease.

c) Discussion

A plot of Na_2O versus Al_2O_3 (Fig. 5.3) shows a positive correlation. As this behaviour is simply related to albite formation, Na_2O/Al_2O_3 ratios have been used in a number of cases to help assess the variations experienced by other elements during Na metasomatism.

Na_2O/Al_2O_3 versus K_2O/Al_2O_3 shows that there is an antipathetic relationship between those two elements (Fig. 5.4) as one moves away from the regional and "transi-

Fig. 5.2. Results from chemical mass balance: gains and losses expressed in grams per 100g of parent for Na, Ca, Al, Fe, K and Si as a function of increasing alteration (increasing Na₂O); see Table 5.7. Square: parent rock; circles: common unmineralized and mineralized albitites (Table 5.5, analyses nos. 5, 7, 8, 17, 19, 21, 22, 23).

Fig. 5.3. Al_2O_3 versus Na_2O for regional, "transitional" and metamorphic rocks. For symbols, see Fig. 4.1.

Fig. 5.4. K_2O/Al_2O_3 versus Na_2O/Al_2O_3 ratios in regional, "transitional" and metasomatic rocks. For symbols, see Fig. 4.1.

tional" rocks fields into the field of the albitites. Silica depletion is clear from Fig. 5.5 ; in the field of the albitites themselves, there is a positive correlation between Na_2O and SiO_2 due to varying albite amounts in different samples, compared to mafic minerals amounts (pyroxene and garnet), which contain lower SiO_2 values.

Although Fe_2O_3 , CaO , MgO and TiO_2 all present a slight general increase in common albitites, such increase is probably related to the volume loss. The mineral reactions found to occur during the alteration process (see Chapter 6), do not account for the introduction of either one of these elements, so it is improbable that they were added together with Na. The variations observed in the percentages from sample to sample must be explained by the biased sampling, because in some cases ideal homogeneity was lacking due to small specimens. $\text{CaO} \times \text{Na}_2\text{O}$ and $\text{Na}_2\text{O}/\text{Al}_2\text{O}_3 \times \text{TiO}_2$ plots illustrate the behaviour of these elements (Figs. 5.6 and 5.7). CaO is higher by about 3-4 times and negatively correlates with Na, reflecting varying albite/mafic minerals concentrations; very high CaO values observed in Ca-rich albitites must reflect original bulk compositions. Though slightly higher, TiO_2 increase is not significant.

Fig. 5.5. Na_2O versus SiO_2 in regional, "transitional" and metasomatic rocks (see text). For symbols, see Fig. 4.1.

Fig. 5.6. Na₂O versus CaO in regional, "transitional" and metasomatic rocks (see text). For symbols, see Fig. 4.1.

Fig. 5.7. $\text{Na}_2\text{O}/\text{Al}_2\text{O}_3$ ratios versus TiO_2 in regional, "transitional" and metasomatic rocks (see text); vertical scale is a log scale. For symbols, see Fig. 4.1.

Also among the trace elements, the small increases in Nb, Zr and less in Y are probably apparent and again may be related to volume loss. Figs. 5.8 and 5.9 exhibit plots of Nb versus $\text{Na}_2\text{O}/\text{Al}_2\text{O}$ and Zr versus Na_2O . Sr increase is perhaps related to CaO increase; though Ca and Sr are still roughly correlated, this correlation is not as good as that found for the regional gneisses.

A drastic decrease in Ba is most certainly the result of K_2O removal as these two elements correlate well; Ba was probably removed as k-feldspar was replaced. Also related to K_2O loss is the decrease in Rb; Rb/Sr ratios are much lower and their averages are : 0.24 (unmineralized albitites), 0.64 (mineralized albitites), 0.21 (Ca-rich albitites).

Th is practically unchanged when compared to values obtained for "transitional" rocks; Th/U ratios are extremely low and average 1.55 in unmineralized albitites and 0.008 in mineralized albitites.

V is clearly higher in all albitites (as high as 870 ppm). A plot of V versus U shows that vanadium is about 10 times higher in altered rocks (Fig. 5.10). Such considerable increase and positive correlation with U seem to suggest that V was also introduced during the alteration process. In igneous rocks, vanadium tends to

Fig. 5.8. $\text{Na}_2\text{O}/\text{Al}_2\text{O}_3$ ratios versus Nb in regional, "transitional" and metasomatic rocks (see text); vertical scale is a log scale. For symbols, see Fig. 4.1.

Fig. 5.9. Na₂O versus Zr in regional, "transitional" and metasomatic rocks (see text); vertical scale is a log scale. For symbols, see Fig. 4.1.

Fig. 5.10. U versus V in regional, "transitional" and metamorphic rocks (see text); both vertical and horizontal scales are log scales. For symbols, see Fig. 4.1.

become concentrated in the SiO_2 undersaturated rocks: averages of 231 ppm (50-55% SiO_2), 225 ppm (45-50% SiO_2) and 319 ppm for the range of 45 - 40% SiO_2 are reported by Evans Jr. and Landergren, (1978). Vanadium is commonly enriched in iron-rich pyroxenes (Evans Jr. and Landergren, 1978), and thus it is probably present in aegirine-augite.

Uranium values are generally very high, even in the unmineralized albitites. Although U correlates well with Na_2O (Fig. 5.11), it best correlates with Fe^{+3} . Fig. 5.12 is a plot of $\text{Fe}^{+3} / \Sigma\text{Fe}$ versus U; examination of that plot seems to suggest that U is better concentrated in oxidized rocks and a ratio of 0.5 and higher seems to be preferable; a $\text{Fe}^{+3} / \Sigma\text{Fe}$ ratio of unity represents 100% oxidation.

5.3.2. Epidosites

The chemistry of epidosites is basically much like that of all other metasomatic rocks except that they are very rich in CaO (much like Ca-rich albitites), poor in or lacking Na_2O and show higher values of LOI. Sodium impoverishment does not allow the construction of plots using Na_2O as a measure of alteration and the use of U is probably best for that purpose. Nonetheless, all trace elements follow the same trends previously described.

Fig. 5.11. Na₂O versus U for regional, "transitional" and metamorphic rocks (see text); vertical scale is a log scale. For symbols, see Fig. 4.1.

Fig. 5.12. $\text{Fe}^{+3}/\Sigma\text{Fe}$ ratios versus U in regional, "transitional" and metasomatic rocks (see text). Note that a $\text{Fe}^{+3}/\Sigma\text{Fe}$ ratio of 1.0 represents 100% oxidation. For symbols, see Fig. 4.1.

suggesting that in spite of their different mineralogy the epidiosites probably originated during the same alteration process that gave place to the albitites.

5.4. $^{87}\text{Sr}/^{86}\text{Sr}$ systematics

Rb-Sr isotopic tracer studies have been done to help address the question of uranium provenance. Although Sr is apparently coenriched with U during Na alteration, this does not necessarily imply that uranium and strontium have originated from a common reservoir. Nevertheless, strontium isotope studies can constrain the possible rock reservoirs with which the alteration solution equilibrated. The results obtained from these studies are summarized in Table 5.8 ; these results were first reported in Kerrich et al (1984).

The regional gneisses hosting the U mineralization have an age of 2.86 Ga (Cordani, 1982), elevated Rb-Sr, and therefore have evolved along a steep $^{87}\text{Sr}/^{86}\text{Sr}$ trajectory through time. Metasomatic alteration accompanying U deposition, resulted in almost total loss of Rb and increase in Sr. Thus Rb/Sr ratios are very low (see 5.3.1. c), such that little ^{87}Sr has evolved.

Table 5.8- Rb-Sr data on whole rocks.

Sample	Rock type	Rb (ppm)	Sr (ppm)	$\frac{Rb}{Sr}$	$\frac{87Sr}{86Sr}$ (measured)	$\frac{87Rb}{86Sr}$	$\frac{87Sr}{86Sr}$ at $t=0.82$ by	$\frac{87Sr}{86Sr}$ at $t=2.86$ by
REGIONAL GNEISSES (this work)								
AN6	hornblende-quartz-microcline-gneiss	155	181	0.8563	0.7753	2.433	0.7468	0.6745
" 8	hornblende-quartz-microcline-gneiss	148	229	0.6463	0.7594	1.833	0.7379	0.6835
" 40	hornblende-quartz-microcline-gneiss	200	58	3.4483	0.9308	9.946	0.8143	0.5185
" 43	hornblende-quartz-microcline-gneiss	139	68	2.0441	0.8509	5.851	0.7824	0.6084
LR20-B*	biotite-gneiss	142	178	0.7977	0.7945	2.33		
LR20-E*	biotite-gneiss	187	72	2.5972	1.0279	7.75		
LR20-F*	biotite-gneiss	208	61	3.4098	1.0886	10.13		
LR20-G*	biotite-gneiss	243	54	4.5000	1.2064	13.63		
LR20-I*	biotite-gneiss	181	103	1.7572	0.9162	5.17		
U ORE DEPOSIT								
BR-01-185.62	aegirine-augite-albite	-	185		0.7232		0.7232	
" 185.80	magnetite-albite	180	238	0.7563	0.73908	2.141		
" 186.00	andradite-albite	25	177	0.1412	0.7281	0.394	0.7234	
" 186.70	andradite-albite	21	126	0.1667	0.7225	0.471	0.7169	
" 188.30	andradite-albite	-	121		0.7220		0.7220	
" 189.63	albite	-	118		0.7226		0.7226	
AN7-LL45-5	andradite-diopside-albite	63	131	0.4092	0.7204	1.359	0.7045	

-: less than 15 ppm Rb

*samples from data in Cordani (1982)

The best current estimate of the age of the uranium mineralization is 820 Ma by U-Pb age determination (Cooper, 1979). To try to determine $^{87}\text{Sr}/^{86}\text{Sr}$ evolution of regional gneisses through time, Cordani's (1982) results were used. An average value of 7.802 for the $^{87}\text{Rb}/^{86}\text{Sr}$ ratio of the regional gneisses is obtained from five samples of augen-gneisses defining an isochron, with an initial $^{87}\text{Sr}/^{86}\text{Sr}$ ratio of 0.705 at the time of last strontium isotope homogenization (2.86 Ga). Combining these values with the elapsed time to mineralization (that is, 2.04 Ga), an average $^{87}\text{Sr}/^{86}\text{Sr}$ of 0.934 at 820 Ma is obtained. This assumes that the samples used to construct the isochron are representative of the gneisses as a whole. An array for the strontium isotope evolution of the gneisses is portrayed in Fig. 5.13, calculated from the mean $^{87}\text{Rb}/^{86}\text{Sr}$ and $\text{Rb}/\text{Sr} \pm 2.185$ of the mean (half of the mean, to facilitate construction of the lines).

Four samples of regional gneisses (Table 5.1, analyses nos. 2, 4, 13, 15) were also used to try to estimate the $^{87}\text{Sr}/^{86}\text{Sr}$ of the gneisses at 820 Ma. Back extrapolation of analytically determined $^{87}\text{Sr}/^{86}\text{Sr}$ ratios, using calculated $^{87}\text{Rb}/^{86}\text{Sr}$ from measured Rb/Sr , gives an average ratio of 0.77 at 820 Ma. This value is significantly lower than that calculated based on the isochron data (0.934), and it is best interpreted to

Fig. 5.13. Strontium isotope evolution diagram for regional gneisses (samples are from this work and from Cordani, 1982) and albitites (see text). For derivation of trajectories see text. Numbers on trajectories represent $^{87}\text{Rb}/^{86}\text{Sr}$ ratios (see Table 5.8).

reflect a decrease of the $^{87}\text{Sr}/^{86}\text{Sr}$ ratio of the gneisses by interaction with the metasomatic fluid during uranium mineralization. It is interesting to note that the two samples exhibiting the highest $^{87}\text{Sr}/^{86}\text{Sr}$ values (similar to those of the regional Archean gneisses) are the furthest away from the zone of alteration (see Fig. 5.1); samples AM 6 and 8 are closer to the deposit area and their $^{87}\text{Sr}/^{86}\text{Sr}$ values approximate those obtained for the albitites.

Samples of common unmineralized (Table 5.5, analyses nos. 4, 5, 6, 8) and mineralized (analyses nos. 19, 23) albitites, together with one sample of mineralized Ca-rich albitite (analysis no. 35) were also used during the course of this study. Measured $^{87}\text{Sr}/^{86}\text{Sr}$ ratios of these are surprisingly uniform and in accord with their low Rb/Sr ratios; $^{87}\text{Sr}/^{86}\text{Sr}$ ratios vary from 0.720 to 0.739, with an average of 0.725. Calculated $^{87}\text{Sr}/^{86}\text{Sr}$ at 820 Ma yields an average value of 0.719; at 2.86 Ga an average $^{87}\text{Sr}/^{86}\text{Sr}$ value of 0.70 is obtained.

Strontium evolution arrays are depicted for all rock samples in Fig. 5.13. Fig. 5.14 portrays the variation in strontium isotopes as a function of strontium contamination.

Fig. 5.14. $^{87}\text{Sr}/^{86}\text{Sr}$ of regional gneisses bordering the ore deposits and of albitites (see text).

Examination of Figs. 5.13 and 5.14 coupled with the data presented above concur to the following:

- at the time of uranium mineralization (at 820 Ma), the $^{87}\text{Sr}/^{86}\text{Sr}$ signature of the albitites (0.719) was significantly less radiogenic than that of the host Archean gneisses (0.934);

- in the vicinity of the zones of metasomatic alteration, the regional gneisses have undergone variable Sr isotopic disturbance towards the lower values characteristic of the albitites. The extent of disturbance correlates with the magnitude of Sr introduction, or contamination, Rb loss, and thus with diminished Rb/Sr ratios;

- from Fig. 5.14, it is obvious that samples with the highest Rb/Sr retain $^{87}\text{Sr}/^{86}\text{Sr}$ ratios which are close to the regional Archean gneisses;

- all observations combined may suggest that it is unlikely that Sr, and by extension uranium, were derived from fluids indigenous to the host gneisses.

Further discussion of these results will be presented in Chapter 8.

5.5. Oxygen isotope studies

5.5.1. Results

Oxygen isotope data on mineral separates by Fuzikawa (1985) and earlier results first reported in Lobato et al (1983 b) are presented in Table 5.9 and depicted in Fig. 5.15. The results are reported according to the rock groups scheme defined by petrography (Chapter 3). All results are reported as $\delta^{18}\text{O}/\text{‰}$ (per mil) values relative to SMOW. The foregoing discussion results from a preliminary assessment of the isotope results by this author; a more detailed investigation is under way by K. Fuzikawa.

In the regional gneisses, $\delta^{18}\text{O}$ values for quartz range from 7.8 to 9.7 per mil, feldspar from 6.3 to 7.8, biotite from 2.2 to 5.6 and one determination for magnetite is 0.4 per mil. In type A "transitional" rocks quartz ranges from 8.76 to 12.1, feldspar from 6.99 to 9.8 and magnetite from 0.4 to 2.5 per mil; two determinations on biotites yielded values of 2.8 and 4.7 per mil. In type B "transitional" rocks quartz ranges from 8.7 to 11.2, feldspar from 6.8 to 8.0, and magnetite from 0.9 to 1.7 per mil; two biotite values of 3.4 and 4.2 and one value of 6.2 for pyroxene were also obtained.

Table 3. 3-Step average (average composition of mineral assemblages (and some whole rocks) from regional "transitional" and metamorphic rocks. Numbers from 1 to 15 refer to those in Fig. 5. 15, the sequence for each rock group is by highest to lowest values of $\delta^{18}O$.

Analysis no.	Sample no.	$\delta^{18}O$ quartz wt. %	$\delta^{18}O$ feldspar wt. %	$\delta^{18}O$ biotite wt. %	$\delta^{18}O$ pyroxene wt. %	$\delta^{18}O$ hornblende wt. %	$\delta^{18}O$ quartz wt. %	$\delta^{18}O$ feldspar wt. %	$\delta^{18}O$ biotite wt. %	$\delta^{18}O$ pyroxene wt. %	$\delta^{18}O$ hornblende wt. %	Main mineral composition (wt. %)
1	100-170	10.1	9.6	9.4	9.4	9.4	9.4	9.4	9.4	9.4	9.4	quartz (100), feldspar (100)
2	100-170	10.1	9.6	9.4	9.4	9.4	9.4	9.4	9.4	9.4	9.4	quartz (100), feldspar (100)
 

Table 3.9
(Cont.)

Analysis no.	Sample no.	$\delta^{18}O$ quartz ‰	$\delta^{18}O$ feldspar ‰	$\delta^{18}O$ biotite ‰	$\delta^{18}O$ pyroxene ‰	$\delta^{18}O$ hornblende ‰	$\delta^{18}O$ whole rock ‰	$\Delta^{18}O$ (quartz-feldspar) ‰	$\Delta^{18}O$ (quartz-pyroxene) ‰	$\Delta^{18}O$ (quartz-hornblende) ‰	$\Delta^{18}O$ fluid ‰	% in mineral composition (all)
34	WV 876	6.4	6.5	0.3				1.9	6.1		1.36	pl(80), biot(10), gr(10)
35	WV 773	5.7	6.1					1.6			2.66	pl(50), gr(30), pyr+amph+ep+qtz+cal+ephsur(20)
36	WV 767		1.8		-0.4	-6.0						pl(85), pyr+ep+amph(15), cal, ur
37	WV 766		1.3	-3.6		-6.7					0.46	pl(70), cal+amph(30)
38	WV 762		1.0		-1.4	-4.4						pl(70), pyr(10), gr(10), amphib+ep(10)
39	WV 875	0.6	-1.2 (pink fel)			-7.9		1.8		8.5	-2.36	pl(80), gr(15), amphib(10), biot+amph(15)
40	WV 777		0.4	-3.0		-6.6		2.4			-0.46	pl(70), amphib(25), cal(1), amphib(2)
41	WV 776	0.0	-1.7		-4.1	-8.7		1.7		8.7	-2.46	unavailable
42	WV 775	-0.5	-1.4			-8.4		0.9		7.9	-2.36	unavailable
43	WV 765		-0.9			-7.4					-1.36	pl(85), pyr+ep+amph(15), ur, ap, flu
44	WV 821		-1.2			-8.2					-2.36	pl(85), pyr+amph(10), amphib(15)
45	WV 774	-1.2	-2.6			-9.5		1.4		8.3	-3.36	unavailable

ur: orthoclase; mic: microcline; pl: plagioclase; feld: feldspar; ab: albite; gr: quartz; biot: biotite; am: amphibole; pyr: pyroxene; ap: apatite; hys: hypersthene; chl: chlorite; epi: epidote; qtz: quartz; ur: uraninite; cal: calcite; am: amphibole; horn: hornblende; flu: fluorite; ur: uraninite
 a: data from samples in Table 3.1 (1983b); all other data are from Fushimi (1985)
 b: calculation of $\delta^{18}O$ of the fluid phase was done using magnetite-water fractionation, defined by the equation:

$$1000 \ln \alpha_{\text{H}_2\text{O}} = -1.07 \left(\frac{T}{1000} \right)^2 - 3.70 \left(\frac{T}{1000} \right) + 3.00$$

for samples which show mineral-mineral fractionations that are at least almost doubly or triply concordant. Temperatures obtained by the feldspar-magnetite pairs are used in the calculation in accordance with the metamorphic evidence and the metamorphic history presented in the course of this thesis.
 c: calculation of $\delta^{18}O$ of the fluid phase was done using albite-water fractionation, defined by the equation:

$$1000 \ln \alpha_{\text{H}_2\text{O}} = 1.39 \left(\frac{T}{1000} \right)^2 - 1.16 \left(\frac{T}{1000} \right) + 1.19$$

using the temperature-mineral fractionations obtained by the quartz-feldspar pairs are used in the calculation.

d: calculations for these samples were done solely to illustrate the shift from hornblende to lighter values. The lack of other mineral pairs for 1 determined here, other than $\delta^{18}O$ fluid values difficult to interpret.

In the zone of albitization all minerals undergo a shift towards lighter values except for a few samples. In albite-gneisses, where microcline and/or quartz are still present in considerable amounts (see Table 5.9), oxygen isotope values of 1.4 and 2.2 were encountered for quartz; albite values range from 0.0 to 9.6 per mil and values for biotite and magnetite are 3.4 and 5.2, -7.2 and -7.6 respectively. In unmineralized albitites quartz values range from 3.4 to 10.3, albite from 0.4 to 8.7, magnetite from -6.7 to -0.8 per mil; one determination in biotite yielded a value of 2.2 per mil. It is important to note that the highest values for quartz and/or albite were encountered in samples rich in either microcline or quartz. In mineralized albitites quartz values range from -1.2 to 6.4 per mil, albite from -2.6 to 4.5, magnetite from -9.5 to -6.0, biotite from -3.6 to 0.3 and pyroxene from -4.1 to -0.4 per mil; once again the highest values correspond to samples containing quartz (in 10 to 30%).

5.5.2. Discussion

a) Isotopic thermometry

The use of oxygen isotope fractionations among minerals for geothermometry has been the subject of discussion by many authors (Bottlinga and Javoy 1973, 1975,

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Javoy, 1977, Deines, 1977, Dahl, 1979, Clayton, 1980). There is no consensus on which fractionation factors should be applied in temperature calculations, nor on the criteria to test for isotopic equilibrium among minerals.

Isotopic thermometry is based on three assumptions: (1) the isotope exchange reactions must have approached equilibrium; (2) the isotopic compositions were not altered subsequent to the establishment of equilibrium; (3) the temperature dependence on the fractionation factors is known from experimental determinations (Faure, 1977). Because isotopic thermometry requires the establishment and preservation of isotopic equilibrium in order to provide meaningful temperatures, concordance among the various mineral-pair fractionations is a necessary condition for assignment of isotopic temperatures (Clayton, 1980). It should also be stressed that textural equilibrium does not necessarily imply isotopic equilibrium.

The quartz-water, feldspar-water and magnetite-water fractionations presented by Bottinga and Javoy (1973) have in part been favoured in this work, because they provide internally consistent results, thus enabling easier comparisons. Recent reviews on this subject (Dahl, 1979, Clayton, 1980) recommend the use of the calibration of Matsuhisa et al (1979) for quartz-water and albite-water

fractionations and that of Becker and Clayton (1976) for magnetite-water. Dahl (1979) concluded that the equilibrium curve of Bottinga and Javoy (1973) for magnetite-water is in good agreement with that proposed by Becker and Clayton (1976). The use of the calibration curve for quartz-water by Matsuhisa et al (1979) results in considerably lower calculated temperatures (as much as -150°C), and was not chosen for that reason. The isotope fractionation equations quoted in this section are listed in Table 5.10. Although calculations have also been done for quartz-biotite, feldspar-pyroxene and feldspar-biotite pairs, the use of these results is not as adequate; first because there is a general disagreement in the available calibrations for biotite and pyroxene (Essene, 1982), and also because these minerals are more susceptible to retrograde exchange. On the other hand, quartz and magnetite are more likely to preserve their record of maximum metamorphic temperatures (Clayton, 1980).

Use of equations in Table 5.10 on oxygen isotope data for Lagoa Real, resulted in a large number of temperature determinations which are displayed in Table 5.11; some of these data are depicted in Fig. 5.16. In regional gneisses temperatures vary from 460° to 680°C and as no concordant temperatures are found among mineral pairs, their meaning is questionable. In type A "transitional"

Table 5.10—Equilibrium mineral-mineral fractionation equations.

Equation	Reference
$1000 \ln qz-pl = 0.97 \frac{10^6}{T^2} - 1.04$	Bottinga and Javoy, 1973
$1000 \ln qz-mg = 5.57 \frac{10^6}{T^2}$	Bottinga and Javoy, 1973
$1000 \ln ab-mg = 3.06 \frac{10^6}{T^2} + 2.54$	Matsuhisa et al., 1979, for albite-water fractionation; Bottinga and Javoy, 1973, for magnetite-water fractionation
$1000 \ln pl-pyr = 1.70 \frac{10^6}{T^2} - 1.04$	Bottinga and Javoy, 1975
$1000 \ln pl-bio = 2.72 \frac{10^6}{T^2} - 1.04 - 0.60$	Bottinga and Javoy, 1975
$1000 \ln qz-bio = 2.54 \frac{10^6}{T^2} + 0.85$	Shieh and Schwarcz, 1974

Table 5.12-Oxygen isotope thermometry. Temperatures are calculated for the pairs qz-fel, qz-bio, qz-mq, fel-mq, fel-bio, fel-pyr. according to mineral-mineral fractionations defined by the equations listed in Table 5.10. The data in the present table is depicted in Fig. 5.16.

	Analysis no.	Sample no.	T(°C) qz-fel	T(°C) qz-bio	T(°C) qz-mq	T(°C) fel-mq	T(°C) fel-bio	T(°C) fel-pyr
Regional gneisses	1	H8Z 728	460					
	2	H8V-894		625				
	3	H8U 626		840				
	4	H8V 763	530	460	595	680	490	
Type A "Transitional" gneisses	5	(01)	360°-380		490	530		
	6	(03)	330		470	520		
	7	H8F 581						
	8	(02)	360		480	510		
	9	H8C 251	460	440			490	
	10	H8U 326	480°-505					
	11	H8U 325	530	350			370	
	12	(06)						
	13	H8V 765	460					
	14	H8U 259						
	15	(05)	460					
	16	(04)	470		540	600		
Type B "Transitional" gneisses	17	H8Z 903			490			
	18	H8X 070	290	450			650	
	19	H8U 320	560°-590	475			480°-490	
	20	H8V 791	625°-905		570	550		670°-700
	21	H8V 892	900		580	560		

Table 5.11
(Cont.)

	Analysis no.	Sample no.	T(°C) qz-fel	T(°C) qz-bio	T(°C) qz-mq	T(°C) fel-mq	T(°C) fel-bio	T(°C) fel-pyr
Albite- gneisses	22	HBU 327					465	
	23	HBU 269						
	24	HBU 260	440		500	510		
	25	HAV 887	765(pl) 560(mic)		515	470(pl) 500(mic)		
Unmineralized Albitites	26	HAV 883	420		530	580		
	27	HBU 328						605
	28	HBU 322						
	29	HAZ 901	410		540	620		
	30	HBU 324		815	690			
	31	HBU 271	505		520	530		
	32	HBU 296				540		
	33	HAV 786				550		
Mineralized Albitites	34	HAV 896	440	420			480	
	35	HAV 773	505					
	36	HAV 787				490		605
	37	HBU 284				475	430	
	38	HBU 282				520		570
	39	HAV 895	460(pink) 360(green)		540	585(pink) 650(green)		
	40	HAV 777				540	550	
	41	M3	488		530	555		570
	42	M2	765		570	555		
	43	HAV 785				580		
	44	HBU 321				555		
	45	M1	560		550	560		

*For comparison, these temperatures were calculated assuming an anorthite-content of 10% in plagioclase in accordance with microprobe results; calculations are done only for samples where plagioclase percentage surpasses that of K-feldspar.

Fig. 5.16. Isotopic temperatures calculated using the $\delta^{18}\text{O}$ data on Table 5.9, for the pairs quartz-feldspar, quartz-magnetite and feldspar-magnetite. Equations are from Table 5.10, and calculated temperatures on Table 5.11.

rocks almost doubly concordant temperatures are found in analyses nos. 5 and 8 (480-530°C, for qz-mg and fel-mg pairs). Almost triply concordant temperatures in analysis no. 9 are found utilizing quartz-biotite and feldspar-biotite pairs; such temperatures (440-490°C) are in reasonable agreement with those mentioned above (that is, 480-530°C). In type B "transitional" rocks doubly concordant temperatures of 550-580°C are observed for analyses nos. 20 and 21 (for qz-mg and fel-mg pairs).

Also in the albitized rocks, concordant temperatures among mineral pairs are not widespread though more common than in the "unaltered" rocks. Among the albite-gneisses, double concordance is observed for analysis no. 24 (for qz-mg and fel-mg pairs) and almost triple concordance for no. 25 (for qz-fel, qz-mg and fel-mg pairs); temperatures range from 550° to 560°C. Temperatures in the range of 505° to 580°C are obtained for the group of unmineralized albitites; temperatures in analysis no. 26 (for qz-mg and fel-mg pairs) are doubly concordant and those in analysis no. 31 (for qz-fel, qz-mg and fel-mg pairs) are triply concordant. In mineralized albitites, double concordance in analyses nos. 39, 41, 42 (qz-mg and fel-mg pairs) and 40 (fel-mg and fel-bio pairs), and triple concordance in analysis no. 45 (qz-fel, qz-mg and fel-mg pairs) are observed, indicating temperatures in the range

of 540-570°C.

From the above observations it is evident that the triplet quartz- feldspar-magnetite provides the least disturbed evidence for isotopic equilibrium.

b) Fluid isotopic signature

Isotopic compositions for the fluids in isotopic equilibrium with the different mineral assemblages were calculated for samples which showed reasonable concordant temperatures implying isotopic equilibrium, except in a few cases (see Table 5.9 , superscript d). In type A "transitional" rocks the fluid composition varies from 5.8 to 8.5 per mil; in type B "transitional" rocks, it varies from 5.8 to 6.8 per mil. On the basis of oxygen isotope composition alone, according to Taylor's (1974, 1979) classification of the various "terrestrial water reservoirs", such values (5.8 to 8.5 per mil) are compatible with a metamorphic origin of the fluids.

In the zone of albitization fluid isotopic signatures from -1.4 to 1.1 per mil are found for the albite-gneisses. Among the unmineralized albitites, oxygen isotope values of -0.8 to 7.3 per mil were encountered for the fluid phase, whereas in the mineralized albitites these values range from 2.6 to -3.7 per mil. Some of

these very light values (0.0 to - 3.7 per mil) are compatible with a signature for meteoric water (Taylor, 1974, 1979) or for pore water of marine sediments (Savin and Yeh, 1981). See Table 5.9.

Fluid composition calculations suffer from the same limitations as those constraining isotopic temperature calculations. Nevertheless, one can speculate that the shift from heavier to lighter isotopic ratios for the metasomatic rocks results from exchange with a fluid isotopically light (as in meteoric waters).

The gradual transition towards lighter values can be explained by progressive exchange with a ^{18}O -depleted fluid under varying water/rock ratios. Accordingly, $\delta^{18}\text{O}$ of the fluid calculated from "transitional" rocks represent a low water/rock ratios, and $\delta^{18}\text{O}$ values of the fluid calculated for the metasomatic rocks (lightest values) represent higher water/rock ratios. The latter values (as low as -3.7 per mil) are probably the closest to the real composition of the fluid. This explanation is readily applicable within the albitized rock groups; heavier values found for analysis no. 26 (7.3 per mil), no. 35 (2.6 per mil) and no. 34 (1.5 per mil) probably represent interaction with the metasomatic fluid in a rock dominated system. It is also very important to point out that these high fluid oxygen isotope values refer to

samples where microcline and/or quartz are still present in considerable amounts; this mineralogical feature indicates incomplete metasomatic alteration as the disappearance of either or both minerals can be used as a tracer for "completion" of the alteration event, as described in other parts of this thesis. Accordingly, more evolved (meaning more altered) albitites yield lighter isotopic fluids.

One could alternatively explain the varying isotopic ratios of the fluid phase by invoking the participation of unrelated fluids of different isotopic compositions characteristic of each alteration stage. However, the trend of alteration described for the Lagoa Real rocks (see Chapter 6), which involves among other things, continuous oxidation (from type B "transitional" rocks to albitites) to the limit of hematite formation and uraninite precipitation, and albitization, makes such speculation difficult to envisage. Based on the sequential development of the alteration phenomena, it can be argued that changes in $\delta^{18}\text{O}$ of the fluid result from isotopic exchange reactions of a single fluid (evolving fluid) at different water/rock ratios. Accordingly, each alteration stage may reflect a specific range of water/rock ratios.

5.5.3. Concluding remarks

Despite the uncertainties in the calculation of meaningful isotopic temperatures, the relative constancy of fractionations among minerals (mainly quartz, feldspar and magnetite) suggests isotopic equilibration for all rock types under the same temperature range. On the other hand, the shift towards lighter values in the more altered rocks implies isotopic exchange with a fluid isotopically lighter. One can invoke two possibilities in order to satisfy this interpretation:

1/ metasomatism proceeded under essentially the same temperature range as the regional metamorphism, though involving fluids with a lighter isotopic signature, under varying water/rock ratios;

2/ metasomatism proceeded at temperatures that were lower than those in effect during regional metamorphism in the presence of an isotopically lighter fluid, under varying water/rock ratios; subsequent thermal reequilibration accompanied by isotopic reequilibration, reset the mineral-mineral fractionations towards lower values.

Considering that the textural attributes of both host mineral assemblages (in "transitional" rocks) and metasomatic assemblages are compatible with a single metamorphic event, and that the isotopic reequilibration

under prograde conditions is easily attainable, the second hypothesis seems to be adequate. Nevertheless, the first hypothesis cannot be ruled out entirely.

CHAPTER 6

METAMORPHISM

6.1. Sequence of Metamorphic Reactions

6.1.1. "Transitional" rocks

Gneissic rocks (regional rocks), mostly of granitic composition, are locally intensively recrystallized and foliated. They are very coarse grained porphyroclastic rocks containing k-feldspar, quartz, oligoclase or albite-oligoclase, amphibole (hastingsite and Fe-pargasite), ilmenite and magnetite. Progressive deformation resulted in the development of foliated, fine to medium grained gneisses basically containing the same mineralogy (type A "transitional" rocks) as the regional rocks. In places, the textural reorganization is accompanied by some mineral transformations which include, the neoformation of albite in the form of rims around pre-existing plagioclase and the partial or almost complete replacement of amphibole by garnet and/or

pyroxene (type B).

A summary of the mineral assemblages includes:

In regional rocks:

orthoclase-quartz-plagioclase-hastingsite

orthoclase-quartz-plagioclase-hastingsite-biotite

In type A "transitional" rocks:

microcline-quartz-plagioclase-hastingsite

microcline-quartz-plagioclase-hastingsite-biotite

In type B "transitional" rocks:

albite-quartz-hasting --> pyrox + magnet

microcline-quartz-oligoclase-hasting --> pyrox

microcline-oligoclase-quartz-hasting --> garnet + magnet

microcline-oligoclase-quartz-hasting --> magnet + fluorite

oligoclase-quartz-microcline-hastingsite --> pyrox +

garnet + magnet + fluorite

oligoclase-quartz-microcline-hastingsite --> garnet +

pyrox + fluorite

In type B "transitional" rocks microcline characteristically shows textural evidence for its incomplete substitution by plagioclase. Vermiform and granular epidote is a secondary alteration after amphibole, garnet and pyroxene.

Microprobe analyses show that amphiboles in both regional and type A "transitional" rocks are predominantly hastingsite and subordinate ferropargasite, both with very high $\text{Fe}^{\text{t}}/\text{Mg}$ ratios (15 to 60 and 44 to 88, respectively). The $\text{Fe}^{\text{t}}/\text{Mg}$ ratio is considerably reduced (3 to 29, but average ~10), and Fe^{+3} predominates, once amphibole starts to be replaced by pyroxene and/or garnet (see Figs. 4.1 and 4.4).

Garnets in type B "transitional" rocks are members of the ugrandite series with andradite varying from 37.86 to 58.36% and grossular from 38.24 to 55.80%, with low to absent spessartine and pyrope end-members. Pyroxenes are members of the hedenbergite-diopside series and plot as ferrosalites and end-member hedenbergite (see Fig. 4.8).

Thomas (1982) investigated the upper thermal stability of hastingsite as a function of temperature, fluid pressure and oxygen fugacity, using the ideal hastingsite composition $\text{NaCa}_2\text{Fe}_4^{+2}\text{Fe}^{+3}\text{Si}_6\text{Al}_2\text{O}_{22}(\text{OH})_2$. His experiments showed that under $f\text{O}_2$ conditions

controlled by FMQ buffer, hastingsite dehydrates to produce the assemblage hedenbergite, magnetite, an andradite-grossular garnet, plagioclase plus fluid. Similar results were obtained for the hornblende ferropargasite (Gilbert, 1966), except that due to the absence of ferric iron, the stability of this amphibole is reduced when compared to hastingsite. Both workers concluded that in the presence of excess silica at oxidation states defined by the FMQ buffer, the high temperature assemblage (dehydrated) field is enlarged; it should also be noted that the dehydration curve for the pure amphibole phase in a natural rock system, would be lowered in temperature due to the presence of additional components.

The alteration assemblage encountered in type B "transitional" rocks is in good agreement with those of experimental studies. Whether plagioclase is part of the alteration assemblage as indicated by Thomas (1982), is unclear; its presence in the original assemblage renders difficult the identification of neoformed plagioclase.

As shown above, not all mineralogical phases are present in the alteration assemblage in each case. Garnet for instance, tends to be less frequent and pyroxene in many cases, is the most abundant phase (Thomas, 1982). A continuous oxidation pattern is manifested by the presence

of zoned pyroxene crystals in which the borders are enriched in Na and Fe^{+3} , compared to cores.

In type A "transitional" rocks, ilmenite was only found in one sample and in type B, only magnetite is present. In both types, sphene is rather abundant as an accessory mineral and is frequently associated with magnetite. Sphenitization of ilmenite or titanomagnetite is relatively common; for sphene formation, the additive ions are Ca^{+2} and SiO_2 . Haggerty (1976) has shown that at high SiO_2 activity, magnetite and sphene will form at the expense of ilmenite, at oxygen fugacities between the FMQ and NNO buffers.

In summary, the first mineral transformations are:

6.1.2. Metasomatic rocks

In places where recrystallization is complete together with total or almost total changes in the compositions of the minerals, the rocks formed are fine to medium polygonal granoblastic albitites characterized by the following assemblage:

albite-aegirine augite-andradite-magnetite

A group of Ca-rich albitites may also contain calcite, wollastonite and very rarely, vesuvianite; in addition, their pyroxene is richer in MgO and garnet contains higher amounts of the grossular end-member. In all mineralized albitites, uraninite is present mostly associated with mafic minerals, and magnetite is invariably altered to hematite.

A group of metasomatic rocks which are not albite bearing, are the epidiosites. Epidiosites are usually mineralized, and contain oligoclase, quartz and epidote replacing pyroxene (aegirine-augite).

a) Albitites

Textural evidence for the substitution of k-feldspar by albite is widespread in most samples. Na addition is confirmed by mass balance studies (Lobato et al, 1982; see Chapter 5). The main reaction involving Na addition

is:

Quartz is almost totally removed from the original assemblage and, at least at the initial stages, is being consumed by amphibole to produce Fe-salite/hedenbergite and grossular-andradite garnet in type B "transitional" rocks.

Besides Na enrichment in albitites, O_2 addition is also very important, and that is shown by comparing the mineralogy in albitites with that of "transitional" rocks. The aegirine content of pyroxenes increases (8-35%, in one exceptional case reaching 80%), indicating continuous oxidation coupled with Na enrichment (see Fig. 4.9). Also the zoning in pyroxenes confirms this same trend. These changes are summarized by the reaction:

The variable chemistry in clinopyroxene composition, from very FeO-rich (up to ~25% FeO total) to sometimes moderately rich in MgO (~10%, reflected in the Enstatite end-member %), probably reflects original bulk composition which can be illustrated by the variable $\text{Fe}^{\text{t}}/\text{Mg}$ ratios of

However, the possible (?) metamict characteristic of a number of grains and the doubtful inclusions of uraninite (many times uraninite grains look as though they are in fact associated with neighbouring crystals), point out to a late origin.

The metasomatic event in Ca rich albitites is reflected in the formation of albite (possibly at the expense of original Ca-plagioclase) and by the enrichment in the aegirine molecule in pyroxene; pyroxene of common calc-silicate rocks is diopside. The oxidation reaction pyroxene $\rightarrow$ andradite is widespread as well, but due to lower $\text{Fe}^{\text{t}}/\text{Mg}$ ratios and higher CaO percentages in their pyroxenes (when compared to other albitites), garnets formed are richer in the grossular molecule (see Fig. 4.5). Also probably due to lower $\text{Fe}^{\text{t}}/\text{Mg}$ ratios, magnetite does not seem to occur as a result of that reaction.

original amphiboles (Fig. 4.1).

As alteration proceeds, the garnet evolved from reaction (1) is probably replaced by an even more andradite rich phase due to the oxidizing trend. However in thin section, it is difficult to distinguish garnets thus formed from those produced as a result of reaction (5), discussed below.

In the Lagoa Real albitites, textural evidence points out to the replacement of aegirine-augite by an andradite garnet. Gustafson (1974) studied the stability relations of hedenbergite and andradite of common iron-rich calc-silicate skarns. At 2000 bars P_{fluid} , hedenbergite + O_2 react to form the anhydrous oxidized assemblage andradite + quartz + magnetite, according to the reaction:

The reaction takes place at oxygen fugacities somewhat above those defined by the NNO buffer. This reaction is common in a number of deposits associated with skarns (Sato, 1980, Sivaprakash, 1981, Burton et al, 1982).

As pointed out by Einaudi (1982), natural assemblages in most cases do not contain phases of end-member composition. Even though as a first approximation, garnets can be treated as pure end-member andradite, the magnesium content of clinopyroxene must be accounted for.

He concluded that the overall composition of salite reflects an approach to equilibrium with andradite at oxidation states intermediate between the NNO and MH buffers. Gamble's (1978, in Burton et al, 1982) experiments indicate that Mg contents in excess of 50 mole percent diopside are necessary to effect a change in the stability of hedenbergite. On the other hand, only 15% johannsenite raised the stability field of hedenbergite by 1.0 log fO_2 at temperatures between 700° and 800° C (Chou, 1978).

In the case of aegirine-augites in the albitites, the high acmite content should considerably raise the stability field of clinopyroxene explaining why both minerals are invariably found together. As in almost all samples where such reaction is observed, magnetite appears altering to hematite at varying degrees, it is suggested that the reaction pyroxene $\rightarrow$ andradite occurs under conditions equal to or more oxidizing than those defined by the hematite-magnetite equilibrium.

Sphenitization is also very common in most albitites and is related to magnetite oxidation (which is titanium rich; ~9% TiO_2) of both generations; that is, either primary or related to pyroxene $\rightarrow$ andradite reaction.

In summary:

aegirine-augite

where Fe_2TiO_4 is TiO_2 rich magnetite

b) Ca-rich albitites

In Ca-rich albitites, the above reactions are basically the same and are being controlled by bulk original composition. Considering the chemical nature of these rocks (see Chapter 5), it is possible that they were originally calc-silicate lenses, commonly found in this terrain. Petrographic studies alone, did not permit to evaluate the relation between the "common" (pyroxene, andradite and albite) albitite minerals with others, typical of Ca-rich assemblages. Nevertheless in thin section, wollastonite, pyroxene and also calcite seem to coexist in textural equilibrium. This could imply that at least wollastonite, pyroxene, calcite and perhaps vesuvianite were already part of the original assemblage,

which was stable but recrystallized during the "albitization".

Wollastonite, always intergrown with calcite, could probably have been the result of the reaction:

Whether its partial replacement by pectolite (~ 8.5% Na₂O) is related to Na metasomatism, is unclear. Pectolite occurs as fine fibres replacing borders of wollastonite, unrecrystallized. Thilo and Funk (1950, in Deer et al, 1978) reported that pectolite loses water above 400°C and changes to a mixture of wollastonite, SiO₂ and Ca₂NaHSi₃O₉. In spite of its textural character, the high Na content and the fact that its formation is reportedly favoured by a silica-saturated environment (Juan et al, 1968, in Deer et al, 1978), suggest that pectolite is part of the metasomatic assemblage, even if at lower temperatures.

Vesuvianite is only observed in two samples, never in association with wollastonite, and its origin is uncertain. As pointed out by Kerrick et al (1973), vesuvianite is commonly the result of the following:

However, the possible (?) metamict characteristic of a number of grains and the doubtful inclusions of uraninite (many times uraninite grains look as though they are in fact associated with neighbouring crystals), point out to a late origin.

The metasomatic event in Ca rich albitites is reflected in the formation of albite (possibly at the expense of original Ca-plagioclase) and by the enrichment in the acmite molecule in pyroxene; pyroxene of common calc-silicate rocks is diopside. The oxidation reaction pyroxene $\rightarrow$ andradite is widespread as well, but due to lower $\text{Fe}^{\text{t}}/\text{Mg}$ ratios and higher CaO percentages in their pyroxenes (when compared to other albitites), garnets formed are richer in the grossular molecule (see Fig. 4.5). Also probably due to lower $\text{Fe}^{\text{t}}/\text{Mg}$ ratios, magnetite does not seem to occur as a result of that reaction.

c) Epidosites

Epidosites have a restricted distribution in the area and proper field relations with other lithologies are obscure. For the present study, only 5 samples were available. Their chemistry points to a Al-Ca-Fe rich bulk composition, again suggesting a calc-silicate precursor. They contain a rather Ca-rich plagioclase (An% up to 18%), quartz, diopsidic pyroxene (enriched in acmite), epidote (containing up to 1% UO_2 , in mineralized samples), sphene and rare uraninite.

Fyfe et al (1958) suggested that in rocks coupled reactions may influence the crystallization of epidote, and that its synthesis from component oxides will be more difficult than its formation by the reaction anorthite + pyroxene + water = epidote + chlorite + quartz:

In epidosites, the replacement of pyroxenes by epidote could be the result of the reaction with anorthite component of original plagioclase during albitization, as follows:

d) Carbonate alteration and other alterations

In a few samples studied, calcite appears forming at the expense of andradite. Taylor and Liou (1978) investigated the low-temperature stability of andradite as a function of X_{CO_2} and fO_2 at constant P_{fluid} of 2000 bars. Their results indicate that the reaction:

occurs at $T=550^\circ C$ at $X_{CO_2} = 0.22$ on the HM buffer. As carbonate alteration in metasomatic rocks is not widespread, it is suggested that CO_2 pressures locally attained sufficient levels so that carbonate crystallized.

A number of secondary minerals are found throughout the samples, in "transitional" and all metasomatic rocks. These include: epidote (from pyroxene, garnet and amphibole); amphibole (from pyroxene); very rare biotite and/or chlorite (in two samples, only); prehnite after pyroxene (in one sample only). Specifically in the cases of epidote and amphibole, both minerals always occur very fine grained, vermiform and interstitial to others, never

exceeding 5%. Such textural observations suggest that this assemblage is of a later generation in relation to well recrystallized minerals like plagioclase, garnet and pyroxene. Though not commonly observed, uraninite can appear as inclusions in amphibole, but not inside prehnite, for instance.

e) Quartz and microcline

Albitites almost never contain much quartz or microcline. It has been pointed out that they usually occur in no more than about 5%, interstitial to other minerals. Microcline clearly appears to be consumed by plagioclase in most samples.

In very few albitites either one of these minerals can occur in up to about 15 or 20%. If occurring recrystallized and with same grain size as all other minerals in the rock suggesting textural equilibrium, this probably indicates that either quartz or microcline have not reacted out, and are metastably coexisting with other mineral phases. On the other hand, a few albitites studied contain quartz that either form micro veins or appear in concentrations; this quartz is always coarser grained than other minerals in the rock, xenoblastic and highly strained. This is indicative of a late origin for

quartz, which does in fact appear as late veins in some localities in the area.

6.2. Fluid inclusion studies

Fluid inclusion studies were carried out by Fuzikawa (1980, 1982), Fuzikawa e Alves (1984) /Alves e Fuzikawa (1984). Their results are summarized bellow:

- preliminary results obtained from the study of fluid inclusions in albites from mineralized and unmineralized albitites indicate pressures of at least 1 kbar. Salinity ranges from ~1.5 to 23% (NaCl equivalent). The presence of dissolved salts of K, Ca and Mg were inferred, and in one sample CO₂ and CH₄ occur;

- fluid inclusions were also studied in carbonates, from mineralized albitites (1), carbonate veins and breccias cross-cutting the mineralized zones (2) and carbonate rich zones away from the mineralization (3). Fluids very rich in CO₂ and CH₄ (and other hydrocarbons) are present in carbonates in (1) and (2); salinities are in the order of 15 to 21% (NaCl equivalent). In (3), the fluid inclusions were essentially aqueous, with very high salinities (~20% NaCl equivalent). δ Cl³⁵ isotope determinations in (1) (between -4.75 and - 7.11) favour a

crustal origin for the carbonates;

- recent results in mineralized and unmineralized albitites confirm earlier studies. Inclusions in albitites from an unmineralized albitite indicate salinities of 15 and 23.5% NaCl (equivalent). Inclusions in albitites from mineralized albitites show salinities >23% NaCl, and are rich in CO₂ and hydrocarbons.

6.3. Conditions of metamorphism

Unfortunately, precise estimates of pressure and temperature conditions are rendered difficult because the dominant mineral assemblage (aegirine-augite, andradite, albite, magnetite) is stable over a wide P-T range. Nevertheless, based on mineral reactions and oxygen isotope data, a certain number of metamorphic constraints can be traced.

Isotopic thermometry was undertaken (see Chapter 5) and temperatures were obtained using quartz-albite, quartz-magnetite, and albite-magnetite pairs. In spite of the many problems involved in using isotopic temperatures as geothermometers for metamorphic rocks (see Clayton, 1979 and Essene, 1982), the isotopic temperatures (500-600°C) will be used as the best approximation.

Constraints on pressure are sparse. The lack of jadeite component in aegirine-augites (jadeite component of ~8%, is present only in hedenbergites in type B "transitional" rocks) rules out high pressure metamorphism (Newton and Wyfe, 1976). Popp and Gilbert (1972, see also Gasparik and Lindsley, 1980) showed that the solubility of jadeite in the Fe^{+3} Al-clinopyroxenes decreases with pressure, and that at 4 kbar the maximum solubility of jadeite component in aegirine decreased from 5-6 mol % at 500°C to 4-5 mol % at 600°C; jadeite should be more soluble in aegirine in silica-undersaturated environments. Lobato et al (1983b) also pointed out that the presence of kyanite coexisting with muscovite in metarhyolites of the lower Espinhaço supergroup near the gneissic terrain border, implies pressures of at least 4 kbar at temperatures below 650°C (Holdaway, 1971, Turner, 1981; see discussion in Chapter 8). Therefore, a tentative estimate of slightly less than 4 kbar for the prevailing pressure is proposed.

6.3.1. "Transitional" rocks

Although the mineral transformations observed in type B "transitional" rocks are in good agreement with Thomas' (1982) experimental results, their direct application is limited and should be done with caution. As already

mentioned, that author found that hastingsite dehydrates to form hedenbergite + andradite/grossular + plagioclase + magnetite + fluid at the FMQ buffer. At temperatures around 550°C and $P_f = P_t = 4$ kbar, this reaction would not proceed and hastingsite would still be stable. The presence of excess SiO_2 in the system would decrease the amphibole stability field, but the magnitude of this deviation is difficult to evaluate. Under conditions of NNO buffer, Thomas (1982) concluded that hastingsite stability is considerably lowered, though no conclusive demonstration of a stability field was obtained.

The onset of the alteration is marked by metamorphic conditions characterized by low water/rock ratios, restricted albitization of original feldspars, widespread recrystallization, and oxidation/dehydration of original hastingsite by hot, oxidizing fluids whose $\log f\text{O}_2$ are probably in the range of -23 to -21, at 550°C (from buffer assemblages by Eugster and Wones, 1962, Wones and Gilbert, 1969). As pointed out by Giggenbach (1984), at very low water/rock mass ratios, the hydrothermal alteration process corresponds to isochemical recrystallization of the primary rock; a similar process could have been dominant. Restricted desilicification is demonstrated by the dehydration reaction of amphibole; as silica activity in the fluid increases due to quartz dissolution, the

stability of amphibole is lowered enhancing its dehydration (see Gilbert et al, 1982, for discussion p. 230). In addition, Fe^{+2} rich amphiboles have a lower stability when compared with Mg-rich phases (Gilbert et al, 1982). It should also be noted that the activity of water is lowered in saline solutions (Garrels and Christ, 1965), thus enhancing amphibole dehydration reaction.

6.3.2 Metasomatic rocks

Continuous oxidizing trends are demonstrated by the mineralogy. Replacement of aegirine-augite by andradite is accompanied in most samples, by the alteration of magnetite to hematite, suggesting that $\log f\text{O}_2$ is in the order of -17, at 550°C .

Because wollastonite is stable with calcite in a number of albitites, equilibrium metamorphic fluids must have been characterized by an X_{CO_2} defined by the reaction: calcite + quartz = wollastonite + CO_2 , at 550°C . For this reaction, at $P_f = P_t = 2000$ bar, on the HM buffer, X_{CO_2} (mole fraction) values are in the range of about 0.10 to 0.15 (see Miyashiro, 1975, Winkler, 1976, Kerrick et al, 1973, Taylor and Liou, 1978). Andradite replacement by calcite is localized; as mentioned in section 6.1.2., reaction (9) occurs at $T = 550^\circ\text{C}$, at $X_{\text{CO}_2} = 0.22$ on the HM buffer (at constant $P_f = P_t$ of 2000

bars). These observations suffice to suggest that during the alteration, CO_2 pressures were moderately low.

Metamorphism of the original rocks probably controlled fluid composition. In section 6.1.2., the mineral reactions listed as (1), (3), (4), (5) and (6) all consume oxygen, reactions (1), (3) and (6) consume silica, and reactions (5) and (9) liberate silica; the alteration of pyroxene to amphibole also liberates silica. This implies that SiO_2 is being released to the fluid at the earlier stages of the alteration; oxidation is widespread throughout all stages and $f\text{O}_2$ of the fluid phase should decrease towards the late stages of alteration; SiO_2 addition to the rocks in later stages is expressed by reactions (5) and (9) and by the presence of late quartz veins. Varying salinity values (from 1 to 25% NaCl equivalent), may suggest that whereas salinity of the fluid was high in early stages of alteration when albite was forming, it became lower in later stages.

It should be mentioned that quartz solubility increases with increasing temperature at all moderate pressures and that the solubility of quartz is not strongly influenced by dissolved salts (Fyfe et al, 1978).

6.3.3. The role of temperature

As mentioned earlier (6.1.2. d), in late stages of alteration, secondary minerals replace pyroxene and/or garnet in albitites. These include amphibole (by far the most abundant), epidote, rare biotite and even rarer chlorite and prehnite. Uraninite can occur as inclusions in amphibole.

This late alteration may represent a shift towards conditions compatible with quartz formation (and stability), as a response to temperature decrease (see Chapter 8); such conditions assist amphibole stability, and it starts to replace pyroxene. In addition, the low Na content of these amphiboles is indicative of low Na activity in the fluid phase in the late stages of metasomatic alteration; sodium impoverishment in the fluid results from widespread formation of albite during early stages of alteration.

The restricted formation of prehnite replacing pyroxene (probably reacting with plagioclase), again lends support for the proposed decrease in temperatures during the late stages. Although prehnite can be stable at relatively high temperatures (Coombs et al, 1959, Liou, 1971), the reaction: hedenbergite + anorthite + fluid = prehnite + magnetite, occurs at lower T and fO_2 than

the hedenbergite --> andradite replacement (Liou, 1973); see Fig. 6.1.

6.3.4. Epidosites

The assemblage, in epidosites (generally mineralized) deviates from the common albitite assemblage. The assemblage is made up of: epidote (Ps 28), oligoclase, aegirine-augite (rich in MgO), quartz, where epidote (and not andradite) appears replacing pyroxene according to reaction (8), section 6.1.2.

The presence of relatively abundant quartz (5 to 35%), and the high anorthite content of plagioclase (An 18%), is indicative of limited interaction of these rocks with the metasomatic fluids. That, and their bulk composition (Al-Ca-Fe-rich) represent key factors for controlling epidote formation as opposed to andradite. Liou (1973) showed that at 5 kbar $P_f = P_t =$ pressure, epidote is iron-rich (Ps 33 $\pm$ 2) at high (HM and CCO) oxygen fugacities, and becomes progressively more aluminous (Ps 25 $\pm$ 3) with decreasing fO_2 values and temperatures (defined by the FMQ buffer). With fO_2 defined by the HM buffer, epidote (Ps 33) is stable up to 748°C, 5 kbar, 678°C, 3 kbar and 635°C, 2 kbar $P_f = P_t$. Under reducing conditions (slightly lower than the FMQ buffer), epidote is not stable at any temperatures and

epidote + quartz gives place to hedenbergite + anorthite + fluid in SiO_2 -rich rocks.

As pyroxenes in epidiosites are Fe^{+3} rich, pyroxene to epidote replacement must have occurred at high $f\text{O}_2$, possibly defined by the magnetite-hematite buffer, in accordance with pyroxene replacement in albitites.

Accordingly, epidiosites may represent a suite of calcium-iron-aluminum rich rocks altered under conditions of low water/rock ratios with the reaction aegir-aug + anorth component + fluid = epidote + quartz, being controlled by bulk composition (availability of Al), under $f\text{O}_2$ defined by the magnetite-hematite pair.

6.4. Final remarks

The metasomatic alteration in the Lagoa Real rocks can be tentatively divided in the following stages:

(I)

- intense recrystallization and dehydration/oxydation of hastingsite producing hedenbergite + garnet (gross/andrad) (with SiO_2 consumption);

- restricted Na enrichment (albitization of feldspars) and quartz removal;

- sphene formation at the expense of ilmenite and magnetite;

- oxygen fugacity of the fluid controlled by slightly higher values than those defined by the FMQ buffer;

(II)

- intense recrystallization accompanied by Na enrichment expressed by the formation of albite replacing all original feldspars;

- Na and O₂ enrichment in hedenbergite giving place to zoned aegirine-augite;

- sphene formation at the expense of magnetite;

- desilicification is completed (quartz is absent);

- limited uraninite precipitation with uraninite forming inclusions in albite and pyroxene;

(III)

- oxidation of aegirine-augite (to produce andradite) and of magnetite (to produce hematite);

- widespread uraninite precipitation with abundant uraninite grains included in andradite and oxides;

- oxygen fugacity of the fluid controlled by the HM buffer;

(IV)

- incipient to moderate alteration of pyroxenes mainly to amphiboles - reestablishment of quartz stability probably due to decreasing temperatures (supported by the appearance of prehnite);

- formation of calcite from mafic minerals, and of calcite and quartz as late veins;

- very limited uraninite precipitation during amphibole formation;

- possible lowering of salinities (low salinity values are supported by fluid inclusion studies), due to early widespread albite formation.

CHAPTER 7

THE URANIUM MINERALIZATION

7.1 The ore deposit

The Lagoa Real uranium province is part of a project being studied and developed by NUCLEBRÁS - Empresas Nucleares Brasileiras, S. A. The project covers an area of approximately 4540 km² being limited by the coordinates 42° 00' - 42° 30' W and 13° 30' - 14° 15' S (Fig. 7.1), located 50 km northeast from the town Caetité in the central southern part of Bahia State, Brazil. The ore deposits are distributed along 33 km of extension from north to south, forming an arcuate zone of 32 anomalies, from which 12 are the object of research and development at the moment. General descriptions of the ore deposits can be found in Geisel et al (1980), Stein et al (1980), Villaca e Hashizume (1982), Raposo e Matos (1982), Ribeiro et al (1984), Brito et al (1984), Raposo et al (1984).

Fig. 7.1. Simplified geological map of the Lagoa Real area showing the distribution of the various uranium deposits ("jazidas", An 03, 10 etc.) in the Paramirim basement terrain. Delineated in black are the albitite bodies.

- (1) TQ : Tertiary/quaternary cover
- (2) Pgr2 : Fine granite bodies
- (3) Pch : Charnoquites
- (4) : Strike and dip
- (5) Pab : Albitites
- (6) Pgrst: São Timóteo granite
- (7) Plb : Ibitira-Brumado metamorphic complex
- (8) : Inferred faults
- (9) Pplgn: Finely foliated plagioclase gneissés
- (10) Pogn : Plagioclase-microcline-gneisses
- (11) Amg : Migmatites
- (12) : Roads
- (13) Pgn : Plag-microcl foliated gneisses
- (14) Pgno : Microcline-plagioclase-gneisses
- (15) : Ore deposits/U anomalies
- (16) : Towns

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- (13)
(14)
(15)
(16)

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The U mineralization is associated with discontinuous long tabular bodies of albitites, which trend $N40^{\circ}E$ to $N30^{\circ}W$ and dip 35° to 90° to SW or NW, with the exception of the two northernmost deposits which dip to the east and those situated in the central part of the region which are almost vertical. Each mineralized albitite body may vary up to one kilometre in length, averaging 6-7 m in width (maximum 30 m), with mineralization extending to 850 m below surface.

The contacts between albitites and hosting regional or "transitional" rocks are generally transitional but sharp contacts are also commonly found. Drilling has shown that each body of albitite consists of "bands" or "zones" of variable mineralogy, the contacts between the zones being transitional. Fig. 7.2 displays a geological section of drill hole BR01 anomaly 05 illustrating the distribution of such zones at depth; the samples from this drill hole were used during the course of this thesis.

The mineralized albitites are white to greyish-white, equigranular, fine to coarse grained, rarely porphyroclastic, usually slightly foliated. The most common types of mineralized albitites are andradite-aegirine augite-albitites and hematite-magnetite-albitites.

Fig. 7.2. Geological section of drill hole BR01 anomaly 05 (see Fig. 7.1.), illustrating the distribution of albitite bands or zones at depth. Reproduced with permission of NUCLEBRÁS-Empresas Nucleares Brasileiras S.A.

AB: albitite

AT: type A "transitional" gneiss

BT: type B "transitional" gneiss

GA: garnet-albitite

GPA: garnet-pyroxene-albitite

PA: pyroxene-albitite

MA: magnetite-albitite

In black: uranium mineralized levels

The mineralized albitites are normally discontinuous, mainly along the foliation. Each albitite body may contain one or more mineralized levels which may be interrupted in places. The contacts between mineralized levels with adjoining lithologies is generally sharp and abrupt.

Inverse faults with movements from east to west are parallel to the general foliation. Incipient to intense folding is observed. Boudinage and shearing along the foliation are also common.

According to Hoffman (1982, in Brito et al, 1984), the folding style may represent the main control for the uraniferous mineralization, at least in two ore zones (anomaly 03 - Rabicha Ore Zone, and anomaly 13 - Cachoeira Ore Zone). The lenticular mineralized bodies are probably located along the limbs of large isoclinal folds; the mineralized levels are best developed at the interception of two directions of foliation (primary and secondary) or may represent axial planes of secondary folds. Brito et al (1984) admit that a structural control for the mineralization is possible, but point out that it is combined with a lithological control. The disseminated mineralized levels seem to be confined along the foliation in the form of ore shoots whose pitch is developed in the direction of the mineral-lineation: to the NE in the

north and SW in the south. The cataclastic foliation (secondary) seems to regulate or condition zones of uranium enrichment. It has been observed in thin section (see Chapter 3), that albite grains usually show a preferred orientation; although unconfirmed, it is expected that such orientation makes an angle with the original foliation in the region.

At present the geological reserve is 93,190 metric tons (tonnes) of U_3O_8 , 2/3 of which are measured and indicated and 1/3 is inferred, contained in ore of average grade 0.15% U_3O_8 .

7.2. Alteration in shear zones

Deformation can produce changes in the structure and texture of a rock, as well as changes in the composition of the minerals or in the mineral assemblages. Bulk compositional changes may also occur by addition or loss of certain components.

Work on shear zones (Ramsay and Graham, 1970) has shown that they are commonly narrow zones of high deformation, and that their deformation state can be interpreted as the result of progressive simple shear. Within the shear zones, the rock develops new mineral fabrics and changes in mineralogy take place. In deformed

crystalline basement rocks which have been deformed under metamorphic conditions of greenschist, blueschist, amphibolite or granulite facies, ductile shear zones are extremely common (Beach, 1980). Metamorphism in shear zones is often retrogressive and the mineralogy of the rock usually shows the characteristics of the metamorphic facies under which it developed (Ramsay, 1980).

A simple observational account of the mineralogy of a shear zone is often sufficient to indicate that a fluid phase was involved in the metamorphism (Beach, 1976 and 1976). New mineral assemblages form as a result of fluid-rock interaction. The importance of shear zones is mostly clear where these structures are mineralized (Kerrick et al, 1977, Kerrich and Allison, 1978).

High deformation increases the internal strain energy within crystals promoting recrystallization leading to grain size reduction (Brodie, 1980). Variations in mineral chemistry and texture across a shear zone in phlogopite peridotite for example, have demonstrated that the degree of recrystallization of the rock increases progressively with the increasing shear strain into the shear zone (Brodie, 1980).

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In recent years, the study of shear zones has received a lot of attention, whether directed to structural (Beach, 1973, 1974, Wilson, 1980), mineralogical (Burns, 1966, Sills, 1983) or chemical (Beach and Tarney, 1978, Floyd and Winchester, 1983) aspects.

The style of deformation in shear zones established by Ramsay and Craham (1970) in the form of simple shear, was based on the analysis of small-scale shear zones, of the order of a few centimetres in width. However, their results have proven to be immediately applicable to similar zones of a much larger scale, showing a number of fabric and textural elements of their smaller analogues (Beach, 1976). That is true for example, in that both foliation and lineation show a constant orientation for large distances, that original metamorphic or igneous textures are preserved outside shear zones and that growth of new minerals, with advanced recrystallization is entirely confined to the shear zones.

The Lagoa Real region lacks detailed structural and geological mapping and for this reason, the relations between host rocks (themselves deformed) and altered products are not well established. Nevertheless, the combination of textural, mineralogical and chemical observations with scattered structural information, allow

the suggestion that the albitites or zones of alteration were developed within large-scale ductile shear zones, associated with extensive metasomatism (see Chapters 3 and 5), thus facilitating fluid migration. This was first suggested in Lobato et al (1982), and later discussed in Lobato et al (1983 a, b).

Original coarse grained k-feldspar rich gneisses are finely foliated and recrystallized, as one approaches the alteration zones. This fabric development is clear in thin section and hand specimen. Stein e Pereira (1979) clearly point out this change in the field; they observed that the k-feldspar gneisses are highly oriented in the proximities of albitites, but preserve their porphyroclastic character at greater distances (~1km), with transposition of the original foliation. Recrystallization alone is responsible for grain size reduction. These rocks appear bordering the albitites with abrupt contacts (as commonly observed in smaller scale shear zones; see Wilson, 1980) and have been designated type A "transitional" rocks in the present work.

Recrystallization coupled with restricted mineral changes mark the onset of fluid-rock interaction in type B "transitional" rocks, with the reaction amphibole --> garnet and/or pyroxene plus limited albitization taking place (see Chapter 6).

The albitites themselves seem to represent the areas of most extensive plastic deformation and recrystallization with the highest degree of fluid-rock interaction. Beach (1976) called attention to the fact that the metamorphic reactions taking place, promoted by the presence of the metasomatic fluid, will result in nucleation of small grains, facilitating an increase in ductility. He also suggested that as the growth of new phases occur along with deformation in the shear zones, it becomes apparent that metasomatism does not continue after deformation has ceased; the converse of this may also be important, that is, deformation ceases when metasomatism ceases. Clearly the deformation increases and controls the permeability.

The basic component in practically all shear zones is that of heterogeneous simple shear; as deformation proceeds, it is very common for the deforming mass to become anisotropic as a result of finite strains which build up with increasing shearing strain (Ramsay, 1980). This heterogeneous behaviour must account for the observed relic portions of unsheared original gneisses found within the zones of alteration.

7.3. Uranium: thoughts on transport and precipitation

It has long been known that uranium is usually transported in the hexavalent state, and in solution it has a pronounced tendency to form complexes with water, organic compounds and anions such as CO_3^{2-} , PO_4^{3-} , VO_4^{3-} , SiO_4^{4-} , etc (Rich et al, 1977). The cations mostly combined with the uranyl ion and corresponding complex anions are Ca^{2+} , K^+ , Na^+ , Ba^{2+} , Mg^{2+} , (Kostov, 1977). Uranium migration frequently occurs in alkaline carbonate solutions forming complex ions of the $[\text{UO}_2(\text{CO}_3)_2(\text{H}_2\text{O}_2)]$ and $[\text{UO}_2(\text{CO}_3)_3]$ types, the deposition of uranium under hydrothermal conditions being realized at 150°C ; sodium uranyl dicarbonate and tr carbonate solutions are perhaps the most frequent uranium carriers (Kostov, 1977).

Current available data for the solubility of uranium minerals are surprisingly poor. High-temperature uranium geochemistry is presently not well characterized, and the key question of how higher temperatures affect uranium solubility is unanswered to this day (see Kimberley, 1978). In recent years however, several researchers in the U.S.S.R. and France have produced some of the data required for the understanding of the origin of uranium deposits formed at temperatures above 100°C . Lemoine (1975, in Kimberley, 1978) for instance, found a solubility maximum for uraninite at about 260°C in water

at a pressure of 750 bar, and oxygen fugacity buffered by equilibrium between UO_2 and hydrous or anhydrous UO_3 .

The association between U mineralization and Na-metasomatism has been noted mainly by Russian geologists (Nikolskii, 1973, Tarkhanov, 1975, Smirnov, 1977, Belevtsev, 1980, Grechishnicov, 1980, Kalyaev, 1980, Tugarinov, 1980, Zhukova, 1980). Some Russian deposits show similarities with the Lagoa Real deposit in that they comprise linear, tectonic controlled, Na-metasomatized zones (albitites). Belevtsev (1980) studied a few of these deposits and proposed that the alteration is characterized by the formation of sodic silicates at earlier stages, with later precipitation of carbonates and oxides. He concludes that U was transported by hydrothermal solutions as complex uranyl-carbonate or bicarbonate ions. Field evidence coupled with experimental results led the same author to conclude that:

- the solubility (1 to 10 mg/l) of hexavalent uranium differs only slightly with temperature and pressure changes ($T = 200^\circ$ to 500°C , and $P = 490$ bar). The solubility increases 10 to 100 times when U is transported in a carbonate medium with potassium and sodium ($T = 200^\circ$ to 400°C). In the presence of CO_2 (observed in fluid inclusions in the area), between 300° and 400°C , K and Na are easily transported, creating conditions for the

formation of sodium-carbonate solutions capable of transporting uranium as a complex compound. Uranyl-carbonate ions are stable only in alkaline media (ph = 7 to 9), at temperatures of 200° to 300°C;

- pressure has an essential influence on the stability of complex uranium ions in solution; higher pressures (200 to 500 bar) dramatically increase uranium concentrations;

- uranium precipitation is a combination of many factors: (a) as Na silicates (mainly albite) form, the $\text{Na}_2\text{O}:\text{CO}_2$ ratio in metasomatic solutions is depressed, and the stability of uranyl-carbonate ions decreases due to the sharp decrease in alkalinity; (b) crystallization of carbonates (possibly stimulated by T decrease) assists in further decomposing the uranyl-carbonate ion and that leads to ore deposition; (c) the presence of reducing agents (Fe^{+2}), promotes the reduction of hexavalent uranium with precipitation of uraninite; in the absence of Fe^{+2} bearing phases (i.e. in parageneses with quartz), the fluids released uranium that combined with silica to form uranium silicates.

Romberger (1978) also provided some information on the stability of uranyl complexes in oxidizing conditions at 25° and 300°C. He derived the following generalizations:

- carbonate complexes are very important at low temperatures in slightly acid to alkaline pH's in the absence of phosphate; their stability decreases at elevated temperatures;

- at elevated temperatures in fluoride-containing uranium-transporting systems, fluoride complexes predominate well into the alkaline region; carbonate species are restricted to very high pH's;

- the stability of most uranyl complexes, with the exception of hydroxide complexes, decreases as temperature increases;

- deposition takes place upon reduction of the U^{4+} state.

The presence of CO_2 rich fluid inclusions (see 6.2) has been demonstrated in a number of hydrothermal uranium deposits (see Spooner, 1981). The same is true for hydrocarbons (Poty in Kostov, 1977).

In the Lagoa Real rocks, uraninite is chiefly related to pyroxene --> andradite replacement, as by far the highest percentage of uraninite grains are included in the garnet. Uraninite is also common inside pyroxene and hematite (magnetite) and may also occur associated with albite, hornblende and rarely with sphene. In conclusion, uraninite precipitation is in great part controlled by reduction of the fluid phase, and the main precipitation event took place when pyroxene and magnetite were being oxidized (forming andradite and hematite respectively).

Similar mechanisms for the precipitation of uraninite or pitchblende involving hematite formation, are well known in the literature (Rich et al, 1977, Nash et al, 1981). Precipitation of uraninite and allanite associated with andradite formation is reported by Derrick (1977) for the Mary Kathleen uranium deposit in Australia. He concluded that an early high-temperature K-metasomatism of Proterozoic hornblende-hornfels was followed by a lower temperature metasomatic event that resulted in extensive garnetization (andradite) and skarn replacement of earlier assemblages, to which uraninite formation is related; fluids associated with this second event were rich in Na (albite/oligoclase is formed), H_2O , O_2 , CO_2 , U and rare-earths. This interpretation was however refuted by Page (1983), who believes that the uranium mineralization

is much later on the basis of geochronological studies.

The question of U transport is not resolved. High CO_2 concentrations in fluid inclusions may support a transport mechanism similar to that suggested by Belevtsev (1980), and described above, though the temperatures in Lagoa Real are higher. Carbonate precipitation is also late stage and may have been inhibited at the initial stages due to high salinities, as suggested by fluid inclusion studies; addition of NaCl to CO_2 - H_2O fluids results in increase in calcite solubility (Fyfe et al, 1978), but as salinities were probably lowered towards late stages (stages (III) and (IV) in section 6.3.5.), carbonate precipitation occurred. CH_4 is also observed in many fluid inclusions. Given carbon compounds, it is possible that organic molecules are present and were thermally cracked at high temperatures. This possibility is further indicated by the association with vanadium well known to be associated with organics in petroleum.

It should be stressed that whereas most experimental works deal with uranium solubility with changing temperature, the ultimate process leading to U transport and precipitation depends on the interplay of many factors. Among these, variations in the redox potential can play a major role in U accumulation as attested by low temperature sandstone-type uranium deposits (Melin, 1957).

7.4. -U and Na metasomatism: other examples

Albite is a common phase in assemblages associated with uraninite and pitchblende; but "uranium deposits in albitites" is a relatively new concept which was first recognized by Russian geologists (Sarcia, 1983). These deposits usually occur within shear and breccia zones, connected with deep-seating lineaments, without obvious magmatism, and may contain substantial reserves of uranium and thorium (Smirnov, 1977). That author classified some Russian deposits on the basis of their structural features and metal contents. He suggests the name "linear albitites" for the hydrothermal, non-magmatic albitites occurring along fault zones cross-cutting the Precambrian basement; these deposits may contain substantial reserves of uranium and thorium primarily structure-controlled.

Although many albitite deposits have been generally described as hydrothermal in nature, the classification "shear zone and breccia deposits" was suggested by Backström (1974). A series of albitite deposits in the Precambrian of the U.S.S.R. have been described by Nikolskii (1973), Tarkhanov (1975), Belevtsev (1980), Grechishnicov (1980), Kalyaev (1980), Tugarinov (1980), Zhukova (1980); although exact locations and detailed descriptions of the mineralogy and chemistry are not provided, a summary of their general characteristics is

presented below:

- uranium bearing albitites occur in metasomatically altered gneisses (originally amphibolite or granulite facies), granites and pegmatites, older than 2 Ga, and in most cases the mineralization is about 1.7 Ga. They are located and related to large faults, and associated with carbonate metasomatism as well;

- in most deposits; the mineralogy includes: albite, alkaline pyroxenes (aegirine and aegirine-augite), amphiboles (riebeckite), quartz (many times absent), chlorite, epidote, carbonates, sulfides, magnetite, hematite;

- uranium minerals may include uraninite, pitchblende, davidite, coffinite and brannerite;

- vanadium and copper may be associated with U;

- in many deposits, Na silicates form first, followed by the precipitation of carbonates and oxides (hematite), assisting uraninite precipitation;

- the albitite bodies form columnar, lenticular flattened or saddle-shaped lodes;

uraninite in all deposits is chiefly concentrated within areas of intense martitization of magnetite; temperatures of formation are 400°-500°C.

More recently, similar associations have been recognized elsewhere. In the Mistamisk Valley of the Labrador Trough, Quebec, Kish and Cuney (1981) describe uraninite-albite veins, deposited in fractures by hydrothermal solutions causing albitization of wall rocks (argillites). The mineralogy includes: albite, chlorite, dolomite, uraninite, and minor quartz, tellurides, sulfides, gold, hematite and organic material. Fluid inclusion studies show oversaturation in NaCl (salinities of ~25% NaCl equivalent) and contain Na, Ca, Mg, Fe, and CO₂. Temperatures of formation are estimated at 300-350°C at 2.5 kbar. Vein emplacement occurred in the waning stages of the Hudsonian Orogeny; hydrothermal solutions probably originated by metamorphism of sodic schists of the Mistamisk area, which are possibly of evaporitic origin.

In the Arjeplog-Arvidsjaur Precambrian uranium province of Sweden (Adamek and Wilson, 1977), pitchblende occurs associated with sodium metasomatism of silicic metavolcanic rocks and granites. The mineralization has a spatial relationship to fault zones which have a NNE trend. Albite replacing k-feldspar, loss of quartz and

formation of riebeckite and chlorite constitute the main mineralogical characteristics. The mineralization postdates albitization; uranium was possibly transported by oxidizing fluids in the form of uranyl-carbonate ions. As pitchblende precipitates, carbonates, chlorite and hematite (from magnetite) formed, with rare garnet.

Other examples are the uraniferous mineralization in the Brousse-Broquies Basin, Aveyron, France (Yerle et Thiry, 1979) and that of the upper Aillik Group in Labrador (White and Martin, 1980).

CHAPTER 8

METAMORPHISM, METASOMATISM AND MINERALIZATION

AT LAGOA REAL: A SYNTHESIS

8.1. Introduction

In the attempt to integrate the metamorphic, metasomatic and mineralizing processes operating in the Lagoa Real area within a geological model, one must necessarily take into account a number of observations and interpretations generated by field relations, petrography, mineral chemistry, rock chemistry, metamorphic history, oxygen isotope geochemistry and fluid inclusion studies. A number of important points should be considered:

- 1) U bearing albitites and epidotes are formed upon metasomatism of amphibolite facies orthogneisses of the Archean and Lower Proterozoic Paramirim basement terrain. The metasomatic alteration is restricted to ductile shear zones that transect the basement. The original assemblage in the gneisses is: orthoclase,

oligoclase or albite-oligoclase, quartz, hornblende (hastingsite and Fe-pargasite), biotite (after hornblende), magnetite, ilmenite and sphene; such rocks are incipiently banded, very coarse grained, porphyroclastic, usually highly deformed.

In the vicinity of ore bodies, these k-feldspar-gneisses acquire a highly foliated fabric, granoblastic and polygonal granoblastic texture, and fine to medium grain size, but maintain the same original mineralogy (type A "transitional" rocks). In some of these rocks, the replacement of hornblende by pyroxene and/or garnet is observed as the first sign of alteration (type B "transitional" rocks); their mineralogy includes: microcline, oligoclase or albite-oligoclase, hornblende (hastingsite and Fe-pargasite), hedenbergite or Fe-salite, a grossular-andradite garnet, magnetite, fluorite, sphene. In most samples of "transitional" gneisses, fine albite rims develop surrounding original plagioclase.

In the zone of most intense alteration, the rocks are fine to medium grained albitites (and also epidositites), exhibiting a polygonal granoblastic texture, usually incipiently and sometimes strongly foliated. The most common mineral assemblage in the albitites is: albite (An 2-7%), aegirine-augite (zoned), andradite (maximum of ~ 10% of the grossular component, replacing pyroxene),

magnetite (mostly altered to hematite and/or sphene), hornblende (hastingsite, Mg-hastingsite and Fe-pargasite, after pyroxene), microcline and quartz (both amounting to ~5%). In a few samples, biotite may occur after hornblende, epidote after all mafic minerals, calcite mostly after andradite and prehnite after pyroxene; microcline and/or quartz (as much as ~20-30%) were observed in rare samples. Ca-rich albitites may contain wollastonite (slightly altering to pectolite), calcite (up to 30%) and vesuvianite, besides the minerals listed above; in addition, they contain pyroxene richer in MgO (up to diopside or salite) and andradite richer in the grossular component (up to 37%).

Mineralized albitites contain uraninite which is associated chiefly with andradite, but also with aegirine-augite, magnetite altering to hematite, albite and hornblende. Epidosites are made up of: aegirine-augite, epidote (Ps = 26 to 28, replacing or pseudomorphing pyroxene), oligoclase (An ~18%), quartz (5-30%), sphene and microcline (less than 2%). In mineralized epidosites, uraninite is usually absent, and U is locked in the epidote.

2) Based on the mineral reactions and mineral associations, the metasomatic process can be divided into

the following stages:

(I) dehydration/oxydation of hornblende, with silica consumption; restricted Na enrichment (albite rims, albite replacing microcline);

(II) Na addition: widespread formation of albite; Na addition and oxidation: formation of zoned aegirine-augite from hedenbergite; dissolution of quartz; limited uraninite precipitation (in albite and pyroxene);

(III) oxidation of aegirine-augite and magnetite: formation of andradite and hematite; widespread uraninite precipitation; formation of hornblende after pyroxene; limited alteration of hornblende to biotite.

3) The geochemical trend from "unaltered" to altered rocks is: removal of SiO_2 , K_2O , Rb and Ba, with an estimated volume loss of 10%; enrichment in Na_2O , Fe_2O_3 , Sr, Pb, V and U.

4) Fluid inclusion studies suggest salinities ranging from ~1.5 to 23% (NaCl equivalent), rich in CO_2 and CH_4 . Albite formation is widely restricted to the early phases of the alteration, reflecting high salinities in the early stages of the alteration as opposed to lower salinities in the later stages.

5) $^{87}\text{Sr}/^{86}\text{Sr}$ studies indicate that probably Sr, and by extension U and its gangue, were not derived from the host gneisses.

6) The general mineralogy of the altered rocks corresponds to epidote-amphibolite facies and temperatures in the range of 500-600°C at moderate pressures (Turner, 1981), which also constitute the estimated range of temperatures for the regional and type A "transitional" rocks. Pressure estimates are in the order of 4 kbars or less.

7) Oxygen isotopes thermometry yields temperatures which are compatible with the metamorphic assemblages.

$\delta^{18}\text{O}$ isotope values of the fluid phase show a range of values of 7.3 to -3.7‰. This variation of values is interpreted as the result of the exchange with a $\delta^{18}\text{O}$ depleted fluid at varying water/rock ratios. A value of -3.7‰ is consistent with a signature for meteoric water (Taylor, 1979).

8.2. Discussion

The texture in altered rocks is open saccharoidal and in general, a noticeable foliation is present. The large

scale of alteration observed in the shear zones involving volume loss, addition and removal of components, requires pervasive interaction with a fluid phase.

To account for the "opening" of such high grade rocks at depths, high fluid pressure ($P_{\text{fluid}} > P_{\text{lithostatic}} + \text{tensile strength}$) should have prevailed at least during the initial shearing process. Once fluid flow was established, P_{fluid} changed, and may have eventually approached hydrostatic equilibrium, assuming that the fluid paths reached the surface (Fyfe et al, 1978). It should also be noted that both silica removal and Na-exchange of microcline should lead to an increase in porosity as a volume loss of about 10% has been estimated.

Given the above considerations, the characteristics of the mineralizing fluid and the conditions under which the fluid moved are constrained. Important points which must be accounted for are:

a) the fluids must have a high Na/K ratio to remove K-feldspar and cause the dramatic increase in sodium. Given the high salinities, it is likely that sodium chloride brines may have been involved;

b) to account for the removal of about 10% silica

and virtual removal of quartz, fluids must be moving up a temperature gradient. If fluids were originated in any normal magmatic source, they would precipitate silica;

c) fluids were oxidizing. Normal deep metamorphic fluids contain excess H_2 and when moving down temperatures, must reduce rocks. (Beach and Fyfe, 1972); this fact, coupled with the oxygen isotope data helps rule out high temperature metamorphic fluids;

d) the fluids must have been capable of carrying uranium and vanadium. The absence of Th in the alteration, also lends support to a non-magmatic origin for the fluid as most uranium mineralizations associated with igneous processes contain appreciable thorium (Adler, 1977, Sørensen, 1977).

8.3. Model of metasomatic and mineral processes

A possible model arises from the regional structure of the area. A major thrust structure has long been reported in the area along the Espinhaço range/basement border (Moutinho da Costa et al, 1976, Schobbenhaus, 1976, Moutinho da Costa e Inda, 1982; see Fig. 2.1). This structural feature has been interpreted by Moutinho da

Costa e Inda (1982) as having been generated by the uplifting of the Paramirim "block", posterior to the main metamorphic event in the Espinhaço; these ideas are also corroborated by Jardim de Sá (1978). Along this segment, those authors suggested that the basement rocks were thrust and folded to the west over the Espinhaço strata; there is no doubt that the latter dips under the basement. Retrograde metamorphism in parts of the basement to the east and the local prograde metamorphism in rocks of the Lower and Middle Espinhaço (appearance of kyanite and chloritoid) are related to the thrusting. This general structure appears to be common for about 200km along a general N-S trend in Bahia and to the south into the State of Minas Gerais (Schobbenhaus, 1976).

To support this line of evidence, are the gravimetric and magnetic studies reported by Motta et al (1981). To the east of the State of Bahia (towards the coast), depths of the crust/mantle interface are calculated at 32km. Such depth increases to about 43km at the Paramirim valley region, where gravity data suggest a geofracture that coincides with the Espinhaço alignment, from about north of Oliveira dos Brejinhos ($12^{\circ}00'$) to a little north of Caetité ($14^{\circ}00'$); it constitutes an elliptical negative anomaly trending $N30^{\circ}/45^{\circ}W$. Thickening of the crust across major thrust faults is a well documented feature

(e. g. Fountain and Salisbury, 1981).

The uranium anomalies make up an arcuate zone parallel to this regional structure. Figueiredo (1980, 1981) and Sighinolfi et al (1981) have drawn attention to the geochemistry of granulites over wide regions of the State of Bahia (i. e. lack of U, Rb depletion, etc.). The regional rocks surrounding the ore deposits are amphibolite facies gneisses exhibiting retrograde effects, with an average U content of 3.8 ppm. This value is much higher than those reported for the granulitic rocks, and is ~1.4 times higher than the average crustal abundance of 2.6 ppm (Dyck, 1978), but similar to those reported by Dostal and Capedri (1978), that are in the range of 1 to 3.43 ppm. The relatively high U content of the regional gneisses has led to speculations that these rocks could be the source of uranium (Cordani, 1982). However, this interpretation is rendered difficult by the $^{87}\text{Sr}/^{86}\text{Sr}$ results.

Similar large scale thrust structures have been reported by many workers on the Southern Appalachians (Cook et al, 1980, Brewer and Oliver, 1980, Cook et al, 1983). Their "thin skinned" thrust model, involves the thrusting of thin sheets (10-20km thick) of older crystalline basement over younger sedimentary formations for distances of over 260 km (see also Meissner et al,

1981).

In the Lagoa Real district, this structure may have resulted from the updoming of the crust to the east, probably by a thermal high and subsequently, the updomed basement was detached and slid to the west, in a mechanism similar to that described by Hyndman (1980; see also Coney, 1983). Jardim de Sá (1978) and Moutinho da Costa e Inda (1982) proposed a mechanism similar to this when interpreting the structural framework of the Paramirim valley region. Acid plutonic rocks farther to the north being dated at 815 Ma (Chapter 2) have a similar age to that proposed for the uranium mineralization.

A 12 km thick gravity driven thrust slice must have had basal temperatures close to the melting temperature of granite (c. f. Hyndman, 1980). Such temperatures can explain the prograde metamorphism observed in the underthrust Espinhaço, near the contact with the basement rocks. When thrusting occurs, compaction of the underplate must occur rapidly, expelling pore water from it. High fluid pressures (similar to or greater than lithostatic pressures) are likely to be attained; this is a necessary requirement to maintain thrust motion, as first stressed by Hubbert and Rubey (1959). Fluids generated at the base by loading and heating would penetrate the basement by hydraulic fracture mechanism,

resulting in retrograde metamorphism of the base of the overriding plate.

Rising fluids will encounter inverted thermal gradients; the extent of retrograde metamorphism depends on the distribution of temperature, and the path taken by the escaping fluids (Crawford and Mark, 1982). The underplate high level rocks would tend to be more oxidized than basement rocks, and cold. Salty fluids rising under the thrust into a hot overplate at high fluid pressures could cause the oxidation, silica removal and sodium metasomatism. The regions of fluid penetration would be confined to zones of weakness in the overthrust plate (shear-zones). See Figs. 8.1, 8.2 and 8.3.

As already mentioned, the appearance of kyanite in the lower Espinhaço sequences is indicative of local higher metamorphic conditions. Some aspects of the retrograde alteration of the basement rocks at Lagoa Real must be considered in light of the basic similar metamorphic conditions of both the altered and "unaltered" rocks. We have seen that the onset of the mineral transformations is marked by the dehydration/oxydation of original hornblende to form anhydrous phases like garnet and/or pyroxene. Taking into consideration the genetic model proposed above, which involves large quantities of

Fig. 8.1. Schematic diagram illustrating mid to late Proterozoic tectonic evolution of the basement/Espinhaço border around the Lagoa Real area, leading to alteration/mineralization (see text). Stippled pattern: Espinhaço sediments; in black and white: basement.

- A: Espinhaço platformal cover rests on the Archean basement
- B: development of a thermal anomaly and updoming of the basement and cover followed by thrusting (C)
- D: metasomatism and mineralization; development of shear zones at Lagoa Real. Inset is presented in figure 8.3.

Fig. 8.2. Schematic diagram displaying the proposed model for the Lagoa Real region (see text). Dewatering of the Espinhaço sediments occurs as a response to load caused by thrusting. Flow is concentrated along thrust plane and in places penetrates zones of structurally enhanced permeability leading to metasomatism and mineralization.

fluids (initially cold), hydration reactions would more likely be expected or alternatively, hydrated phases like hornblende and biotite should have remained stable. The dehydration/oxydation reaction of hornblende is itself controlled by the change towards higher temperatures; as dissilicification is promoted by rising temperatures, silica activities in the fluid rise, facilitating hornblende dehydration. It should again be pointed out (see Chapter 6) that the high salinities in the fluid phase must have considerably lowered the activity of water, again enhancing dehydration. It is also important to note that towards the later stages of the alteration, hornblende (which may contain uraninite inclusions) starts to form after pyroxene; quartz results from this and pyroxene $\rightarrow$ andradite reactions and late quartz veins are found throughout the area; calcite precipitates (probably in part due to lowering salinities). Later establishment of lower T conditions is suggested by the limited replacement of pyroxene by prehnite.

Oxygen isotope systematics of the Lagoa Real rocks lends support to the thrust model. The trend towards lighter isotopic ratios in the more altered rocks and the relative constancy of isotopic fractionations among minerals in all rock types, is compatible with an early low temperature metasomatic episode, followed by prograde

isotopic reequilibration. This process could be explained by a thrust mechanism occurring in a single fast phase, so that maximum temperatures were attained rapidly.

Oxburgh and Turcotte (1974) studied the thermal gradients and regional metamorphism in overthrust terrains (in particular the Eastern Alps); they considered a thrust sheet of 15 km thick with a temperature of T at the base of the slab and a temperature of zero at the top of the underlying rocks. They showed that after thrusting there is an initial cooling in the slab at all depths. After 4 m. y., steady temperature increases occur at all depths. The initial temperature distribution and associated negative thermal gradients are extinguished in less than a million years, and by 3.3 m. y. the gradient through the slab is nearly linear and has about half the value of the initial gradient. Towards this stage, normal metamorphic conditions should be reestablished as temperature reequilibration is under way.

The suggestion that the metasomatic alteration/mineralization at Lagoa Real is linked to the thrust mechanism is chosen as adequate because, not only a thrust structure has been mapped in the region supported by field relations and gravimetric data, but also because such a process can explain the various facets of the alteration/mineralization discussed throughout this

thesis. It is however important to point out, that attempting to model the mechanism(s) by which a thrust structure evolved in the region is complicated by the poor geological knowledge at the present, and therefore transcends the scope of the present study.

If the temperature path in Lagoa Real is interpreted in a similar fashion, it can be concluded that either the altered rocks exposed today do not come from very deep levels of the basement slab, because there is no evidence of low temperature alterations, or later reequilibration with higher temperatures (return to normal metamorphic thermal gradients) have masked initial lower temperature assemblages, as previously suggested for oxygen isotopic studies.

Inverted metamorphism in regional metamorphic terrains associated with thrusting has gained a lot of attention in recent years (Graham and England, 1976, Crawford and Mark, 1982, Arita, 1983). Inverted metamorphic conditions for the Lagoa Real region is summarized in Fig. 8.3.

Fluids that are moving up a thermal gradient can do so by moving down or upwards into rocks. Either process, if involving a light, salty meteoric water moving from an oxidized zone and precipitating uranium upon reduction,

Fig. 8.3. Schematic representation of geothermal gradients across the Lagoa Real crustal section, before (A), during (B) and after (C) thrust.

could explain oxidation, silica removal, etc. In considering the former case as a second possible model, the great problem is to have a mechanism for very deep descent of such fluids into a zone at over 500° C at ~12 km depth. In this case, one possible mechanism might involve a rising diapiric pluton at even greater depth with a fracture system penetrating from great depth; such a model was suggested for the area by Geisel (1981). Major problems with this model are that at the P-T conditions envisioned, rocks are plastic and impermeable (Rutter and Brodie, 1985) and it is very difficult to produce deep conditions where $P_{\text{fluid}} > P_{\text{lithostatic}}$ by a descent model.

8.4. Fluid source: implications

As pointed out in Chapter 2, the metasediments of the Espinhaço sequence reflect a diverse spectrum of platform continental, coastal and marine sedimentary facies; a succession of marine pelites predominate to the west whereas to the east, fluvial-deltaic sediments are more abundant (Jardim de Sá, 1978). The nature of the fluids which might be released from such rocks is not known.

$^{87}\text{Sr}/^{86}\text{Sr}$ isotope determinations on samples of graphite-phylrites and muscovite-quartz-schists of the Espinhaço were carried out by Kerrich et al (1984).

Although their significance is questionable in light of the dramatic lithological variations encountered in the Espinhaço sequence, some of the results may have some bearing on the present discussion. Five samples were analysed and calculations were done according to arguments introduced in Chapter 5; at 820 Ma, the calculated $^{87}\text{Sr}/^{86}\text{Sr}$ values are as follows:

graphite-phyllite	0.7419
graphite-phyllite	0.7385
hematite-phyllite	0.7311
muscov-qz-schist	0.7490
muscov-qz-schist	0.7002

The average from these values is 0.732. These values are consistently lower than those obtained for the regional gneisses (from 0.7379 to 0.8143; see Table 5.8), and approximate the values obtained for the albitites (0.7045 to 0.7234). Modern continental surface waters have a strontium isotope signature dominated by the solution and recycling of Phanerozoic marine carbonates (0.708), depending on the precise nature of the terrain being weathered (Faure, 1977). The original fluids contained in the variable lithologies of the Espinhaço sequence were presumably of widely differing Sr contents. Thus, recharge fluids contained in aquifers within the

Espinhaço might acquire a strontium isotopic signature between that of marine carbonates (0.708) and the average of the group as a whole (0.732).

An analogy can be made with modern formation brines, where a low $\delta^{18}\text{O}$ fluid is possible (Taylor, 1979). Modern formation brines are saline waters, rich in dissolved species (Na as one of them) and may be oxidizing (Hanor, 1979).

A final point to be mentioned, is the constant association of vanadium (up to 870 ppm) with the mineralized rocks. Its association with the mineralization again lends support to a low temperature sedimentary (biological) origin of the fluid. Vanadium is an important species associated with a number of low temperature uranium deposits in sediments where it is in part responsible for uranium precipitation (Adler, 1974). One should still add that the presence of CH_4 in large amounts as reported in Chapter 6, further contributes to this interpretation.

APPENDIX I

MICROPROBE ANALYSES OF MINERALS FROM REGIONAL, "TRANSITIONAL" AND METASOMATIC ROCKS

- I-1. Plagioclases from regional gneisses
- I-2. Amphiboles from regional gneisses
- I-3. Biotites from regional gneisses
- I-4. Plagioclases from "transitional" rocks
- I-5. Amphiboles from type A "transitional" rocks
- I-6. Amphiboles from type B "transitional" rocks
- I-7. Biotites from "transitional" rocks
- I-8. Clinopyroxenes from type B "transitional" rocks
- I-9. Garnets from type B "transitional" rocks
- I-10. Epidotes from "transitional" rocks
- I-11. Plagioclases from metasomatic rocks
- I-12. Clinopyroxenes from metasomatic rocks
- I-13. Garnets from Ca-rich albitites
- I-14. Garnets from common albitites
- I-15. Amphiboles from metasomatic rocks
- I-16. Epidotes from epidiosites

Table 1.1-Analyses of plagioclases from regional gneisses.

Sample no.	1	2	3	4	5	6	7	8	9	10	11
	wt%	wt%	wt%	wt%	wt%	wt%	wt%	wt%	wt%	wt%	wt%
SiO ₂ (wt%)	64.54	63.38	64.02	63.63	64.52	64.47	64.52	64.46	68.54		68.65
Al ₂ O ₃	21.81	21.99	22.56	22.23	21.84	22.05	21.76	20.05	19.86	19.42	19.44
CaO	3.15	3.19	3.68	3.79	3.06	3.22	3.21	0.86	0.60	0.40	0.21
Na ₂ O	11.19	10.63	9.35	9.68	10.07	10.08	10.49	9.73	10.98	11.54	10.81
K ₂ O	0.10	0.05	0.07	0.09	0.19	0.16	0.16	0.26	0.22	0.11	0.77
Total	100.79	99.24	99.68	99.42	99.68	99.98	100.12	100.36	100.20	99.13	99.88

Number of ions on the basis of 32 oxygens											
Si	11.352	11.305	11.320	11.308	11.418	11.382	11.395	12.022	11.944	11.939	12.013
Al	4.520	4.622	4.680	4.655	4.555	4.587	4.524	4.089	4.056	4.038	4.009
Ca	0.594	0.610	0.697	0.722	0.580	0.609	0.607	0.199	0.112	0.076	0.039
Na	3.816	3.676	3.205	3.335	3.455	3.450	3.592	3.265	3.710	3.948	3.668
K	0.022	0.011	0.016	0.020	0.043	0.036	0.036	0.057	0.023	0.025	0.172
Arb	13.39	14.19	17.79	17.78	14.23	14.87	14.34	4.58	2.89	1.87	1.02
Arb	86.10	85.55	81.80	81.80	84.72	84.25	84.81	93.77	95.84	97.52	94.55
Orb	0.51	0.25	0.40	0.50	1.05	0.88	0.85	1.65	1.26	0.61	4.43

1-7: oligoclase; 8-11: albite

Table 1.1: Analyses of amphiboles from regional gneisses.

Sample no.	1	2	3	4	5	6	7	8	9	10	11	12	13	14
	4430	4430	4430	4430	448	448	4433	4434	4437	4437	4437	4437	4437	4437
SiO ₂ (wt)	40.40	39.25	38.14	39.32	37.77	37.13	36.73	38.78	37.75	36.88	36.51	35.94	36.42	36.48
Al ₂ O ₃	9.30	10.44	10.80	10.66	12.09	12.00	12.19	11.54	12.53	13.33	12.25	14.88	11.32	12.15
FeO	11.61	12.47	12.56	12.54	11.38	11.21	11.57	10.75	12.19	10.11	12.31	10.46	11.02	12.44
Fe ²⁺ O	10.19	29.95	27.63	30.49	29.51	29.75	27.41	29.02	27.99	26.42	25.58	25.15	28.79	28.12
Fe ³⁺ O ₃	1.58	2.79	5.63	2.60	1.75	2.76	4.48	1.93	4.66	4.10	7.48	5.99	2.47	4.80
MgO	1.23	0.96	1.13	0.97	0.61	0.35	0.63	1.12	0.48	0.73	0.63	0.69	0.60	0.64
MnO	0.61	0.53	0.59	0.64	0.49	0.46	0.42	0.54	0.31	0.68	0.43	0.72	0.35	0.61
TiO	1.43	0.52	0.38	0.78	0.46	0.35	0.23	0.48	0.50	0.52	0.36	0.46	0.48	0.53
CaO	10.61	11.07	11.07	10.68	10.66	10.68	10.70	10.82	10.84	10.23	10.32	10.10	10.77	10.73
Na ₂ O	2.08	1.98	1.57	2.11	1.98	1.54	1.65	1.70	1.61	1.03	1.27	1.53	1.32	1.34
K ₂ O	1.71	1.92	1.94	1.94	2.13	2.10	2.39	2.30	2.14	2.03	2.13	2.23	2.44	2.14
Total	99.63	99.61	98.90	99.63	97.77	96.70	97.19	94.23	95.25	97.08	97.53	97.32	97.73	99.64

Number of ions on the basis of 23 oxygens

Si	4.444	4.304	4.167	4.309	4.179	4.134	4.084	4.280	4.113	4.165	3.972	3.833	4.277	4.167
Al	1.554	1.885	1.933	1.891	1.921	1.844	1.784	1.749	1.987	1.935	2.038	2.148	1.723	1.833
Al _{IV}	0.259	0.281	0.262	0.325	0.495	0.471	0.417	0.469	0.506	0.708	0.334	0.607	0.561	0.491
Fe ²⁺	4.028	4.024	3.705	4.052	4.079	3.971	3.935	3.930	3.791	3.576	3.502	3.621	3.915	3.788
Fe ³⁺	0.190	0.338	0.693	0.334	0.215	0.340	0.556	0.235	0.568	0.699	0.918	0.721	0.302	0.576
Mg	0.292	0.229	0.272	0.232	0.100	0.084	0.154	0.270	0.114	0.176	0.154	0.167	0.145	0.157
Mn	0.062	0.072	0.081	0.085	0.068	0.084	0.088	0.078	0.082	0.082	0.115	0.089	0.076	0.082
Ti	0.172	0.089	0.064	0.084	0.106	0.119	0.028	0.054	0.061	0.063	0.068	0.076	0.089	0.063
Fe ²⁺ /Fe ³⁺	0.056	0.064	0.039	0.783	0.041	0.051	0.051	0.057	0.085	0.105	0.089	0.112	0.038	0.074
Ca	1.514	1.595	1.516	1.522	1.548	1.490	1.490	1.474	1.518	1.774	1.409	1.739	1.474	1.433
Na	0.132	0.090	0.065	0.085	0.070	0.054	0.054	0.063	0.087	0.121	0.102	0.128	0.086	0.089
K	0.515	0.545	0.647	0.561	0.526	0.435	0.465	0.464	0.344	0.292	0.301	0.354	0.349	0.325
K	0.348	0.393	0.404	0.397	0.444	0.443	0.503	0.475	0.442	0.419	0.444	0.463	0.458	0.443
Fe ²⁺ /Mg	14.70	19.29	14.44	19.10	43.42	50.74	25.74	5.68	37.99	23.41	25.52	25.36	29.53	24.10
Fe ²⁺ /Fe ³⁺	0.94	0.95	0.94	0.95	0.96	0.99	0.97	0.94	0.97	0.94	0.97	0.94	0.97	0.97

1, 4, 5, 6, 8, 10, 13: Ferro-pergarnite; 2, 3, 7, 9, 11, 12, 14: Magnetite

Calculated according to Palmer et al. (1976)

Def. total iron

Table I-3. Analyses of biotites from regional gneisses.

Sample no.	1 AM30	2 AM30	3 AM30	4 AM30	5 AM37	6 AM37
SiO ₂ (wt%)	33.54	34.00	34.43	34.54	34.03	33.96
TiO ₂	3.47	4.25	2.11	2.01	1.67	2.07
Al ₂ O ₃	12.41	12.49	13.41	13.87	14.50	14.71
Cr ₂ O ₃	0.12	0.11	0.03	0.10	0.14	0.05
FeO*	35.73	34.41	34.44	34.08	35.92	35.63
MnO	0.45	0.33	0.47	0.41	0.53	0.55
MgO	1.56	1.48	2.04	1.59	1.13	1.17
CaO	0.07	0.00	0.15	0.04	0.04	0.02
Na ₂ O	0.33	0.04	0.00	0.00	0.00	0.24
K ₂ O	9.56	9.02	8.63	8.77	8.95	9.34
Total	99.66	96.13	95.71	95.41	96.91	97.74
Ions on the basis of 22 oxygens						
Si	5.529	5.596	5.666	5.688	5.574	5.521
Al	2.411	2.404	2.334	2.312	2.426	2.479
Al	0.000	0.019	0.266	0.379	0.373	0.339
Ti	0.48	0.529	0.261	0.249	0.206	0.253
Mg	0.388	0.363	0.500	0.390	0.276	0.294
Fe	4.926	4.736	4.740	4.693	4.921	4.944
Mn	0.063	0.046	0.066	0.057	0.074	0.076
Cr	0.016	0.014	0.004	0.013	0.018	0.006
Na	0.105	0.013	0.000	0.000	0.000	0.076
Ca	0.012	0.000	0.026	0.007	0.007	0.003
K	1.900	1.894	1.811	1.842	1.870	1.937
Fe ^T /Mg	12.45	13.17	9.60	12.17	18.10	17.35
Fe ^T /Fe ^T -Mg	0.93	0.93	0.90	0.92	0.95	0.95

*FeO calculated as total iron.

Fe^T: Total iron.

Table 1-4. Analyses of plagioclases in "transitional rocks".

Sample no.	1	2	3	4	5	6	7	8	9	10	11	12
	AN2	AN2	AN2	AN2	AN2	AN2	AN2	AN2	AN2	AN2	AN2	AN2
SiO ₂ (wt%)	66.89	67.60	67.22	66.69	66.60	66.99	66.47	66.20	66.62	66.03	66.73	66.99
Al ₂ O ₃	20.64	20.13	20.99	21.49	20.33	20.29	20.99	20.35	20.22	21.46	20.70	20.26
CaO	1.99	0.43	0.55	2.04	1.95	1.66	1.62	1.64	1.16	1.77	1.99	0.66
MgO	10.36	11.64	10.96	10.39	10.93	10.03	10.99	10.52	10.74	10.99	10.76	11.74
FeO	0.23	0.09	0.09	0.20	0.14	0.10	0.20	0.19	0.21	0.21	0.20	0.13
Total	100.09	99.99	99.81	100.90	99.89	99.07	100.27	98.99	98.95	100.35	100.27	99.77

Ions on the basis of 32 oxygens

Si	11.725	11.867	11.766	11.614	11.720	11.816	11.651	11.739	11.792	11.573	11.692	11.790
Al	6.263	6.157	6.129	6.410	6.216	6.217	6.335	6.252	6.219	6.432	6.276	6.199
Ca	0.366	0.081	0.103	0.391	0.349	0.314	0.304	0.312	0.220	0.332	0.353	0.124
Mg	1.521	1.955	1.779	1.505	1.729	1.430	1.735	1.617	1.696	1.697	1.655	1.803
Fe	0.051	0.081	0.02	0.044	0.040	0.033	0.045	0.041	0.047	0.047	0.045	0.029
AN	9.30	1.99	2.68	9.69	9.67	9.33	7.65	7.85	5.56	9.15	9.71	2.99
AN	89.40	97.51	96.79	89.18	90.55	91.07	91.46	91.12	93.34	90.69	90.19	96.31
AN	1.31	0.50	0.52	1.13	0.94	0.60	1.09	1.03	1.20	1.15	1.10	0.70

Table 1-4.
(Cont.)

Sample no.	11	14	15	16	17	18	19	20	21	22	23	24	25	26
	wt%					ppm					wt%			
	49.51					120.95					140.19			
	49.57m	idem	idem	idem	idem	129.00m	idem	idem	idem	idem	52.63	idem	idem	idem
SiO ₂ (wt)	66.81	66.19	66.97	66.71	67.34	63.70	64.82	64.56	64.56	63.35	67.18	66.71	65.09	69.17
Al ₂ O ₃	20.17	20.22	20.27	20.71	20.55	22.98	22.14	22.16	22.74	21.15	20.61	20.60	22.39	20.12
CaO	1.22	1.21	1.15	1.23	0.53	3.25	2.58	3.29	3.17	4.14	0.19	1.98	3.10	0.71
MgO	10.45	10.94	10.64	10.43	12.07	9.83	9.60	9.73	9.80	9.26	11.41	10.43	9.81	11.07
K ₂ O	0.20	0.22	0.20	0.22	0.14	0.20	0.22	0.27	0.25	0.28	0.16	0.19	0.24	0.22
Total	99.63	98.98	99.98	99.80	100.83	99.96	99.80	100.21	100.32	100.18	99.95	100.11	100.63	100.49

Ions on the basis of 32 oxygens														
Si	11.780	11.765	11.796	11.755	11.761	11.751	11.751	11.765	11.727	11.786	11.776	11.698	11.797	11.762
Al	4.258	4.222	4.208	4.300	4.217	4.783	4.687	4.638	4.702	4.817	4.757	4.796	4.620	4.767
Ca	0.232	0.230	0.255	0.232	0.099	0.615	0.607	0.621	0.596	0.783	0.036	0.172	0.502	0.132
Mg	3.994	3.759	3.634	3.563	4.075	3.166	3.148	3.121	3.134	3.170	4.014	3.545	3.130	3.735
K	0.045	0.050	0.045	0.049	0.031	0.085	0.089	0.081	0.079	0.063	0.036	0.062	0.094	0.089
Arb	5.99	5.69	6.44	6.04	2.35	15.28	12.54	15.51	14.95	19.50	0.87	9.19	14.67	1.39
Arb	92.08	93.08	92.19	92.67	96.91	83.61	84.19	82.98	83.64	78.93	98.25	89.53	83.98	95.17
Orb	1.17	1.23	1.14	1.29	0.74	1.12	1.27	1.51	1.40	1.57	0.99	1.07	1.35	1.25

1, 4, 5, 7, 10, 15, 16, 19, 20, 21, 22: Plagioclase, isolated grains; 2: Plagioclase, center; 3: Same grain as 2, border; 4: Large porphyroblast; 5: Plagioclase, center; 9: Same as 8, border; 11: Rimmed plagioclase, center; 12: Same as 11, rim; 13: Coarse grain, center; 14: Same as 11, border; 17: Rim surrounding fine grain; 23: Plagioclase altered to sericite, border; 24: Same as 21, center; 25: Porphyroblast, center; 26: Same as 25, border

*type 9

Table 1-4.
(Cont.)

Sample no.	27 HRO1 208.92- 209.00m	28 idem	29 idem	30 idem	31 HRO1 18.0m- 18.10m	32 idem	33 idem	34 idem	35 HRO1* 107.11- 107.14m	36 idem	37 idem	38 idem	39 idem	40 idem	41 HRO1 29.07- 29.15m	42 idem	43 idem	44 idem
SiO ₂ (wt%)	67.46	66.60	67.04	64.96	64.69	64.72	65.32	65.06	67.45	68.03	67.80	67.11	66.66	66.96	65.18	65.42	64.99	64.11
Al ₂ O ₃	20.81	21.66	21.87	22.45	22.16	21.58	21.87	21.19	20.12	20.23	20.09	20.30	20.55	20.15	22.00	21.58	22.32	22.15
CaO	0.86	2.79	2.70	2.91	2.78	3.01	2.90	2.30	1.18	1.18	1.16	1.14	1.27	1.10	3.05	2.82	3.32	3.02
MgO	10.51	8.74	8.16	9.68	9.52	9.96	10.21	10.36	11.00	11.11	11.11	11.17	11.13	11.47	10.01	10.29	9.78	10.32
K ₂ O	0.16	0.20	0.18	0.10	0.19	0.10	0.29	0.10	0.14	0.09	0.06	0.14	0.09	0.13	0.14	0.00	0.16	0.09
Total	100.19	99.99	100.15	100.10	99.34	99.37	100.59	99.83	100.09	100.64	100.40	99.92	99.71	100.01	100.38	100.11	100.57	99.69

Ions on the basis of 32 oxygens																		
Si	11.777	11.640	11.668	11.410	11.444	11.471	11.453	11.596	107.11	11.836	11.833	11.782	11.779	11.755	11.440	11.503	11.390	11.354
Al	4.281	4.461	4.485	4.647	4.619	4.507	4.519	4.395	4.191	4.169	4.132	4.199	4.261	4.210	4.550	4.471	4.610	4.623
Ca	0.157	0.522	0.503	0.548	0.527	0.572	0.545	0.434	0.221	0.220	0.213	0.214	0.239	0.207	0.574	0.531	0.623	0.573
Mg	3.686	2.962	2.821	3.297	3.265	3.423	3.471	3.536	3.732	3.749	3.827	3.801	3.880	3.904	3.406	3.508	3.323	3.544
K	0.036	0.045	0.040	0.022	0.043	0.023	0.065	0.022	0.031	0.02	0.013	0.040	0.020	0.029	0.031	0.000	0.036	0.000
ApB	4.04	14.81	14.96	14.16	13.74	14.23	13.35	10.87	5.55	5.52	5.26	5.29	5.90	5.00	14.30	13.15	15.65	13.85
ApB	95.04	81.93	83.85	85.26	85.14	85.21	85.06	88.57	93.66	93.98	94.41	93.72	93.61	94.30	94.92	96.45	93.45	95.66
OpB	0.92	1.26	1.19	0.54	4.12	0.56	1.59	0.56	0.274	0.50	0.31	0.99	0.50	0.20	0.74	0.00	0.90	0.49

29-44: Plagioclase, isolated grains.

27: Rinned crystal, along rim; 28: Same as 27, centre.

*type B

Table 1-4.
(Cont.)

Sample no.	45 PR01 17.02- 17.10m	46 17.10m	47 17.10m	48 AA3 PAA- 413	49 17.10m	50 17.10m	51 17.10m	52 17.10m	53 AA3- HVV791	54 17.10m	55 AA6 HVV794	56 AA6 16.0.10m	57 PR01 16.0.10m	58 17.10m	59 17.10m
SiO ₂ (wt%)	67.90	66.96	66.43	65.71	66.35	65.00	68.04	68.21	67.98	67.80	67.23	66.96	66.80	66.61	65.05
Al ₂ O ₃	20.15	22.27	22.40	21.47	22.05	20.79	19.78	19.51	19.92	19.82	21.08	22.45	22.38	19.50	21.78
CaO	0.51	3.20	3.46	3.44	3.61	2.34	0.68	0.64	1.64	1.26	1.78	2.51	3.14	0.22	2.62
Na ₂ O	11.30	9.69	9.77	9.76	9.90	10.97	11.15	11.80	10.77	10.98	10.02	9.68	10.02	11.16	10.29
K ₂ O	0.15	0.16	0.21	0.26	0.25	0.08	0.09	0.12	0.26	0.24	0.13	0.10	0.17	0.05	0.15
Total	100.01	100.28	100.27	100.68	100.16	99.18	99.24	100.28	100.57	100.10	100.24	100.10	100.51	99.54	99.89

Ions on the basis of 32 oxygens															
Si	11.872	11.408	11.342	11.507	11.357	11.559	11.979	11.920	11.852	11.869	11.728	11.410	17.371	12.011	11.420
Al	4.151	4.608	4.647	4.430	4.586	4.356	4.000	4.018	4.092	4.089	4.333	4.667	4.628	4.013	4.525
Ca	0.096	0.602	0.653	0.653	0.683	0.446	0.128	0.120	0.306	0.236	0.333	0.548	0.590	0.041	0.495
Na	3.831	3.292	3.342	3.314	3.398	3.782	3.816	3.998	3.640	3.727	3.889	3.797	3.409	3.788	3.518
K	0.033	0.016	0.042	0.058	0.056	0.080	0.020	0.12	0.058	0.054	0.029	0.022	0.038	0.011	0.034
Ar	2.41	15.29	16.18	16.22	16.54	10.50	3.24	2.89	7.65	5.84	8.87	14.16	14.62	1.07	12.23
Ab	96.74	83.80	82.66	82.33	82.09	89.07	96.25	96.46	90.91	92.78	90.36	85.26	84.44	98.63	86.93
Or	0.04	0.91	1.17	1.44	1.36	0.43	0.51	0.65	1.44	1.33	0.77	0.58	0.96	0.29	0.83

45, 48, 49, 50, 53, 54, 55, 56, 59: Plagioclase, isolated grains.

46: Rimmed crystal, along rim; 47: Same as 46, centre; 51: Rimmed crystal, centre; 52: Same as 51, rim; 57: Rimmed crystal, centre; 58: Same as 57, rim.

*Plagioclase in type B "transitional rocks".

Table 1-5 Analysis of amphiboles in type A "transitional rocks"

Sample no.	1 8801 18.06- 18.10a	2	3	4	5	6 8801 29.07- 29.15a	7	8	9	10	11 8801 208.92- 209.00a	12	13 8801 160.00- 160.18a	14 8801 107.00- 107.10a	15	16 8801 12.63- 12.67a
SiO ₂ (wt)	58.29	57.75	58.00	58.06	58.09	57.38	57.71	57.25	57.33	57.32	56.91	57.66	60.31	61.92	61.36	58.27
Al ₂ O ₃	12.05	12.83	12.29	12.18	12.12	11.97	12.84	12.46	12.43	12.85	10.96	11.63	10.63	9.20	9.68	10.71
FeO*	31.27	31.97	31.61	32.30	32.46	32.67	32.36	32.25	32.67	32.63	33.69	32.27	22.99	23.96	23.67	33.00
Fe ₂ O ₃ *	29.76	28.67	28.69	29.62	28.85	28.98	27.89	28.18	28.75	27.68	28.62	28.86	19.11	20.67	20.53	29.36
Fe ₂ O ₃ †	2.65	3.98	5.25	3.19	6.03	6.56	6.76	6.52	6.35	5.50	5.68	3.79	3.87	3.56	3.69	3.92
MgO	0.36	0.29	0.30	0.36	0.36	0.27	0.27	0.21	0.27	0.23	0.29	0.61	6.00	6.57	6.90	0.31
MnO	0.63	0.62	0.57	0.50	0.63	0.61	0.69	0.63	0.66	0.55	0.56	0.66	1.27	0.67	0.66	0.71
TiO	0.78	0.71	0.92	0.66	0.95	0.96	0.95	0.77	0.99	0.99	0.50	0.69	0.39	0.36	0.35	0.93
CaO	10.55	10.62	10.60	10.62	10.31	10.81	10.69	10.86	10.43	10.65	10.28	10.69	10.69	10.96	10.37	10.62
Na ₂ O	1.85	1.51	1.53	1.78	1.70	1.52	1.10	1.60	1.60	1.36	1.91	1.72	1.66	1.66	2.11	1.63
K ₂ O	2.06	2.16	2.16	2.10	1.82	2.10	2.20	2.19	2.70	2.10	2.10	2.13	1.60	1.61	1.57	2.06
Total	98.79	98.87	100.11	99.96	98.66	98.62	98.92	98.07	99.25	98.93	97.68	97.66	95.33	96.72	96.48	98.66
Number of ions on the basis of 23 oxygens																
Si	6.196	6.087	6.0612	6.156	6.169	6.075	6.066	6.061	6.031	6.016	6.099	6.163	6.427	6.566	6.508	6.236
Al	1.906	1.913	1.939	1.966	1.951	1.925	1.956	1.939	1.969	1.996	1.902	1.957	1.573	1.611	1.692	1.770
Al	0.692	0.525	0.372	0.676	0.655	0.369	0.5022	0.489	0.636	0.655	0.216	0.392	0.625	0.293	0.305	0.299
Fe ³⁺	4.027	3.939	3.827	3.979	3.895	3.985	3.751	3.835	3.995	3.730	3.956	3.960	2.566	2.717	2.703	4.026
Fe ²⁺	0.298	0.671	0.630	0.389	0.690	0.556	0.576	0.553	0.529	0.667	0.701	0.668	0.666	0.619	0.613	0.668
Mg	0.082	0.069	0.071	0.092	0.087	0.065	0.065	0.053	0.065	0.055	0.071	0.100	1.626	1.539	1.595	0.075
Mn	0.086	0.085	0.077	0.101	0.099	0.086	0.087	0.099	0.087	0.075	0.081	0.087	0.171	0.061	0.069	0.089
Ti	0.095	0.086	0.110	0.080	0.115	0.115	0.115	0.096	0.108	0.108	0.062	0.085	0.067	0.061	0.061	0.116
Exc. Oct.	0.079	0.077	0.088	0.075	0.101	0.095	0.076	0.082	0.091	0.091	0.086	0.075	0.081	0.073	0.117	0.069
Ca	1.629	1.635	1.612	1.660	1.733	1.661	1.662	1.626	1.605	1.606	1.620	1.666	1.626	1.662	1.769	1.652
Na ⁺	0.091	0.088	0.100	0.085	0.116	0.063	0.086	0.096	0.106	0.106	0.096	0.083	0.093	0.096	0.136	0.079
Na ⁺	0.699	0.366	0.373	0.673	0.616	0.616	0.259	0.367	0.315	0.320	0.516	0.666	0.356	0.615	0.510	0.616
K	0.629	0.666	0.666	0.632	0.375	0.635	0.655	0.656	0.556	0.632	0.663	0.666	0.325	0.293	0.315	0.629
Fe ³⁺ /Mg	55.91	63.07	61.56	56.16	51.30	69.17	67.66	67.33	68.92	69.96	66.32	66.93	2.33	2.08	1.99	61.33
Fe ³⁺ /Fe ²⁺	0.99	0.99	0.99	0.99	0.99	0.99	0.99	0.99	0.99	0.99	0.99	0.99	0.69	0.67	0.67	0.99

1,3,4-16: Hastingsite 1,2,8: Fe-pargasite

*Calculated according to Rasmussen et al. (1976)

Fe³⁺ total iron.

Table 1-1. Analyses of biotites in "transitional rocks".

Sample no.	1	2	3	4	5	6	7	8
	MO 39			MO 01	MO 01		AMG	MO 39
	49.51			29.15m	16.10m	16mm	MO 39m	52.63
	49.57m	16mm	16mm					52.67m
SiO ₂ (wt)	35.61	35.10	35.68	33.61	36.06	35.00	36.07	32.71
TiO ₂	2.26	2.11	2.18	2.50	2.82	2.26	2.20	1.26
Al ₂ O ₃	15.90	15.96	16.28	16.48	16.68	15.70	16.62	16.25
Cr ₂ O ₃	0.00	0.05	0.00	0.09	0.00	0.00	0.00	0.07
FeO	36.95	36.71	36.09	35.51	35.70	36.26	36.62	35.75
MnO	0.32	0.29	0.31	0.33	0.43	0.56	0.44	0.47
MgO	1.76	1.66	1.83	0.46	0.56	0.55	7.73	0.81
CaO	0.00	0.00	0.00	0.00	0.06	0.06	0.00	0.02
Na ₂ O	0.09	0.20	0.16	0.00	0.06	0.19	0.18	0.00
K ₂ O	9.36	9.06	9.15	9.01	9.03	9.16	10.09	9.09
Total	90.21	97.10	97.66	95.99	97.78	98.46	96.70	96.43

Ions on the basis of 22 oxygens									
Si	5.706	5.686	5.715	5.555	5.627	5.678	5.709	5.396	
Al	2.796	2.396	2.295	2.445	2.373	2.322	2.291	2.606	
Al	0.328	0.351	0.407	0.376	0.421	0.412	0.360	0.556	
Ti	0.272	0.257	0.261	0.311	0.296	0.267	0.257	0.156	
Mg	0.415	0.396	0.437	0.113	0.135	0.130	1.786	0.101	
Fe	4.682	4.703	4.567	4.908	4.822	4.541	1.192	4.912	
Mn	0.043	0.040	0.042	0.046	0.059	0.075	0.056	0.066	
Cr	0.000	0.006	0.000	0.012	0.000	0.000	0.000	0.009	
Mg	0.028	0.063	0.050	0.000	0.019	0.056	0.056	0.000	
Ca	0.000	0.000	0.000	0.000	0.010	0.007	0.000	0.006	
K	1.908	1.868	1.869	1.899	1.860	1.869	1.995	2.081	
Fe ²⁺ /Mg	11.37	11.97	10.55	43.72	36.20	15.51	1.42	49.57	
Fe ²⁺ /Fe ³⁺	0.92	0.92	0.91	0.96	0.97	0.97	0.66	0.98	

FeO calculated as total iron (Fe²⁺)
1-3: Biotite as only mafic mineral

4-7: Biotite from amphibole

Table I-8. Analyses of clinopyroxenes replacing hastingsite in type 9 "transitional rocks".

Sample no.	1 RR01 129.95- 129.00m	2 Idem	3 Idem	4 Idem	5 Idem	6 RR01 107.00- 107.10m	7 RR01	8 Idem	9 Idem	10 M13 HAV 791	11 Idem	12 Idem
SiO ₂ (wt)	49.96	49.78	49.64	49.07	47.25	51.71	50.01	50.47	49.53	48.11	46.20	46.61
TiO ₂	0.10	0.15	0.11	0.14	0.14	0.10	0.14	0.06	0.01	0.05	0.10	0.10
Al ₂ O ₃	0.84	1.58	1.25	0.57	1.56	1.35	1.96	1.33	1.59	0.23	1.16	1.24
FeO ^c	20.20	23.46	23.78	19.53	24.00	18.12	19.94	18.38	20.14	27.04	27.70	27.13
FeO ⁺	19.92	23.46	23.78	17.46	20.85	15.70	19.78	18.37	16.99	25.85	22.49	22.08
Fe ₂ O ₃ ^a	0.31	0.00	0.00	1.96	3.50	2.69	0.11	0.01	3.49	1.32	5.79	5.61
MnO	1.04	1.13	1.37	1.02	1.14	0.71	0.43	0.46	0.45	1.27	0.66	0.73
MgO	4.75	2.41	2.14	5.16	2.39	6.07	5.33	5.63	5.93	0.37	0.21	0.24
CaO	22.29	20.46	20.21	21.91	20.90	14.12	20.43	21.26	21.18	23.38	21.99	21.38
Na ₂ O	0.39	0.66	0.60	0.57	0.80	2.00	0.47	0.46	0.88	0.41	1.62	1.50
Cr ₂ O ₃	0.00	0.02	0.00	0.00	0.00	0.14	0.14	0.00	0.02	0.00	0.00	0.00
Total	99.55	99.65	98.90	97.97	98.08	98.32	99.19	98.47	99.73	99.99	100.12	99.49

Number of ions on the basis of 6 oxygens

Si	1.988	2.000	2.000	1.976	1.936	2.000	1.980	2.003	1.984	1.940	1.901	1.919
Al	0.012	0.000	0.000	0.024	0.064	0.000	0.020	0.000	0.056	0.011	0.000	0.060
Al	0.027	0.075	0.060	0.003	0.011	0.062	0.071	0.062	0.018	0.000	0.000	0.000
Fe ²⁺	0.663	0.788	0.908	0.601	0.714	0.514	0.655	0.610	0.558	0.990	0.774	0.761
Fe ³⁺	0.00	0.000	0.000	0.056	0.108	0.079	0.003	0.000	0.163	0.041	0.179	0.174
Mg	0.252	0.144	0.130	0.310	0.146	0.354	0.314	0.333	0.347	0.023	0.013	0.015
Mn	0.035	0.038	0.047	0.035	0.040	0.024	0.014	0.016	0.015	1.044	0.023	0.025
Ti	0.003	0.05	0.003	0.004	0.004	0.003	0.004	0.002	0.000	0.002	0.003	0.003
Cr	0.000	0.001	0.000	0.000	0.000	0.004	0.004	0.000	0.001	0.000	0.000	0.000
Exc. Oct.	0.019	0.051	0.046	0.010	0.023	0.040	0.067	0.023	0.042	0.000	0.000	0.000
Ca	0.991	0.881	0.880	0.946	0.913	0.760	0.967	0.904	0.891	0.947	0.965	0.943
Na	0.030	0.051	0.047	0.045	0.044	0.152	0.067	0.066	0.067	0.033	0.129	0.120
En%	14.87	7.96	7.13	16.68	9.23	21.75	17.13	14.03	19.32	1.19	0.74	0.96
Fs%	34.38	43.47	44.44	32.39	40.28	31.57	35.47	33.01	31.07	46.45	44.17	44.25
Wo%	50.75	48.57	48.41	50.92	51.49	46.68	47.20	48.95	49.61	51.96	55.09	54.99
Actite%	-	-	-	6.40	-	15.14	-	-	12.23	4.23	18.24	17.69
Jadite%	3.94	4.00	6.32	-	-	-	9.30	6.42	-	-	-	-
Cal. Cats%	-	-	-	-	12.34	-	-	-	-	-	-	-
Name	Fe-salite	Meden-salite	Meden-salite	Fe-salite	Meden-salite	Fe-salite	Fe-salite	Fe-salite	Fe-salite	Meden-salite	Idem	Idem

1: centre of grain; 2,3: same, border; 4: centre of grain; 5: same, border

^aCalculated according to Papike et al. (1974).

FeO⁺: total iron as FeO.

Table 1-9. Analyses of garnets from type B "transitional rocks".

Anal. no.	Centre										Rim			
	1	2	3	4	Average	5	6	7	8	Interm. Average	9	10	11	Average
SiO ₂ (wt%)	36.89	36.99	37.02	37.11	36.95	37.66	36.66	37.25	36.88	37.11	35.80	36.02	35.91	35.92
TiO ₂	0.05	0.06	0.06	0.07	0.05	0.01	0.11	0.01	0.08	0.05	0.23	0.18	0.26	0.22
Al ₂ O ₃	20.09	20.23	20.88	20.20	20.25	17.53	17.61	17.89	17.91	17.73	11.02	15.21	9.96	12.08
FeO ¹	27.43	26.61	26.37	26.57	26.74	19.97	19.65	21.81	19.86	20.22	21.24	18.52	21.83	20.53
FeO ₃ ²													9.11	
Fe ³⁺					28.06					18.19			11.16	
MnO	6.63	6.38	7.38	6.46	6.71	6.89	6.91	7.38	6.69	6.97	4.49	6.20	3.77	4.82
MgO	0.11	0.14	0.15	0.21	0.21	0.11	0.25	0.27	0.37	0.30	0.22	0.21	0.22	0.22
CaO	9.79	9.96	9.94	10.00	9.82	19.38	17.86	16.85	17.78	17.87	25.27	21.52	26.38	24.38
Na ₂ O	0.11	0.16	0.53	0.57	0.36		0.23		0.49	0.18	0.44	0.22	0.68	0.45
Total	100.72	100.51	101.73	101.84	101.07	101.78	99.88	101.05	100.06	100.83	98.78	98.13	98.95	98.62

Number of ions on the basis of 24 oxygens

Si	6.061	6.078	6.078	6.078	6.078	6.078	6.078	6.078	6.078	6.078	6.078	6.078	6.078
Ti	0.006	0.006	0.006	0.006	0.006	0.006	0.006	0.006	0.006	0.006	0.006	0.006	0.006
Al	1.901	1.901	1.901	1.901	1.901	1.901	1.901	1.901	1.901	1.901	1.901	1.901	1.901
Fe ²⁺	1.291	1.291	1.291	1.291	1.291	1.291	1.291	1.291	1.291	1.291	1.291	1.291	1.291
Fe ³⁺	0.929	0.929	0.929	0.929	0.929	0.929	0.929	0.929	0.929	0.929	0.929	0.929	0.929
Mn	1.771	1.771	1.771	1.771	1.771	1.771	1.771	1.771	1.771	1.771	1.771	1.771	1.771
Ca	0.108	0.108	0.108	0.108	0.108	0.108	0.108	0.108	0.108	0.108	0.108	0.108	0.108
Na	0.90	0.90	0.90	0.90	0.90	0.90	0.90	0.90	0.90	0.90	0.90	0.90	0.90
Pyrope %	16.81	16.81	16.81	16.81	16.81	16.81	16.81	16.81	16.81	16.81	16.81	16.81	16.81
Gross %	30.37	30.37	30.37	30.37	30.37	30.37	30.37	30.37	30.37	30.37	30.37	30.37	30.37
Alm. %	52.31	52.31	52.31	52.31	52.31	52.31	52.31	52.31	52.31	52.31	52.31	52.31	52.31

¹Calculated according to Barriaga (1981).

²FeO₃: total iron as FeO₃.

Sample BAA 463: 100 - zoned crystals; central part: anal. no. 1, 2, 3, 4; intermediate: anal. no. 5, 6, 7, 8; rim: anal. no. 9, 10, 11.

Table 1-9.
(Cont.)

Anal. no.	1	2	3	Average	1	2	Average
SiO ₂	36.01	36.04	36.16	36.14	35.88	35.95	35.91
TiO ₂	0.57	0.38	0.52	0.49	0.69	0.43	0.56
Al ₂ O ₃	7.67	7.29	7.60	7.50	6.40	6.09	6.24
Fe ₂ O ₃	25.21	25.27	25.26	25.25	25.84	25.37	25.60
Fe ₂ O ₃ *				16.46			19.75
FeO*				7.90			5.24
MnO	0.44	1.18	0.48	1.05	3.09	1.06	3.07
MgO	0.24	0.14	0.0	0.13	0.12	0.07	0.09
CaO	28.94	28.57	28.63	28.71	28.12	28.98	28.55
Mg ₂ O	0.76	0.52	0.00	0.43	0.00	0.02	0.01
Total	100.27	99.59	99.25	99.70	100.14	99.97	100.03

Number of ions on the basis of 26 oxygens

Si	6.034	5.996
Ti	0.061	0.070
Al	1.876	1.228
Fe ³⁺	2.069	2.467
Fe ²⁺	1.104	0.735
Mn	0.148	0.434
Mg	0.032	0.027
Ca	5.136	5.108
Mg	0.139	0.003
Andrad	56.168	66.89
Pyrope	0.618	0.40
Speress	2.794	7.80
Gross	34.248	25.00
Alm		

*Calculated according to Hurliga (1983)
Fe³⁺, total iron as Fe³⁺

Sample AM3 RAA 413

Sample AM9 HAA7903

Table I-9.
(Cont.)

Anal.	1	2	3	4	5	6	7	Average
SiO ₂ (wt)	36.63	35.94	36.32	36.59	36.64	36.89	35.95	36.42
TiO ₂	0.45	0.44	0.47	0.43	0.42	0.44	0.46	0.44
Al ₂ O ₃	11.10	10.13	11.28	9.83	12.08	11.13	10.73	10.90
Fe ₂ O ₃ ^t	21.42	23.18	22.37	23.53	21.40	22.57	23.10	22.57
Fe ₂ O ₃ ^a								10.38
FeO ^a								10.96
MnO	2.06	1.48	1.75	1.71	2.37	2.13	2.16	1.98
MgO	0.15	0.17	0.18	0.18	0.20	0.18	0.20	0.18
CaO	26.66	27.81	27.60	27.71	26.41	26.79	27.08	27.11
Na ₂ O	0.49	0.46	0.58	0.60	0.61	0.59	0.63	0.57
Total	99.36	99.62	100.55	100.10	100.11	100.71	100.28	100.17

Number of ions on the basis of 24 oxygens

Si	6.019
Ti	0.055
Al	2.123
Fe ³⁺	1.292
Fe ²⁺	1.515
Mn	0.277
Mg	0.044
Ca	4.401
Na	0.193
Andrad.	17.468
Pyrope	0.468
Sperss.	5.418
Grossul.	55.808
Almandine	0.068

^aCalculated according to Barriga (1993)Fe₂O₃^t: total iron as Fe₂O₃

Sample BR 01 128.95 - 129.00 m

Table I-10. Analyses of epidotes in transitional rocks.

	1	2	3	4	5	6	7	8	9
Sample no.	444	Idem	ANJ RAA413	Idem	Idem	Idem	Idem	ANJ RAA368	Idem
SiO ₂ (wt%)	38.23	38.14	36.12	36.30	36.47	35.57	36.54	36.79	36.34
TiO ₂	0.00	0.00	0.01	0.01	0.02	0.04	0.00	0.00	0.00
Al ₂ O ₃	24.09	24.16	21.73	21.85	22.00	21.95	22.80	23.39	22.80
Fe ₂ O ₃ *	14.45	14.37	15.69	15.94	15.69	16.18	16.12	16.04	16.37
MnO	0.07	0.08	0.06	0.08	0.10	0.00	0.12	0.32	0.49
MgO	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
CaO	23.45	23.21	20.78	20.77	21.16	20.48	22.95	23.47	23.02
Na ₂ O	0.00	0.10	0.00	0.00	0.00	0.00	0.00	0.12	0.00
Y ₂ O ₃	NA	NA	0.05	0.07	0.04	0.01	NA	NA	NA
UO ₂ ^a	"	"	0.19	0.23	0.04	0.13	"	"	"
La ₂ O ₃	"	"	0.95	1.04	0.83	1.26	"	"	"
Ce ₂ O ₃	"	"	2.03	2.13	1.88	2.47	"	"	"
Total	100.29	100.10	97.60	98.42	98.08	97.99	98.43	100.00	99.02

Ions on the basis of 25 oxygens

Si	5.953
Ti	0.000
Al	4.420
Fe ³⁺	1.693
Mn	0.009
Mg	0.000
Ca	3.912
Na	0.000

1,2: epidote (olstacite) from pyroxene; 3-6: brown epidote
 7: green epidote (olstacite) associated to garnet
 8,9: green epidote (olstacite) associated to garnet

*total iron as Fe₂O₃ NA: not analyzed

^a: oxidation state not determined; values reported as total oxide

Table 1.11. Analyses of plagioclases from metamorphic rocks.

Sample no.	1	2	3	4	5	6	7	8	9	10	11	12	13	14	15
	HR01	HR01	HR01	HR01	HR01	HR01	HR01	HR01	HR01	HR01	HR01	HR01	HR01	HR01	HR01
	104.75	104.75	105.80	105.80	105.80	105.80	105.80	105.80	105.80	105.80	105.80	105.80	105.80	105.80	105.80
	104.85m	104.85m	105.90m	105.90m	105.90m	105.90m	105.90m	105.90m	105.90m	105.90m	105.90m	105.90m	105.90m	105.90m	105.90m
	104.85m	104.85m	105.90m	105.90m	105.90m	105.90m	105.90m	105.90m	105.90m	105.90m	105.90m	105.90m	105.90m	105.90m	105.90m
	67.41	66.82	68.64	67.77	67.41	67.63	67.79	67.32	67.32	67.75	66.91	66.01	66.74	67.57	68.00
SiO ₂ (wt%)	19.52	19.27	19.45	20.06	20.00	20.56	20.79	20.30	20.85	21.13	21.46	20.76	21.09	19.85	19.74
Al ₂ O ₃	0.19	0.25	0.56	0.64	0.66	1.10	1.04	0.55	0.64	1.11	0.89	1.02	1.05	1.42	1.27
CaO	11.86	12.25	13.41	12.00	11.00	10.27	10.79	11.03	10.63	9.70	10.27	10.71	10.31	10.73	10.19
MgO	0.04	0.07	0.16	0.11	0.11	0.24	0.22	0.26	0.18	0.14	0.26	0.13	0.19	0.21	0.18
Total	98.99	98.66	100.76	100.64	99.19	99.80	100.11	99.46	99.62	99.87	99.79	98.63	99.34	99.60	99.36
Loss on the basis of 32 oxygens															
Si	11.917	11.486	11.971	11.922	11.877	11.812	11.763	11.818	11.797	11.810	11.710	11.717	11.734	11.868	11.939
Al	4.085	4.019	1.996	4.124	4.152	4.239	4.281	4.206	4.305	4.341	4.426	4.342	4.369	4.104	4.084
Ca	0.146	0.044	0.105	0.120	0.125	0.206	0.195	0.104	0.120	0.207	0.167	0.194	0.198	0.244	0.239
Mg	4.057	4.225	1.866	4.059	3.758	3.464	3.657	3.760	3.614	3.278	3.485	3.686	3.514	3.654	3.469
K	0.002	0.016	0.036	0.038	0.025	0.054	0.089	0.058	0.040	0.040	0.054	0.029	0.043	0.047	0.040
FeO	0.44	1.11	2.61	2.46	1.19	5.51	4.99	2.64	3.19	5.44	4.50	4.96	5.27	6.29	6.37
Na ₂ O	99.07	94.52	96.50	96.27	96.18	91.06	93.75	95.80	95.75	97.98	93.98	94.28	93.60	92.52	92.55
Sum	0.08	0.37	0.89	0.90	0.63	1.43	1.26	1.49	1.07	1.14	1.56	0.75	1.13	1.19	1.08

Table 1-11.6
(Cont.)

Sample no.	16	17	18	19	20	21	22	23	24	25	26	27	28	29	30
	447	1143-2	1444	1445	1446	1447	1448	1449	1450	1451	1452	1453	1454	1455	1456
SiO ₂ (wt%)	69.07	68.09	68.20	67.77	67.81	66.31	67.16	66.77	66.15	67.00	67.11	69.09	68.06	68.87	69.01
Al ₂ O ₃	19.44	19.31	19.66	20.06	20.07	21.30	20.79	20.86	21.65	20.91	20.36	20.00	19.89	19.82	19.54
CaO	0.20	0.60	0.51	0.64	0.66	0.89	1.17	1.16	1.11	1.85	1.27	0.65	0.44	0.89	0.30
MgO	11.29	11.79	11.57	12.00	11.00	10.65	10.45	10.68	11.22	10.74	11.16	11.64	11.45	10.20	11.31
K ₂ O	0.09	0.28	0.16	0.17	0.11	0.24	0.23	0.23	0.16	0.29	0.20	0.11	0.01	0.11	0.09
Total	99.61	100.07	100.10	100.64	99.18	99.79	99.82	99.92	100.29	100.91	100.10	100.49	100.25	99.94	100.31

	Loss on the basis of 12 oxygen														
Si	11.966	11.916	11.924	11.922	11.877	11.612	11.770	11.713	11.541	11.704	11.766	11.864	11.845	11.946	12.014
Al	4.029	3.994	4.050	4.124	4.152	4.486	4.292	4.312	4.467	4.306	4.206	4.104	4.093	4.074	4.012
Ca	0.0432	0.113	0.096	0.120	0.125	0.163	0.220	0.259	0.208	0.271	0.289	0.421	0.082	0.166	0.056
Mg	3.962	4.006	3.922	4.059	3.754	3.622	3.550	3.433	3.809	3.697	3.793	3.914	4.012	3.222	3.813
K	0.020	0.063	0.036	0.034	0.025	0.054	0.051	0.051	0.036	0.065	0.065	0.024	0.002	0.066	0.020
Na	3.13	2.69	2.36	2.44	1.19	4.35	5.75	6.58	5.14	6.41	5.85	1.09	2.01	4.37	1.44
Fe	96.20	95.81	96.76	96.27	96.44	94.25	92.91	92.12	91.98	91.54	91.05	90.60	90.74	90.79	90.05
Loss	0.30	1.30	0.84	0.90	0.63	1.40	1.35	1.31	0.84	1.63	1.30	0.51	0.05	0.64	0.51

Table 1-11.
(Cont.)

Sample no.	Si	Al ^{IV}	Al ^{VI}	Fe	Mg	Ca	Na	K	Sum	SiO ₂	Al ₂ O ₃	FeO	MgO	CaO	Na ₂ O	K ₂ O	Total	Si	Al ^{IV}	Al ^{VI}	Fe	Mg	Ca	Na	K	Sum								
105.6A	66.85	67.08	66.71	66.91	66.96	68.61	67.79	68.01	66.73	68.67	69.11	68.08	68.60	62.36	63.86	65.69	66.77	66.85	67.08	66.71	66.91	66.96	68.61	67.79	68.01	66.73	68.67	69.11	68.08	68.60	62.36	63.86	65.69	66.77
105.7A	20.21	21.36	21.47	20.03	20.52	19.88	19.81	19.52	20.02	19.38	19.83	22.80	22.16	23.30	22.98	21.36	21.96	20.21	21.36	21.47	20.03	20.52	19.88	19.81	19.52	20.02	19.38	19.83	22.80	22.16	23.30	22.98	21.36	21.96
105.7B	1.05	1.52	1.11	1.66	1.43	1.01	1.46	1.41	1.62	0.96	0.74	1.79	1.92	1.86	8.83	3.17	1.12	1.05	1.09	10.60	10.53	11.21	11.28	11.30	11.41	11.05	11.66	11.11	9.86	9.85	9.59	9.35	9.16	
105.7C	9.91	10.89	10.40	10.53	11.21	11.28	11.30	11.41	11.05	11.66	11.11	9.86	9.85	9.59	9.35	9.16	9.21	9.91	10.89	10.40	10.53	11.21	11.28	11.30	11.41	11.05	11.66	11.11	9.86	9.85	9.59	9.35	9.16	
105.7D	0.21	0.22	0.19	0.20	0.27	0.26	0.20	0.21	0.13	0.16	0.45	0.16	0.18	0.17	0.13	0.28	0.21	0.21	0.22	0.19	0.20	0.27	0.26	0.20	0.21	0.13	0.16	0.45	0.16	0.18	0.17	0.13	0.28	0.21
Total	99.85	100.55	100.10	99.11	100.37	100.66	100.58	100.79	99.60	100.69	100.60	100.17	100.51	99.08	100.59	99.85	99.20	99.85	100.55	100.10	99.11	100.37	100.66	100.58	100.79	99.60	100.69	100.60	100.17	100.51	99.08	100.59	99.85	99.20

Ions on the basis of 12 oxygens

Si	11.96	11.68	11.66	11.82	11.72	11.98	11.83	11.88	11.76	11.98	12.00	11.10	11.11	11.16	11.17	11.56	11.67
Al	4.157	4.140	4.123	4.120	4.124	3.995	4.000	4.023	4.170	3.976	3.975	4.685	4.595	4.663	4.770	4.811	4.842
Ca	0.196	0.208	0.203	0.216	0.268	0.188	0.271	0.268	0.306	0.152	0.185	0.717	0.739	0.735	0.836	0.998	0.992
Mg	3.580	3.508	3.525	3.607	3.406	3.806	3.825	3.777	3.778	3.934	3.787	3.178	3.361	3.366	3.275	3.191	3.119
K	0.087	0.089	0.082	0.085	0.080	0.082	0.083	0.087	0.089	0.031	0.033	0.032	0.080	0.027	0.029	0.063	0.067
Fe	5.86	7.38	6.88	7.08	6.69	8.68	6.59	6.66	7.46	3.80	3.70	17.39	17.85	17.89	20.19	15.52	15.67
Al ^{IV}	91.28	91.36	92.85	91.82	92.05	93.81	92.31	92.89	91.85	95.85	95.86	81.85	81.17	81.86	79.10	82.85	83.07
Al ^{VI}	1.30	1.27	1.11	1.15	1.46	1.53	1.08	1.15	0.71	0.75	0.85	0.76	0.98	0.62	0.71	1.63	1.76

1,2,8,9: albite in pyrrhotite-albite (mineralized)
 3,7,19,21: albite in pyrrhotite-albite
 10,11,27,29: biotite-albite
 14,15: microcline-pyrrhotite
 16,18: Ca-rich-albite (pyrrhotite-albite)
 22,24: quartz-pyrrhotite
 25,26: pyrrhotite-albite containing prepyrrhotite (mineralized)

10: amorphous-albite (mineralized)
 31-41: magnetite-albite
 42-45: type 1 epidote
 46,47: type 2 epidote

Table 1-12 Analysis of chloropyroxene in metamorphic rocks.

Sample no.	1	2	3	4	5	6	7	8	9	10	11	12	13	14	15
	BR01 105.70- 106.70s	105s	105s	105s	105s	105s	BR01 105.62- 105.70s	105s	105s	105s	105s	BR01 106.00- 106.10s	106s	BR01 106.90- 107.00s	107s
SiO ₂ (wt)	48.89	49.29	49.30	49.26	48.25	48.87	47.98	48.70	49.10	49.14	47.70	50.90	50.39	50.00	49.93
TiO ₂	0.05	0.08	0.02	0.11	0.05	0.02	0.07	0.05	0.08	0.05	0.07	0.08	0.06	0.08	0.02
Al ₂ O ₃	1.52	1.33	2.91	1.67	1.35	1.36	1.33	1.39	1.52	1.11	1.31	0.63	1.38	1.31	1.29
FeO ¹	23.07	21.51	21.69	21.71	22.95	23.03	23.60	24.30	24.73	22.56	25.07	16.20	20.41	19.55	21.16
FeO ²	16.08	15.11	12.16	16.68	16.83	16.63	16.63	17.23	15.08	17.09	17.67	13.62	16.89	16.65	16.01
FeO ³	7.81	7.11	10.56	7.81	8.16	7.33	3.77	7.85	10.56	6.17	8.65	5.31	6.13	5.67	5.72
MgO	0.73	0.71	0.89	0.67	0.80	0.39	0.86	0.30	0.30	0.66	0.69	0.36	0.66	0.49	0.37
MgO	6.16	6.77	6.16	6.95	6.32	6.12	3.77	2.97	3.03	6.60	3.18	7.02	5.25	6.16	5.08
CaO	19.14	18.96	19.13	19.38	19.11	18.89	19.13	17.88	18.16	20.36	18.77	20.76	19.31	20.67	20.66
MnO	2.20	2.26	2.38	2.25	1.32	2.26	2.28	2.66	1.36	1.79	2.99	1.73	2.38	1.67	1.71
Cr ₂ O ₃	0.13	0.15	0.09	0.07	0.11	0.07	0.07	0.16	0.06	0.08	0.13	0.05	0.08	0.00	0.12
Total	103.47	99.75	99.61	100.61	98.07	99.65	99.65	99.60	101.56	100.81	100.31	100.20	100.29	100.28	100.59
Number of ions on the basis of 6 oxygens															
Si	1.925	1.966	1.895	1.928	1.953	1.959	1.921	1.968	1.920	1.956	1.908	1.971	1.988	1.968	1.969
Al	0.077	0.096	0.137	0.088	0.067	0.061	0.063	0.052	0.070	0.052	0.062	0.029	0.056	0.052	0.051
Al	0.000	0.008	0.008	0.000	0.016	0.011	0.000	0.016	0.000	0.000	0.000	0.000	0.027	0.008	0.009
Fe ¹	0.528	0.489	0.408	0.461	0.437	0.585	0.557	0.577	0.492	0.561	0.568	0.435	0.485	0.471	0.526
Fe ²	0.232	0.211	0.317	0.230	0.126	0.219	0.233	0.236	0.377	0.183	0.256	0.155	0.180	0.166	0.169
Mg	0.263	0.280	0.366	0.286	0.261	0.266	0.225	0.177	0.177	0.258	0.190	0.405	0.305	0.356	0.286
Mn	0.026	0.026	0.023	0.022	0.011	0.011	0.015	0.017	0.017	0.022	0.022	0.011	0.015	0.016	0.012
Ti	0.002	0.002	0.001	0.005	0.002	0.001	0.002	0.002	0.002	0.001	0.002	0.001	0.002	0.002	0.001
Cr	0.009	0.009	0.009	0.002	0.009	0.002	0.002	0.009	0.002	0.002	0.002	0.002	0.002	0.002	0.008
Enrich.	0.033	0.027	0.115	0.025	0.067	0.033	0.039	0.027	0.006	0.008	0.026	0.009	0.017	0.020	0.011
Ca	0.808	0.801	0.817	0.813	0.829	0.795	0.821	0.766	0.780	0.859	0.826	0.869	0.826	0.856	0.856
Mg	0.168	0.171	0.18	0.171	0.108	0.172	0.177	0.206	0.253	0.137	0.201	0.130	0.177	0.126	0.131
Enrich.	15.38	17.76	23.03	16.13	15.09	15.38	16.06	11.63	12.36	15.39	12.01	23.61	19.40	21.20	17.56
Fe	13.66	11.55	25.55	30.63	36.91	36.66	36.75	37.91	36.66	33.42	37.02	25.55	30.60	27.99	31.27
Mg	51.18	50.71	51.42	51.66	48.00	50.48	51.21	50.42	51.20	51.19	50.97	50.63	50.50	50.91	51.19
Total	23.70	22.45	32.00	23.53	16.88	23.27	23.77	25.67	32.10	16.68	26.05	15.75	15.99	17.62	16.15
Note	Aspirine- Apite	Aspirine- Apite	Aspirine- Apite	Aspirine- Apite	Aspirine- Apite	Aspirine- Apite	Aspirine- Apite	Aspirine- Apite	Aspirine- Apite	Aspirine- Apite	Aspirine- Apite	Aspirine- Apite	Aspirine- Apite	Aspirine- Apite	Aspirine- Apite

1: center of grain; 2: side, border; 3: center of grain; border; 4: 15: garnet-cyanite-magnetite

* Calculated according to Papias et al. (1976)

Fe¹: total iron as FeO

Table 1-12

(Cont.)

Sample no.	16 BRO1 109.90- 109.00s	17 109.90- 109.00s	18 109.90- 109.00s	19 109.90- 109.00s	20 109.90- 109.00s	21 109.90- 109.00s	22 109.90- 109.00s	23 109.90- 109.00s	24 109.90- 109.00s	25 109.90- 109.00s	26 109.90- 109.00s	27 BRO1 179.90- 180.00s	28 109.90- 109.00s	29 109.90- 109.00s	30 109.90- 109.00s	31 109.90- 109.00s
SiO ₂ (wt)	50.33	51.50	51.30	51.29	50.97	50.03	52.34	51.35	52.66	51.25	51.69	50.13	50.36	49.95	49.91	49.76
TiO ₂	0.00	0.00	0.03	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.07	0.00	0.01	0.00
Al ₂ O ₃	1.15	2.36	2.53	2.61	2.20	2.66	2.09	2.25	1.96	2.61	1.95	1.86	1.28	1.36	1.21	0.77
FeO ^a	16.83	15.16	14.63	13.56	15.26	16.52	16.52	16.52	13.37	9.96	11.68	21.01	21.56	21.21	20.09	19.21
Fe ²⁺	13.51	9.31	6.46	5.99	5.53	5.65	7.91	9.79	6.26	6.56	5.18	16.65	16.31	16.90	13.11	16.78
Fe ³⁺	3.31	5.85	8.17	7.57	9.73	10.87	8.61	6.76	7.10	6.05	6.52	4.36	5.25	4.31	6.98	2.43
MnO	0.65	0.18	0.29	0.19	0.24	0.17	0.17	0.26	0.19	0.19	0.15	0.79	0.63	0.72	0.62	1.13
MgO	6.79	9.50	9.61	9.16	9.07	9.26	9.30	9.27	9.25	10.66	9.67	5.66	5.11	5.51	5.29	6.92
CaO	20.09	18.83	18.69	19.38	19.02	19.71	18.71	18.61	19.36	19.75	19.60	18.82	18.39	19.36	19.75	22.17
Na ₂ O	1.96	3.59	3.30	2.98	2.91	2.97	3.30	2.96	2.88	2.82	3.11	2.36	2.73	2.05	2.38	0.63
Cr ₂ O ₃	0.00	0.00	0.05	0.00	0.02	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
Total	99.79	101.26	100.36	100.03	99.52	98.99	100.61	99.69	100.31	99.69	98.31	101.36	100.96	100.96	100.26	100.97

Number of ions on the basis of 6 oxygens

Si	1.961	1.929	1.936	1.935	1.968	1.955	1.971	1.976	1.978	1.988	1.969	1.939	1.956	1.962	1.964	1.962
Al	0.000	0.072	0.066	0.065	0.062	0.065	0.029	0.029	0.022	0.012	0.031	0.061	0.066	0.058	0.052	0.035
Al	0.016	0.022	0.067	0.062	0.067	0.067	0.063	0.078	0.065	0.086	0.052	0.026	0.012	0.009	0.009	0.000
Fe ²⁺	0.660	0.466	0.206	0.189	0.273	0.271	0.269	0.362	0.259	0.205	0.165	0.667	0.668	0.666	0.626	0.666
Fe ³⁺	0.173	0.308	0.257	0.238	0.215	0.216	0.208	0.160	0.161	0.107	0.207	0.212	0.236	0.205	0.228	0.081
Mg	0.371	0.676	0.666	0.515	0.660	0.672	0.670	0.670	0.516	0.563	0.569	0.315	0.295	0.319	0.308	0.603
Mn	0.015	0.012	0.009	0.012	0.008	0.006	0.005	0.008	0.006	0.006	0.005	0.006	0.002	0.002	0.007	0.037
Ti	0.000	0.001	0.001	0.001	0.002	0.002	0.002	0.003	0.003	0.006	0.001	0.000	0.002	0.001	0.000	0.000
Cr	0.000	0.000	0.001	0.000	0.001	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.002	0.010	0.002
Sum, Oct.	0.011	0.000	0.000	0.000	0.005	0.011	0.000	0.010	0.012	0.006	0.000	0.066	0.030	0.039	0.000	0.071
Ca	0.979	0.755	0.755	0.796	0.779	0.768	0.755	0.751	0.779	0.79	0.792	0.760	0.766	0.806	0.826	0.927
Na	0.168	0.261	0.261	0.216	0.216	0.221	0.267	0.219	0.210	0.206	0.260	0.176	0.205	0.156	0.160	0.066
Sum	22.69	13.97	13.52	16.62	16.62	17.22	17.90	16.62	13.30	16.95	16.65	20.15	19.38	19.63	19.70	21.66
Fe ²⁺	26.68	11.91	16.16	12.72	16.06	17.91	16.99	20.16	16.66	12.96	10.96	29.92	10.67	10.09	27.61	19.16
Fe ³⁺	50.91	56.11	52.12	52.66	51.56	50.95	51.21	49.71	50.06	50.77	52.54	49.93	50.15	50.98	52.69	46.16
Sum, Oct.	16.71	13.17	10.61	29.16	26.52	26.15	23.19	16.05	22.69	11.25	25.96	23.63	26.96	21.16	23.17	9.11
Na	Fe ²⁺ Fe ³⁺ Fe ²⁺ Fe ³⁺ Fe ²⁺ Fe ³⁺ Fe ²⁺ Fe ³⁺ Fe ²⁺ Fe ³⁺ Fe ²⁺ Fe ³⁺ Fe ²⁺ Fe ³⁺ Fe ²⁺ Fe ³⁺	Fe ²⁺ Fe ³⁺ Fe ²⁺ Fe ³⁺ Fe ²⁺ Fe ³⁺ Fe ²⁺ Fe ³⁺ Fe ²⁺ Fe ³⁺ Fe ²⁺ Fe ³⁺ Fe ²⁺ Fe ³⁺ Fe ²⁺ Fe ³⁺	Fe ²⁺ Fe ³⁺ Fe ²⁺ Fe ³⁺ Fe ²⁺ Fe ³⁺ Fe ²⁺ Fe ³⁺ Fe ²⁺ Fe ³⁺ Fe ²⁺ Fe ³⁺ Fe ²⁺ Fe ³⁺ Fe ²⁺ Fe ³⁺	Fe ²⁺ Fe ³⁺ Fe ²⁺ Fe ³⁺ Fe ²⁺ Fe ³⁺ Fe ²⁺ Fe ³⁺ Fe ²⁺ Fe ³⁺ Fe ²⁺ Fe ³⁺ Fe ²⁺ Fe ³⁺ Fe ²⁺ Fe ³⁺	Fe ²⁺ Fe ³⁺ Fe ²⁺ Fe ³⁺ Fe ²⁺ Fe ³⁺ Fe ²⁺ Fe ³⁺ Fe ²⁺ Fe ³⁺ Fe ²⁺ Fe ³⁺ Fe ²⁺ Fe ³⁺ Fe ²⁺ Fe ³⁺	Fe ²⁺ Fe ³⁺ Fe ²⁺ Fe ³⁺ Fe ²⁺ Fe ³⁺ Fe ²⁺ Fe ³⁺ Fe ²⁺ Fe ³⁺ Fe ²⁺ Fe ³⁺ Fe ²⁺ Fe ³⁺ Fe ²⁺ Fe ³⁺	Fe ²⁺ Fe ³⁺ Fe ²⁺ Fe ³⁺ Fe ²⁺ Fe ³⁺ Fe ²⁺ Fe ³⁺ Fe ²⁺ Fe ³⁺ Fe ²⁺ Fe ³⁺ Fe ²⁺ Fe ³⁺ Fe ²⁺ Fe ³⁺	Fe ²⁺ Fe ³⁺ Fe ²⁺ Fe ³⁺ Fe ²⁺ Fe ³⁺ Fe ²⁺ Fe ³⁺ Fe ²⁺ Fe ³⁺ Fe ²⁺ Fe ³⁺ Fe ²⁺ Fe ³⁺ Fe ²⁺ Fe ³⁺	Fe ²⁺ Fe ³⁺ Fe ²⁺ Fe ³⁺ Fe ²⁺ Fe ³⁺ Fe ²⁺ Fe ³⁺ Fe ²⁺ Fe ³⁺ Fe ²⁺ Fe ³⁺ Fe ²⁺ Fe ³⁺ Fe ²⁺ Fe ³⁺	Fe ²⁺ Fe ³⁺ Fe ²⁺ Fe ³⁺ Fe ²⁺ Fe ³⁺ Fe ²⁺ Fe ³⁺ Fe ²⁺ Fe ³⁺ Fe ²⁺ Fe ³⁺ Fe ²⁺ Fe ³⁺ Fe ²⁺ Fe ³⁺	Fe ²⁺ Fe ³⁺ Fe ²⁺ Fe ³⁺ Fe ²⁺ Fe ³⁺ Fe ²⁺ Fe ³⁺ Fe ²⁺ Fe ³⁺ Fe ²⁺ Fe ³⁺ Fe ²⁺ Fe ³⁺ Fe ²⁺ Fe ³⁺	Fe ²⁺ Fe ³⁺ Fe ²⁺ Fe ³⁺ Fe ²⁺ Fe ³⁺ Fe ²⁺ Fe ³⁺ Fe ²⁺ Fe ³⁺ Fe ²⁺ Fe ³⁺ Fe ²⁺ Fe ³⁺ Fe ²⁺ Fe ³⁺	Fe ²⁺ Fe ³⁺ Fe ²⁺ Fe ³⁺ Fe ²⁺ Fe ³⁺ Fe ²⁺ Fe ³⁺ Fe ²⁺ Fe ³⁺ Fe ²⁺ Fe ³⁺ Fe ²⁺ Fe ³⁺ Fe ²⁺ Fe ³⁺	Fe ²⁺ Fe ³⁺ Fe ²⁺ Fe ³⁺ Fe ²⁺ Fe ³⁺ Fe ²⁺ Fe ³⁺ Fe ²⁺ Fe ³⁺ Fe ²⁺ Fe ³⁺ Fe ²⁺ Fe ³⁺ Fe ²⁺ Fe ³⁺	Fe ²⁺ Fe ³⁺ Fe ²⁺ Fe ³⁺ Fe ²⁺ Fe ³⁺ Fe ²⁺ Fe ³⁺ Fe ²⁺ Fe ³⁺ Fe ²⁺ Fe ³⁺ Fe ²⁺ Fe ³⁺ Fe ²⁺ Fe ³⁺	Fe ²⁺ Fe ³⁺ Fe ²⁺ Fe ³⁺ Fe ²⁺ Fe ³⁺ Fe ²⁺ Fe ³⁺ Fe ²⁺ Fe ³⁺ Fe ²⁺ Fe ³⁺ Fe ²⁺ Fe ³⁺ Fe ²⁺ Fe ³⁺

17: border of grain; 19: same, intercrystalline; 19: same, centre; 20: another grain, border; 21: same, centre;
 23: another grain, border; 26: same, centre; 25: centre of grain; 26: same, border; 27: border of grain; 28: same,
 along border; 29: same, centre; 30: another grain, border; 31: same, centre; 16, 17, 25: garnet-orthopyroxene-ilmenite
 27: mineralized garnet-orthopyroxene-ilmenite

^a Calculated according to Perkins et al. (1976).
 FeO: total iron as FeO

Table 1-12.
(Cont.)

Sample no.	32 8801 185.10- 185.25a	33	34	35	36	37	38 8801 187.15- 187.25a	39	40	41	42	43	44	45 8801 187.80- 187.92a	46	47
	100%	100%	100%	100%	100%	100%	100%	100%	100%	100%	100%	100%	100%	100%	100%	100%
SiO ₂ (wt)	69.55	50.71	69.69	51.70	69.65	51.66	50.22	66.98	49.57	69.66	69.79	69.57	50.67	51.06	49.33	69.51
TiO ₂	0.10	0.01	0.10	0.06	0.06	0.08	0.02	0.00	0.00	0.00	0.06	0.06	0.06	0.06	0.05	0.06
Al ₂ O ₃	1.25	0.89	1.56	0.61	1.67	0.58	1.46	1.57	0.56	1.12	1.60	1.56	0.56	0.73	1.69	1.60
FeO*	21.68	16.77	22.02	16.66	23.06	17.63	19.92	20.67	19.92	20.03	21.07	22.15	16.37	19.06	22.55	18.27
Fe ₂ O ₃	17.66	16.09	17.02	13.86	15.90	13.73	11.30	11.48	13.57	11.32	13.10	12.96	13.76	14.22	16.36	6.61
Fe ₂ O ₃ ²	6.68	2.97	5.56	3.11	9.06	6.11	9.56	10.33	6.96	9.68	8.95	10.23	5.12	5.36	6.98	6.07
MnO	0.67	0.39	0.48	0.66	0.51	0.36	0.63	0.63	0.53	0.53	0.39	0.37	0.66	0.63	0.62	0.63
MgO	3.63	7.31	6.02	7.61	3.62	7.43	5.23	6.80	6.12	5.88	6.95	5.01	6.55	6.25	3.66	5.79
CaO	16.99	21.73	20.02	21.93	16.36	21.73	17.66	16.57	19.82	19.21	19.28	17.86	21.36	21.29	19.68	21.36
Mg ₂ Si	2.26	1.36	1.93	1.27	2.78	1.32	3.56	3.15	2.68	3.10	2.71	3.05	1.60	1.73	2.33	1.69
Cr ₂ O ₃	0.08	0.07	0.06	0.12	0.08	0.06	0.06	0.05	0.28	0.06	0.05	0.00	0.06	0.06	0.10	0.12
Total	98.66	99.32	100.22	100.69	100.19	100.62	99.50	99.05	100.15	100.36	100.59	100.59	100.00	101.17	100.08	99.12

Number of ions on the basis of 8 oxygens

Si	1.986	1.978	1.951	1.986	1.950	1.971	1.956	1.950	1.950	1.925	1.938	1.929	1.966	1.968	1.969	1.960
Al	0.016	0.022	0.049	0.016	0.050	0.026	0.062	0.070	0.025	0.051	0.062	0.071	0.036	0.032	0.051	0.060
Al	0.065	0.019	0.026	0.013	0.018	0.000	0.006	0.003	0.000	0.000	0.002	0.001	0.003	0.001	0.019	0.025
Fe ₂₊	0.991	0.46	0.561	0.665	0.521	0.666	0.368	0.368	0.666	0.369	0.627	0.621	0.668	0.656	0.560	0.661
Fe ₃₊	0.135	0.087	0.165	0.080	0.239	0.119	0.281	0.306	0.207	0.296	0.299	0.300	0.150	0.155	0.205	0.121
Mg	0.216	0.625	0.236	0.636	0.213	0.628	0.306	0.282	0.357	0.361	0.287	0.291	0.380	0.359	0.236	0.360
Mn	0.016	0.013	0.016	0.016	0.017	0.012	0.016	0.016	0.016	0.017	0.011	0.012	0.015	0.016	0.016	0.016
Ti	0.003	0.003	0.003	0.001	0.002	0.002	0.001	0.000	0.000	0.000	0.002	0.000	0.001	0.001	0.001	0.002
Cr	0.003	0.002	0.001	0.006	0.002	0.001	0.002	0.002	0.009	0.002	0.002	0.000	0.001	0.002	0.003	0.006
Encl. Cr	0.011	0.006	0.007	0.003	0.012	0.006	0.000	0.000	0.011	0.013	0.000	0.025	0.003	0.000	0.000	0.000
Ca	0.815	0.908	0.966	0.902	0.776	0.899	0.737	0.796	0.831	0.901	0.906	0.765	0.962	0.979	0.925	0.900
Mg	0.176	0.086	0.166	0.085	0.213	0.089	0.268	0.261	0.166	0.236	0.205	0.230	0.121	0.129	0.179	0.116
En	13.62	23.70	16.38	26.63	16.09	26.15	21.57	19.65	21.97	22.56	19.92	19.95	22.22	21.16	13.57	19.76
Fs	16.39	25.67a	16.15	26.97	16.52	25.05	26.16	25.66	27.22	26.39	29.11	29.92	26.20	27.01	16.22	27.99
Ms	50.19	50.65	51.67	50.61	51.39	50.79	52.30	56.66	50.92	53.03	52.96	51.11	51.57	51.91	52.21	52.31
Appl. En	19.55	10.83	19.32	10.79	26.12	12.29	30.93	11.07	21.29	25.56	26.66	30.07	15.53	15.99	12.79	15.15
Fe ₂ SiO ₄	1.986	1.978	1.951	1.986	1.950	1.971	1.956	1.950	1.950	1.925	1.938	1.929	1.966	1.968	1.969	1.960
Fe ₃ SiO ₄	0.135	0.087	0.165	0.080	0.239	0.119	0.281	0.306	0.207	0.296	0.299	0.300	0.150	0.155	0.205	0.121
Fe ₂ SiO ₄	0.135	0.087	0.165	0.080	0.239	0.119	0.281	0.306	0.207	0.296	0.299	0.300	0.150	0.155	0.205	0.121
Fe ₃ SiO ₄	0.135	0.087	0.165	0.080	0.239	0.119	0.281	0.306	0.207	0.296	0.299	0.300	0.150	0.155	0.205	0.121
Fe ₂ SiO ₄	0.135	0.087	0.165	0.080	0.239	0.119	0.281	0.306	0.207	0.296	0.299	0.300	0.150	0.155	0.205	0.121
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Fe ₃ SiO ₄	0.135	0.087	0.165	0.080	0.239	0.119	0.281	0.306	0.207	0.296	0.299	0.300	0.150	0.155	0.205	0.121
Fe ₂ SiO ₄	0.135	0.087	0.165	0.080	0.239	0.119	0.281	0.306	0.207	0.296	0.299	0.300	0.150	0.155	0.205	0.121
Fe ₃ SiO ₄	0.135	0.087	0.165	0.080	0.239	0.119	0.281	0.306	0.207	0.296	0.299	0.300	0.150	0.155	0.205	0.121
Fe ₂ SiO ₄	0.135	0.087	0.165	0.080	0.239	0.119	0.281	0.306	0.207	0.296	0.299	0.300	0.150	0.155	0.205	0.121
Fe ₃ SiO ₄	0.135	0.087	0.165	0.080	0.239	0.119	0.281	0.306	0.207	0.296	0.299	0.300	0.150	0.155	0.205	0.121
Fe ₂ SiO ₄	0.135	0.087	0.165	0.080	0.239	0.119	0.281	0.306	0.207	0.296	0.299	0.300	0.150	0.155	0.205	0.121
Fe ₃ SiO ₄	0.135	0.087	0.165	0.080	0.239	0.119	0.281	0.306	0.207	0.296	0.299	0.300	0.150	0.155	0.205	0.121
Fe ₂ SiO ₄	0.135	0.087	0.165	0.080	0.239	0.119	0.281	0.306	0.207	0.296	0.299	0.300	0.150	0.155	0.205	0.121
Fe ₃ SiO ₄	0.135	0.087	0.165	0.080	0.239	0.119	0.281	0.306	0.207	0.296	0.299	0.300	0.150	0.155	0.205	0.121
Fe ₂ SiO ₄	0.135	0.087	0.165	0.080	0.239	0.119	0.281	0.306	0.207	0.296	0.299	0.300	0.150	0.155	0.205	0.121
Fe ₃ SiO ₄	0.135	0.087	0.165	0.080	0.239	0.119	0.281	0.306	0.207	0.296	0.299	0.300	0.150	0.155	0.205	0.121
Fe ₂ SiO ₄	0.135	0.087	0.165	0.080	0.239	0.119	0.281	0.306	0.207	0.296	0.299	0.300	0.150	0.155	0.205	0.121
Fe ₃ SiO ₄	0.135	0.087	0.165	0.080	0.239	0.119	0.281	0.306	0.207	0.296	0.299	0.300	0.150	0.155	0.205	0.121
Fe ₂ SiO ₄	0.135	0.087	0.165	0.080	0.239	0.119	0.281	0.306	0.207	0.296	0.299	0.300	0.150	0.155	0.205	0.121
Fe ₃ SiO ₄	0.135	0.087	0.165	0.080	0.239	0.119	0.281	0.306	0.207	0.296	0.299	0.300	0.150	0.155	0.205	0.121
Fe ₂ SiO ₄	0.135	0.087	0.165	0.080	0.239	0.119	0.281	0.306	0.207	0.296	0.299	0.300	0.150	0.155	0.205	0.121
Fe ₃ SiO ₄	0.135	0.087	0.165	0.080	0.239	0.119	0.281	0.306	0.207	0.296	0.299	0.300	0.150	0.155	0.205	0.121
Fe ₂ SiO ₄	0.135	0.087	0.165	0.080	0.239	0.119	0.281	0.306	0.207	0.296	0.299	0.300	0.150	0.155	0.205	0.121
Fe ₃ SiO ₄	0.135	0.087	0.165	0.080	0.239	0.119	0.281	0.306	0.207	0.296	0.299	0.300	0.150	0.155	0.205	0.121
Fe ₂ SiO ₄	0.135	0.087	0.165	0.080	0.239	0.119	0.281	0.306	0.207	0.296	0.299	0.300	0.150	0.155	0.205	0.121
Fe ₃ SiO ₄	0.135	0.087	0.165	0.080	0.239	0.119	0.281	0.306	0.207	0.296	0.299	0.300	0.150	0.155	0.205	0.121
Fe ₂ SiO ₄	0.135	0.087	0.165	0.080	0.239	0.119	0.281	0.306	0.207	0.296	0.299	0.300	0.150	0.155	0.205	0.121
Fe ₃ SiO ₄	0.135	0.087	0.165	0.080	0.239	0.119	0.281	0.306	0.207	0.296	0.299	0.300	0.150	0.155	0.205	0.121
Fe ₂ SiO ₄	0.135	0.087	0.165	0.080	0.239	0.119	0.281	0.306	0.207	0.296	0.299	0.300	0.150	0.155	0.205	0.121
Fe ₃ SiO ₄	0.135	0.087	0.165	0.080	0.239	0.119	0.281	0.306	0.207	0.296	0.299	0.300	0.150	0.155	0.205	0.121
Fe ₂ SiO ₄	0.135	0.087	0.165	0.080	0.239	0.119	0.281	0.306	0.207	0.296	0.299	0.300	0.150	0.155	0.205	0.121
Fe ₃ SiO ₄	0.135	0.087	0.165	0.080	0.239	0.119	0.281	0.306	0.207	0.296	0.299	0.300	0.150	0.155	0.205	0.121
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Fe ₃ SiO ₄	0.135	0.087	0.165	0.080	0.239	0.119	0.281	0.306	0.207	0.296	0.299	0.300	0.150	0.155	0.205	0.121
Fe ₂ SiO ₄	0.135	0.087	0.165	0.080	0.239	0.119	0.281	0.306	0.207	0.296	0.299	0.300	0.150	0.155	0.205	0.121
Fe ₃ SiO ₄	0.135	0.087	0.165	0.080	0.239	0.119	0.281	0.306	0.207	0.296	0.299	0.300	0.150	0.155	0.205	0.121
Fe ₂ SiO ₄	0.135	0.087	0.165	0.080	0.239	0.119	0.281	0.306	0.207	0.296	0.299	0.300	0.150	0.155	0.205	0.121
Fe ₃ SiO ₄	0.135	0.087	0.165	0.080	0.239	0.119	0.281	0.306	0.207	0.296	0.299	0.300	0.150	0.155	0.205	0.121
Fe ₂ SiO ₄	0.135	0.087	0.165	0.080	0.239	0.119	0.281	0.306	0.207	0.296	0.299	0.300	0.150	0.155	0.205	0.121
Fe ₃ SiO ₄	0.135	0.087	0.165	0.080	0.239	0.119	0.281	0.306	0.207	0.296	0.299	0.300	0.150	0.155	0.205	0.121
Fe ₂ SiO ₄	0.135	0.087	0.165	0.080	0.239	0.119	0.281	0.306	0.207	0.296	0.299	0.300	0.150	0.155	0.205	0.121
Fe ₃ SiO ₄	0.135	0.087	0.165	0.080	0.239	0.119	0.281	0.306	0.207	0.296	0.299	0.300	0.150	0.155	0.205	0.121
Fe ₂ SiO ₄																

Table 1-12
(Cont.)

Sample no.	48 core	49 core	50 core	51 core	52 core	53 core	54 core	55 core	56 core	57 core	58 core	59 core	60 core	61 core	62 core
	48	49	50	51	52	53	54	55	56	57	58	59	60	61	62
SiO ₂ (wt)	49.11	49.37	48.56	50.33	49.59	48.27	51.63	48.76	50.84	52.78	53.51	54.06	54.65	54.68	54.54
SiO ₂	0.03	0.00	0.06	0.02	0.08	0.08	0.02	0.03	0.12	0.00	0.00	0.00	0.00	0.00	0.01
Al ₂ O ₃	1.03	0.68	1.22	0.56	0.82	1.18	0.68	1.51	3.00	1.92	3.63	0.62	0.91	0.91	0.95
FeO	20.29	21.98	23.49	17.80	17.43	22.94	15.89	23.67	16.46	14.75	7.09	7.11	7.51	7.21	7.58
FeO	15.42	15.06	15.39	14.14	13.30	18.24	10.99	15.95	9.01	8.78	5.43	5.12	5.92	5.90	5.46
Fe ₂ O ₃	5.41	7.69	8.44	4.08	4.58	5.22	5.44	9.60	8.39	6.63	1.84	2.21	1.74	1.44	2.24
MnO	0.43	0.52	0.50	0.45	0.41	0.57	0.39	0.95	0.22	0.39	0.34	0.30	0.53	0.29	0.33
MgO	5.70	4.51	3.61	7.08	7.48	4.43	4.50	1.78	4.98	7.97	12.74	12.73	12.33	12.92	12.29
CaO	20.79	19.00	19.52	20.59	20.81	21.08	20.75	18.98	17.26	17.79	23.89	23.36	23.33	22.40	22.69
MgO	1.55	2.67	2.70	1.42	1.42	1.39	1.81	2.44	3.99	3.71	1.11	1.25	1.01	1.14	1.48
Cr ₂ O ₃	0.13	0.24	0.29	0.09	0.19	0.05	0.13	0.04	0.04	0.02	0.03	0.04	0.04	0.05	0.06
Total	99.60	98.74	99.90	98.74	99.69	100.51	100.14	100.94	99.69	99.73	99.14	99.69	99.69	99.75	100.15

Number of ions on the basis of 6 oxygens

Si	1.940	1.948	1.934	1.944	1.952	1.918	1.971	1.920	1.940	1.999	2.002	2.008	2.022	2.021	2.015
Al	0.046	0.032	0.057	0.021	0.038	0.055	0.029	0.070	0.080	0.001	0.000	0.000	0.000	0.000	0.000
Al	0.000	0.000	0.000	0.005	0.000	0.000	0.002	0.000	0.075	0.004	0.008	0.007	0.022	0.040	0.041
Fe ²⁺	0.509	0.502	0.529	0.444	0.438	0.606	0.352	0.525	0.254	0.278	0.170	0.199	0.184	0.182	0.172
Fe ³⁺	0.161	0.231	0.253	0.120	0.136	0.156	0.157	0.255	0.270	0.189	0.052	0.062	0.049	0.041	0.062
Mg	0.336	0.268	0.214	0.415	0.439	0.262	0.485	0.222	0.391	0.450	0.712	0.705	0.682	0.712	0.677
Mn	0.014	0.018	0.017	0.015	0.014	0.019	0.013	0.028	0.007	0.004	0.011	0.009	0.017	0.009	0.010
Ca	0.001	0.000	0.002	0.001	0.002	0.002	0.005	0.001	0.003	0.000	0.000	0.000	0.000	0.000	0.000
Cr	0.009	0.008	0.009	0.003	0.006	0.002	0.005	0.003	0.001	0.001	0.001	0.001	0.001	0.001	0.002
Encl. Oct.	0.025	0.026	0.023	0.024	0.024	0.048	0.013	0.032	0.003	0.006	0.000	0.000	0.000	0.000	0.000
Ca	0.980	0.769	0.790	0.968	0.879	0.898	0.952	0.401	3.705	0.722	0.362	0.930	0.928	0.988	0.998
Mg	0.159	0.206	0.208	0.108	0.108	0.107	0.138	0.188	3.292	3.272	0.081	0.080	0.073	0.082	0.106
Encl.	19.65	17.61	13.97	23.73	25.01	16.85	29.73	16.33	29.94	11.03	19.02	19.29	16.02	19.94	14.74
Fe	29.53	32.62	30.51	26.46	26.96	34.33	20.95	31.93	18.91	19.18	9.12	9.57	10.25	10.21	9.94
Mg	51.01	49.94	51.52	49.63	50.03	50.92	50.62	51.74	52.14	49.79	51.65	51.94	51.72	49.91	51.42
Amphib	14.54	23.83	24.37	12.94	14.46	14.00	14.30	25.70	14.40	27.39	4.09	9.00	7.27	4.14	17.40
Sum	Fe-Asphre- sillite	Asphre- sillite	Asphre- sillite	Fe- sillite	Fe- sillite	Fe- sillite	Fe- sillite	Asphre- sillite	Asphre- sillite	Asphre- sillite	Asphre- sillite	Asphre- sillite	Asphre- sillite	Asphre- sillite	Asphre- sillite

48: core grain, centre of grain; 49: same, border; 50: same, along border; 51: another core grain, centre; 52: same, core centre
 53: same, border; 54: another grain, centre; 55: same, border; 56: border of grain; 57: same, centre; 58: centre of grain
 59: same, border; 60: another grain, centre; 61: same, border
 48-57: mineralized garnet-pyroxene-albite; 58-62: garnet-pyroxene-albite (Ca-rich)

*Calculated according to Papike et al. (1974)
 FeO: total iron as FeO

Table I-13. Analyses of garnets from Ca-rich albitites.

Anal. no.	1	2	3	4	5	Average	1	2	Average
SiO ₂ (wt%)	36.17	36.52	36.95	36.49	36.63	36.55	36.21	36.42	36.31
TiO ₂	1.69	1.26	0.45	0.80	1.19	1.08	1.05	1.11	1.08
Al ₂ O ₃	5.04	6.39	7.41	7.39	6.42	6.61	7.51	49.09	7.30
Fe ₂ O ₃ ^t	23.19	21.13	20.51	20.30	20.70	21.17	20.98	20.73	20.45
Fe ₂ O ₃ ^a						21.27			20.45
FeO ^a									
MnO	0.0	0.05	0.26	0.12	0.05	0.09	0.07	0.05	0.06
MgO	0.07	0.12	0.03	0.11	0.12	0.09	0.07	0.07	0.07
CaO	33.25	33.95	33.16	33.29	30.58	33.65	30.42	30.78	30.60
Na ₂ O	0.10	0.13	0.22	0.10	0.11	0.13	0.05	0.02	0.03
Total	99.51	99.55	99.99	98.60	100.20	99.37	100.36	100.27	100.30

Number of ions on the basis of 24 oxygens

Si	5.979	5.899
Ti	0.133	0.137
Al	1.274	1.394
Fe ³⁺	2.606	2.542
Fe ²⁺		
Mn	0.012	0.008
Mg	0.022	0.017
Ca	5.494	6.006
Na	0.041	0.009
Andrad %	64.41%	62.17
Pyrope %	0.40	0.30
Spess %	0.23	0.15
Gross %	36.55	37.37
Almand %		

*Calculated according to Harrold (1953)

Fe₂O₃^t: total iron as Fe₂O₃

Sample AM7 LL45-2

Sample AM7 LL45-1

Table 1-10. Analyses of garnets from albitites

Anal. no.	1	2	3	4	5	6	7	8	9	10	Average
SiO ₂ (wt%)	36.48	36.41	36.07	36.04	36.33	35.80	36.02	36.29	36.43	35.39	36.13
TiO ₂	0.45	0.48	0.55	0.49	0.41	0.52	0.41	0.42	0.34	0.56	0.47
Al ₂ O ₃	3.04	2.65	2.38	2.40	2.89	2.96	2.03	3.43	2.93	1.79	2.65
Fe ₂ O ₃ ¹	28.50	29.34	29.27	29.86	28.98	29.31	29.92	28.43	29.86	30.12	29.36
Fe ₂ O ₃ ²											26.07
FeO ²											2.96
MnO	0.79	0.64	0.67	0.99	0.66	0.77	0.75	0.85	0.98	0.86	0.79
MgO	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
CaO	31.05	31.32	31.21	30.29	31.70	31.64	32.05	30.69	30.77	31.34	31.21
Na	NA	NA	NA	NA	NA	NA	NA	NA	NA	NA	NA
Total	100.31	100.84	100.15	100.07	100.97	101.00	101.24	100.11	101.31	100.06	100.61

Number of ions on the basis of 24 oxygens

Si	6.036
Ti	0.059
Al	0.522
Fe ³⁺	3.277
Fe ²⁺	0.414
Mn	0.112
Mg	
Ca	5.586
Andrad	86.26%
Pyrope	
Spessa	1.96%
Gross	11.77%
Alm	

¹Calculated according to Barriga (1983)Fe₂O₃²: total iron as Fe₂O₃

Sample BR01 100.90 - 109.00 m

NA: not analysed

Table I-1.
(Cont.)

Anal. no.	1	2	3	4	5	6	7	Average
SiO ₂	36.00	35.52	35.90	35.90	35.44	35.54	37.93	36.09
TiO ₂	0.49	0.47	0.54	0.54	0.44	0.52	0.52	0.51
Al ₂ O ₃	2.11	2.57	2.43	2.71	2.94	2.39	2.15	2.47
Fe ₂ O ₃ ¹	29.66	29.22	30.74	30.94	29.55	29.26	29.71	29.59
Fe ₂ O ₃ ²								26.19
FeO ²								3.05
MnO	0.69	0.57	0.77	0.57	0.61	0.79	0.52	0.64
MgO	0.00	0.00	0.02	0.13	0.16	0.03	0.18	0.07
CaO	31.00	31.21	31.37	31.37	30.40	30.77	31.42	31.07
Na ₂ O	0.00	0.03	0.16	0.07	0.35	0.22	0.59	0.20
Total	99.95	100.59	99.68	100.06	99.32	99.29	100.50	100.64

Number of ions on the basis of 24 oxygens

Si	6.033
Ti	0.064
Al	0.487
Fe ³⁺	3.295
Fe ²⁺	0.427
Mn	0.091
Mg	0.017
Ca	5.565
Na	0.065
Andrad	97.108
Pyrope	0.318
Speress	1.608
Grossul	10.908
Almand	

¹Calculated according to Barriga (1983).
Fe₂O₃²: total iron as Fe₂O₃

Sample BR01 198.00-198.13 m

Table 1-14
(Cont.)

Anal. no.	1	2	3	Average	1	2	3	4	5	6	Average	1	2	3	4	Average
SiO ₂	35.11	35.00	36.09	35.40	35.32	35.99	35.56	35.70	36.49	36.03	35.95	36.92	36.21	36.79	36.32	36.63
TiO ₂	0.61	0.90	0.56	0.56	0.26	0.51	0.63	0.66	0.66	0.64	0.63	0.56	0.60	0.31	0.29	0.36
Al ₂ O ₃	3.10	2.37	2.68	2.72	2.39	2.43	1.99	2.29	2.73	2.90	2.62	2.62	3.05	3.61	3.06	3.03
FeO ₂ ¹	29.86	29.96	29.76	29.92	29.96	29.23	29.99	29.23	29.06	29.86	29.20	29.36	29.97	29.63	29.96	29.86
FeO ₂ ²				26.19							26.33					25.77
FeO ³				2.45							2.56					2.51
MgO	0.67	0.40	0.64	0.70	0.77	0.63	0.72	0.79	0.99	0.95	0.81	1.05	1.02	1.05	1.06	1.06
Mg	00	00	00		0.11	0.06	0.10	0.11	0.11	0.08	0.09	0.13	0.16	0.11	0.13	0.13
CaO	31.99	31.80	31.20	31.53	31.27	31.26	30.90	30.72	30.93	30.75	30.97	30.73	31.21	31.46	31.18	31.15
MgO	00	00	00		NA	NA	NA	NA	NA	NA	NA	00	00	00	00	
Total	99.96	99.63	100.13	99.93	99.08	100.19	99.51	99.24	100.73	99.93	99.77	101.07	100.90	101.19	100.96	101.02

Number of ions on the basis of 24 oxygens

Si	5.966	6.037	6.063
Ti	0.071	0.056	0.062
Al	0.560	0.680	0.993
Fe ³⁺	1.321	1.337	1.279
Fe ²⁺	0.766	0.363	0.390
Mg	0.089	0.115	0.166
Ca	5.692	5.568	5.536
Na			
Total %	99.96	99.63	99.63
Pyrope %		0.19	0.56
Diopside %	1.73	2.32	2.56
Gross %	12.32	10.16	12.63
Alm %			

¹Calculated according to Barriac (1993)
FeO₂¹: total iron as FeO₂

NA: not analyzed

Sample BR01 109.63-109.70 μ

Sample BR01 105.62-105.70 μ

Sample BR01 106.70-106.79 μ

Table 1-14
(Cont.)

Anal. no.	1	2	3	4	5	6	7	8	Average	1	2	3	4	5	Average
SiO ₂	75.46	75.79	75.12	75.46	75.46	75.46	75.46	75.46	75.46	75.46	75.46	75.46	75.46	75.46	75.46
Al ₂ O ₃	0.57	0.51	0.50	0.50	0.52	0.53	0.55	0.75	0.57	0.73	0.35	0.48	0.79	0.63	0.54
Fe ₂ O ₃	2.16	2.49	2.21	2.21	2.09	2.48	2.71	2.54	2.41	3.55	2.73	2.70	3.09	2.90	2.99
MgO	29.71	29.88	29.05	28.47	29.26	29.51	29.46	27.54	28.91	28.06	29.34	28.86	28.57	28.65	28.66
CaO								26.82							25.82
Na ₂ O								2.24							2.55
K ₂ O	0.79	0.63	0.70	0.69	0.66	0.59	0.53	0.70	0.66	0.88	0.93	0.82	0.85	0.90	0.88
MnO	0.09	0.11	0.12	0.14	0.13	0.16	0.17	0.16	0.13	0.00	0.00	0.00	0.00	0.00	0.00
CO ₂	30.57	31.02	31.50	31.06	30.79	31.16	31.06	31.76	31.11	31.86	30.66	31.28	31.73	31.69	31.44
H ₂ O	0.18	0.32	0.39	0.37	0.61	0.54	0.64	0.63	0.65	0.00	0.00	0.00	0.00	0.00	0.00
Total	99.64	99.71	99.59	99.60	99.70	100.15	99.74	97.74	99.76	100.14	99.60	99.55	100.09	99.86	99.85

Number of ions on the basis of 24 oxygens

Si	5.987	5.952
Al	0.072	0.086
Al	0.479	0.593
Fe ³⁺	3.152	3.273
Fe ²⁺	0.315	0.399
Mg	0.094	0.125
Mg	0.033	
Ca	5.619	5.676
Na	0.167	
Total	87.99	88.65
Pyroxene	0.57	
Spinel	1.65	2.17
Glass	10.39	11.19
Sum		

*Calculated according to Burridge (1963).

Sample 9901 108.75-108.85 °

Sample 9901 107.15-107.25 °

Table 1-15. Phosphates which appear as secondary product after exposure to ultrasonic waves.

Sample no.	1 0801 105.65- 105.73m	2 0801 105.70- 105.78m	3 idem	4 idem	5 0801 105.27- 105.33m	6 idem	7 idem	8 idem	9 idem	10 0801 104.05- 104.10m	11 idem	12 idem	13 0801 106.90- 106.00m	14 idem
SiO ₂ (wt%)	61.61	60.16	59.61	59.26	61.63	64.33	63.79	63.11	64.38	61.65	61.21	62.56	65.27	62.21
Al ₂ O ₃	11.08	11.28	11.10	11.20	8.25	7.81	7.75	8.17	7.72	9.48	9.40	9.18	6.64	7.26
Fe ₂ O ₃	24.86	26.28	26.37	26.16	22.25	19.42	20.49	21.24	20.05	24.89	24.67	22.10	21.73	23.39
FeO	48.88	19.42	19.24	19.21	18.36	16.71	17.80	17.99	17.92	20.89	18.96	18.37	19.09	19.89
Fe ₂ O ₃ + FeO	6.48	7.62	7.92	7.73	6.32	3.65	3.56	3.61	2.36	6.66	6.57	6.16	2.96	3.89
MgO	4.28	5.93	5.70	5.54	9.46	10.33	9.56	9.18	9.81	6.55	7.10	6.55	9.89	8.23
MnO	0.87	0.57	0.29	0.31	0.61	0.29	3.66	0.36	0.36	0.33	0.28	1.40	0.29	0.27
TiO ₂	0.09	0.22	0.14	0.00	1.30	0.36	0.39	0.88	0.63	0.18	0.27	0.27	0.13	0.06
CaO	8.39	10.97	10.99	10.95	10.05	10.48	10.88	10.20	10.52	10.83	11.03	11.05	11.08	12.97
Mg ₂ SiO ₄	1.36	1.55	1.48	1.53	2.86	2.05	2.26	2.36	2.39	1.85	1.75	1.58	1.50	2.08
SiO ₂	1.53	1.90	1.71	1.83	1.70	1.67	1.68	1.63	1.70	1.61	1.61	1.35	1.03	1.19
Total	97.03	99.48	97.98	97.35	99.14	97.67	97.58	97.03	97.75	97.86	97.98	96.85	96.85	98.84

Number of ions on the basis of 23 oxygens

Si	6.449	6.191	6.177	6.192	6.856	6.758	6.718	6.669	6.805	6.487	6.406	6.626	6.977	6.381
Al	1.550	1.809	1.823	1.808	1.384	1.242	1.281	1.331	1.195	1.513	1.594	1.378	1.023	1.336
Al	0.484	0.239	0.228	0.275	0.086	0.162	0.120	0.159	0.128	0.236	0.139	0.127	0.163	0.080
Fe ₂ O ₃	2.659	2.503	2.523	2.534	2.336	2.131	2.259	2.328	2.288	2.708	2.465	2.392	2.448	2.595
Fe ₂ O ₃	0.779	0.686	0.934	0.917	0.686	0.386	0.365	0.420	0.389	0.569	0.588	0.645	0.361	0.436
Mg	1.658	1.362	1.331	1.308	2.166	2.367	2.254	2.114	2.232	1.528	1.695	1.584	2.042	1.912
Mn	0.115	0.076	0.038	0.061	0.053	0.037	0.059	0.045	0.046	0.046	0.037	0.053	0.038	0.036
Ti	0.000	0.025	0.016	0.000	0.034	0.039	0.065	0.051	0.049	0.021	0.031	0.032	0.015	0.005
Ca ₂ SiO ₄	0.295	0.088	0.092	0.089	0.146	0.112	0.104	0.119	0.096	0.086	0.096	0.073	0.078	0.082
Ca	1.636	1.911	1.884	1.851	1.638	1.765	1.756	0.169	1.721	1.916	1.837	1.843	1.825	2.167
Mg ₂ SiO ₄	0.271	0.101	0.082	0.099	0.214	0.163	0.161	0.190	0.203	0.096	0.087	0.088	0.091	0.080
Mg ₂ SiO ₄	0.254	0.162	0.167	0.127	0.629	0.463	0.526	0.519	0.508	0.463	0.440	0.514	0.757	0.631
K	0.304	0.373	0.362	0.368	0.329	0.286	0.321	0.322	0.311	0.321	0.279	0.288	0.302	0.237
Fe ₂ O ₃ + FeO	2.30	2.54	2.62	2.68	1.34	1.09	1.19	1.32	1.16	2.16	1.99	1.68	1.39	1.61
Fe ₂ O ₃ + FeO	0.69	0.72	0.72	0.72	0.57	0.52	0.56	0.57	0.56	0.68	0.65	0.59	0.58	0.62

1-5, 7-14: Haplophosphate

6: Haplophosphate

1: albitite

2: garnet-albitite

5, 10, 13: garnet-sper-albitite

* Calculated according to Papike et al. (1974).

Fe₂: total iron.

Table 1-13
(Cont.)

	15	16	17	18	19	20	21	22	23	24	25	26	27	28	29
Sample	400 944874	400 944874	400 944874	400 944874	400 944874	400 944874	400 944874	400 944874	400 944874	400 944874	400 944874	400 944874	400 944874	400 944874	400 944874
SiO ₂ (wt)	61.71	61.53	60.95	60.88	59.38	60.36	58.71	65.01	58.63	51.26	63.68	61.30	66.85	65.45	65.06
Al ₂ O ₃	10.23	10.08	10.13	10.06	11.12	10.26	11.71	6.77	7.97	6.02	10.05	9.56	9.76	9.13	9.16
FeO	20.30	20.29	22.39	24.15	25.89	25.02	27.38	21.86	21.36	25.83	16.93	16.87	16.88	16.38	15.88
Fe ₂ O ₃	16.39	16.22	15.95	17.62	16.76	19.81	20.53	18.38	18.37	19.97	11.73	10.87	11.10	11.27	10.16
Fe ₂ O ₃ /FeO	6.12	6.76	7.07	7.26	6.57	6.78	7.03	1.83	1.30	3.63	3.33	4.46	6.11	1.67	1.57
MgO	3.39	3.49	3.32	3.13	6.13	6.33	6.77	9.11	10.23	7.25	13.22	13.30	12.96	13.90	13.63
CaO	0.21	0.20	0.27	0.26	0.26	0.79	0.37	0.55	0.63	0.36	0.36	0.33	0.26	0.27	0.16
Na ₂ O	0.16	0.23	0.22	0.11	0.25	0.06	0.20	0.16	0.11	0.08	0.13	0.10	0.07	0.06	0.10
CaO	11.36	10.82	11.39	11.38	11.26	11.09	11.17	10.88	11.16	11.10	11.32	11.36	10.11	10.75	10.09
MgO	2.56	1.69	1.73	1.50	1.26	1.69	1.30	1.38	0.96	2.01	2.86	2.97	2.96	2.61	2.56
FeO	1.61	1.69	1.79	2.06	1.88	1.83	2.18	1.28	0.55	1.30	1.30	1.48	0.88	0.98	0.75
Total	97.72	96.30	99.23	99.6	98.66	97.68	98.77	97.56	97.63	97.17	98.18	97.97	96.62	98.61	97.56

Number of ions on the basis of 23 oxygens

Si	6.893	6.810	6.267	6.236	6.762	6.399	6.028	6.982	7.364	6.966	6.894	6.608	6.888	6.627	6.666
Al	1.027	1.009	1.733	1.736	1.838	1.684	1.922	1.089	12.634	1.456	1.569	1.538	1.369	1.333	1.796
Al	0.763	0.239	0.160	0.196	0.282	0.276	0.266	0.126	0.093	0.029	0.251	0.193	0.298	0.266	0.236
Fe ₂₊	2.127	1.836	1.032	2.259	2.361	2.562	2.686	2.399	2.199	2.623	1.688	1.352	1.385	1.383	1.293
Fe ₃₊	0.675	0.763	0.890	0.839	1.006	0.736	0.983	0.962	0.375	0.680	0.395	0.607	0.643	0.686	0.596
Mg	3.198	2.163	1.891	1.629	1.622	1.676	1.116	2.082	2.388	1.697	2.988	2.992	2.981	2.979	2.999
Ca	0.027	0.030	0.125	0.127	0.063	0.096	0.089	0.071	0.093	0.066	0.063	0.062	0.090	0.086	0.082
Na	0.078	0.077	0.023	0.013	0.026	0.006	0.026	0.076	0.112	0.088	0.096	0.091	0.075	0.073	0.083
Sum Oct.	0.078	0.096	0.086	0.061	0.096	0.086	0.086	0.089	0.109	0.083	0.082	0.089	0.110	0.116	0.109
Ca	1.829	1.789	1.862	1.869	1.875	1.853	1.879	1.786	1.935	1.982	1.882	1.887	1.616	1.689	1.586
Mg ₂₊	0.097	0.113	0.076	0.089	0.067	0.078	0.089	0.113	0.085	0.129	0.166	0.163	0.233	0.195	0.237
Mg ₃₊	0.689	0.333	0.638	0.376	0.313	0.635	0.331	0.356	0.196	0.557	0.655	0.638	0.607	0.567	0.609
Ca	0.276	0.295	0.366	0.369	0.373	0.366	0.377	0.238	0.096	0.289	0.286	0.281	0.167	0.163	0.161
Fe ₂ /Fe ₃	1.20	1.21	1.39	1.39	2.61	2.38	2.29	1.38	1.19	1.86	0.63	0.63	0.63	0.61	0.62
Fe ₂ /Fe ₂₊₃	0.38	0.39	0.61	0.66	0.71	0.69	0.77	0.58	0.56	0.66	0.39	0.39	0.39	0.38	0.38

17-26: non-bridging

15, 16, 25-26: bridging

15: garnet-pyroxene doublets

17: pyroxene doublets

21, 22, 25: clinopyroxene-pyroxene doublets

27: clinopyroxene doublets (no pyroxene)

*Calculated according to Peacor et al., 1976a.
Fe₂: total iron.

4

OF/DE

4

MICROCOPY RESOLUTION TEST CHART
NATIONAL BUREAU OF STANDARDS
STANDARD REFERENCE MATERIAL 1010a
(ANSI and ISO TEST CHART No. 2)

Table 1-15 Analyses of epidotes from epidiorites

Sample no.	10	11	12	13	14	15	16	17	18	19	20	21	22	23
	100	100	100	100	100	100	100	100	100	100	100	100	100	100
$\text{SiO}_2(\text{wt})$	58.00	57.55	57.63	57.61	57.78	57.32	57.50	57.66	57.5	57.98	57.48	57.61	57.95	57.95
TiO_2	0.06	0.08	0.06	0.06	0.07	0.06	0.09	0.07	0.00	0.08	0.00	0.00	0.00	0.00
Al_2O_3	26.07	26.55	25.87	25.78	25.92	25.65	25.7	25.21	25.86	25.20	22.1	25.72	22.58	22.58
Fe_2O_3^*	13.03	13.50	13.48	13.47	13.11	13.48	13.61	13.68	13.69	14.31	14.27	13.57	14.21	14.21
FeO	0.02	0.05	0.05	0.05	0.06	0.00	0.05	0.00	0.06	0.06	0.07	0.08	0.08	0.08
MgO	0.00	0.01	0.00	0.00	0.00	0.00	0.00	0.01	0.00	0.02	0.00	0.00	0.00	0.00
CaO	23.90	23.52	22.91	23.28	23.00	23.15	23.68	23.60	23.11	22.11	22.95	22.76	22.26	22.26
Na_2O	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
K_2O	0.09	0.26	0.06	0.06	0.16	0.16	0.17	0.16	0.20	0.03	0.08	0.06	0.08	0.08
Li_2O	0.68	0.11	0.14	0.00	0.19	0.23	0.22	0.13	0.21	0.61	0.08	0.06	0.02	0.02
CaF_2	0.95	0.32	0.27	0.05	0.08	0.05	0.07	0.05	0.08	0.09	0.11	0.11	0.10	0.10
Ce_2O_3	0.03	0.05	0.04	0.06	0.17	0.07	0.08	0.03	0.06	0.07	0.10	0.11	0.11	0.11
Total	99.83	99.86	99.35	99.62	99.62	99.97	99.36	99.58	99.52	99.56	99.65	99.38	99.11	99.11

Number of ions on the basis of 25 oxygens

Si	5.943	5.912
Ti	0.007	0.000
Al	6.410	6.412
Fe^{3+}	7.16	7.161
Fe	0.007	0.008
Mg	0.000	0.000
Ca	1.926	1.903
Na	0.000	0.000

10-19: epidote in type 1 epidiorites, replacing pyroxene

21-23: epidote (distalite) in type 2 epidiorites, replacing pyroxene

*total iron as Fe^{3+}

* oxidation state not determined: values reported as total oxide

APPENDIX II

SUMMARY OF ANALYTICAL TECHNIQUES

II - 1 Electron microprobe mineral analyses

Mineral chemical analyses were obtained on 25 mm diameter polished thin sections coated with carbon, using a Material Analysis Company model 400 electron microprobe equipped with three diffractometers and KRISSEL automation. Working conditions were 15 to 20 kV excitation voltage and 0.25 to 0.5 nA beam current (depending on the mineral being analysed), at counting conditions of 30 seconds or 20000 counts. The standards used were well analysed natural minerals and synthetic glasses, selected to minimize necessary corrections (MAGIC corrections were used throughout). Replicate determinations indicate a precision generally better than 2% of the amount present, for major elements.

II - 2 Whole rock geochemistry

Whole rock samples were diamond cut into several slabs ~0.5 cm thick, subsequently grinded with Al_2O_3 abrasives to eliminate brass contamination, and thoroughly washed and dried, and lightly hammered between paper or cloths to avoid any contact with metal, and the resulting fragments subsequently split to about 50 g. This amount was in turn powdered for ~60 seconds in a Bleuler mill.

Major elements (with the exception of Na) were determined by X-ray fluorescence spectrometry, in a Phillips PW 1450 spectrometer fitted with a Cr tube, using the heavy absorber fusion technique of Norrish and Hutton (1969), where a glass disk is made, composed of 2.0000 g of flux (lithium tetraborate, lithium carbonate and lanthanum oxide), 0.0267 g of sodium nitrate and 0.3733 g of sample. Calibration monitor samples FS-94 and FS-84 (supplied by Dr. K. Norrish, CSIRO, Australia) were used for calculation of the concentrations in the samples. Na was determined in pressed pellets of rock powder, with reference calibration lines defined by international standards.

Most trace elements were determined on pressed pellets, by X-RF spectrometry (W tube) with reference to selected international standards. Data were reduced using interference and matrix calculations through program TRACE developed by G. Barker, H. Hunter and T. LaTour (unpublished, U.W.O.).

International standards and duplicate unknown samples were run for accuracy check, and results indicate that major element determinations are generally accurate within 1-2% of the amount present,

and trace elements accurate within 10% (20% for Nb and Cu).

Sc, Hf, Ta, U and Th were determined by instrumental neutron activation analysis at Neutron Activation Services, Hamilton, Ontario. Results from samples in Chapter 5 which only display Sr, Rb, Y and Zr as trace elements, were analysed by X-RF, done by X-ray Assay Labs., Don Mills, Ontario.

Ferrous iron was determined after the method of Wilson (1955). International standards were analysed as an unknown with each batch, and results indicate that accuracy and precision are about 5% of the amount present. Volatiles were determined by weight loss on rock powder following heating at 1100°C.

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