The Turiy Massif, Kola Peninsula, Russia:

Open-system disequilibrium

by

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ABSTRACT

The Turiy Massif, that lies within the Kandalaksha Graben in the Kola Peninsula, Russia, consists of five complexes dominated by carbonatites, melilitolites and ijolites. Overall, the massif is geochemically heterogeneous in terms of age, parental magma sources, and crystallization conditions.

Rb/Sr isochron ages define two events, one at 588 ± 72 Ma (pyroxenite) and the other at 363 ± 3.5 Ma (phoscorite). The latter age is similar to an age of 373 ± 5.1 Ma (ijolite) obtained by Kramm et al. (1993) for Turiy, and to the age suggested for widespread alkaline magmatism throughout the Kola Peninsula.

$\epsilon_{\text{Sr}}(\text{T}) - \epsilon_{\text{Nd}}(\text{T})$ data from different rock-types form clusters which rarely overlap, and, of all the samples analyzed, the phoscorites show the most depleted isotopic signatures. Most of the data from the silicate rocks lie along the Kola Carbonatite Line (KCL) (Kramm, 1993) and involve two-component mixing between an enriched mantle end-member located within the spinel stability field with trace element and isotopic characteristics similar to modern EM I basalts, and a depleted, asthenospheric end-member located within the garnet-dolomite stability field, with trace element, but not isotopic characteristics of modern HIMU basalts.

The central complex carbonatites show unusually low degrees of trace element enrichment for such rocks, and their isotopic ratios lie below the KCL. A third mantle component similar to a remelted, ancient, lithospheric carbonatite may have been involved in their genesis. The $\delta^{13}\text{C}_{\text{PDB}} - \delta^{18}\text{O}_{\text{SMOW}}$ signatures of these carbonatites suggest derivation from a magma source with heavy $\delta^{13}\text{C}_{\text{PDB}}$ values common to other carbonatites emplaced along the Kandalaksha Graben. Crustal contamination, indicated by high $\epsilon_{\text{Sr}}(\text{T})$ and $\delta^{18}\text{O}_{\text{SMOW}}$ isotope data, has affected both the silicate and carbonatite magmas of the satellite complexes during their emplacement.

Several minerals, including apatite, garnet, magnetite, melilite, mica and pyroxene, show systematic variations in composition according to the rock type in which they occur. As one example, octahedrally co-ordinated Al only occurs in garnet and mica from the melilitolites.

The strong textural heterogeneity of many of the samples, coupled with the variations in isotopic and mineralogical compositions, suggest that the Turiy magmas formed, evolved, and crystallized under open-system conditions.

...the quote "there is no problem so big that one cannot
solve it, and even more so for a bigger, longer, and
harder. This quarterly incomplete list of acknowledgments is
...who have challenged me to delve deeply into the mysteries
...have sang, laughed and cried with me during the last four
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"Many know much about this stone,

everyone knows something,
but no-one knows quite enough"

Author unknown, France (1492)

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CHAPTER 1

INTRODUCTION

1.1 Carbonatites - an overview

Carbonatites are igneous rocks, containing > 50% carbonate minerals (Le Maitre, 1989), such as calcite, dolomite, ankerite and siderite. They occur as both extrusive and intrusive rocks, and, although relatively rare on a world-wide scale, they have been found in every continent (Woolley & Kempe, 1989; Hall et al., 1995). Tectonically, alkaline massifs which contain carbonatites are typically emplaced along zones of weakness at craton boundaries (Kapustin, 1980), and commonly within graben structures (Keller et al., 1990). They are rarely found in “oceanic” environments, except in areas of unusually thick lithosphere, such as the Canaries and Cape Verde Islands (Woolley & Kempe, 1989). The isotopic signatures of modern (continental) carbonatites are identical to those found in most oceanic island basalts (Basu & Tatsumoto, 1980; Bell & Blenkinsop, 1987b; Nelson et al., 1988), and thus it is likely that these two rocks types share a common source. The simplest explanation is that this source lies within the asthenosphere which underlies both oceanic and continental lithosphere. However, the isotopic signatures of Canadian carbonatites indicate that the same source has been involved in carbonatite production since the Archean (Bell et al., 1982; Bell & Blenkinsop, 1987a). This creates an apparent paradox in that the constancy and isolation of this source over such an extended time period is difficult to reconcile with its location within an actively convecting asthenospheric mantle.

The only carbonatite volcano currently active at the present time is Oldonyo L’Engai* in Tanzania (Bell & Keller, 1995), but carbonatites have been erupting and

---

* The spelling of Oldonyo L’Engai used throughout this thesis is taken from Donovan (1978). It is based on the original Masai name for the volcano which translates as “The Mountain of God” where “Engai” means both “God” and “rain” in the Masai language.

intruding consistently throughout geological time since the Archean (Woolley, 1989). They normally occur as either a) hypabyssal stock-like intrusions, b) volcanic veins, or c) effusive material (Barker, 1989). Measurements from Oldonyo L'Engai indicate that carbonatites have the lowest eruptive temperatures and viscosities of any known magmas (Pinkerton et al., 1995). Most carbonatites occur in association with a variety of silicate rocks, including kimberlites, olivine melilitites, nephelinites and phonolites in extrusive and hypabassal environments. In intrusive and hypabyssal massifs, they occur in association with syenites, pyroxenites, ijolites, melilitolites, phoscorites and related rocks (definitions of rock names used in this thesis can be found in Appendix Bi). Due to their high concentration of trace elements, they are commonly mined for rare-mineral deposits.

1.2 Why study carbonatites?

Mantle-derived carbonatites are both a rare and persistent component of the igneous geological record of our planet. Their extremely high trace element contents means that their geochemical characteristics are rarely affected by crustal contamination (Bell & Blenkinsop, 1987a,b). Hence, carbonatites provide a means for us to track the geochemical characteristics of the sub-continental upper mantle (SCUM) from the Archean to the present day. Bell et al. (1982) used Canadian carbonatite isotope data to characterize the SCUM beneath the Canadian shield area since the Archean, and found that the carbonatite source reservoir had evolved as a closed-system throughout the entire time period.

That primary carbonate and CO₂ (found in mineral inclusions) exist within the mantle, and have done since the Earth's formation, is not disputed. Dolomite and magnesite are both stable under certain conditions within the upper mantle (Falloon & Green, 1990). All mantle-derived magmas contain a variety of type and quantity of volatile material, including such volatiles as F, Cl and H₂O, all of which may be found in

carbonatite magmas. The fundamental question to be addressed is: how does carbonate material become concentrated in the mantle, become isolated from any associated silicate material, and rise to the surface of the earth as a carbonatite magma?

Thus, the nature of the relationship or relationships between carbonatites and their spatially and temporally associated silicate rocks is still a matter for great debate amongst the scientific community. Possible relationships include silicate-carbonate liquid immiscibility (Kjarsgaard & Hamilton, 1989), or the fractionation of a carbonate-rich, SiO_2 -poor silicate magma (Gittins, 1989). The issue of the possible existence of primary carbonatite magmas is one of the most contentious issues currently being debated. Criteria with which to recognise a primary carbonatite magma have not been identified, and there are difficulties in envisaging a mechanism by which a carbonatite magma can rise to the surface without dissolving in associated magmas or reacting with wall-rock. Bell (1997) referred to this process as “chemical death”.

The main reasons, therefore, for studying carbonatites, are: (i) to aid our understanding of the spatial and temporal nature of mantle processes within the Earth, and (ii) to understand the relationship(s) between silicate and carbonate melts. To do this, it is necessary to understand the nature of the carbonatite magma source, the way in which the magma is derived from that source, the means by which that magma travels from its source to its final destination, and its crystallization history on arrival.

1.3 Why Turiy?

The Turiy Massif is located on the southern shore of the Kola Peninsula in north-west Russia, just inside the Kandalaksha Graben. Palaeozoic in age, Turiy is one of a large suite of ca. 370 Ma alkaline intrusions located within the Kola Peninsula (Figure 1.1) (Kramm et al., 1993). This suite of intrusions includes Khibina, situated north of Turiy, which is the largest known syenite intrusion on Earth, and which contains the

Figure 1.1 Map of the Kola Peninsula showing the main Palaeozoic alkaline intrusions and major structural features.

largest apatite deposit in the world. Another intrusion, Kovdor, situated north-west of Turiy close to the Russia-Finland border, contains a suite of rocks similar to those found in Turiy. Other members of the suite include Afrikanda, Kandaguba, Lovozero (just east of Khibina), and Sokli. Many of the intrusions, such as Kovdor, Khibina, Lovozero and Afrikanda, have been mined for apatite, magnetite, perovskite, phlogopite, and rare-element minerals. The Khibina massif contains a very diverse mineralogy, and is host to 76 minerals which are unique to the intrusion. Turiy was drilled to > 300 m depth in the early 1970s (Ronensen et al., 1978) to determine whether mining was feasible in this massif, but the apatite and magnetite ores (Turiy's main ore potential) were of too low a grade to warrant development of the prospect. However, the rocks themselves are of enormous geological interest.

Turiy is one of the few alkaline intrusions in the world whose silicate mineralogy is dominated by melilite-bearing rocks. Other examples include Oka (Quebec) (Gold et al., 1986), Gardiner (Greenland) (Nielsen, 1980), Iron Hill (Colorado) (Armbrustmacher & Futa, 1985), Rangwe (Tanzania) (Le Bas, 1977), Kovdor (Kola Peninsula) (Arzamastsev, 1994), and several intrusions (e.g. Kugda) from the Maimecha-Kotui province (Siberia) (Egorov, 1969). The majority of the melilitolites at Turiy are well preserved, and show a variety of modal mineralogy, textures, and chemical compositions. Pristine carbonatites, some of which contain melilite and may be derived from or related to the melilite-bearing silicate rocks, are also abundant in the centre of the Turiy intrusion which will be described in more detail in later chapters. Turiy also contains phoscorites, ijolites, pyroxenites, and a variety of hypabyssal alkaline rocks which form a dyke-suite along the coast, around the massif.

The Turiy massif provides an excellent setting to evaluate the mantle sources of both silicate and carbonate magmas. The aims of the study include the use of trace element and isotopic studies to help: (i) establish the age of the rocks, (ii) evaluate

whether there has been more than one episode of magmatic activity, and (iii) establish the geochemical characteristics of the mantle sources that have been involved in the genesis of these magmas. Kramm et al. (1993) suggested that crustal contamination had also been involved in Turiy's melilitolite magmas, and this assertion needs to be quantified. Geochemical investigations may also help to understand the relationships between the melts - are there similar isotopic signatures in carbonatite-silicate magma pairs, indicating similar sources, crystal fractionation or liquid immiscibility? How do the melilitolites relate to the carbonatites? Are the pyroxenites cumulate, and if so, from which magmas have they separated?

Studies such as that done by Nash (1972) on the Iron Hill Complex in Colorado, showed that there were systematic differences in the composition of minerals which exhibited solid-solution variations between different rock types. Such systematic variations were also noted by Kapustin (1980) in some of the Kola carbonatite intrusions. This provides a means of examining magma chemistry and changes in crystallizing conditions between the different magma types. A number of minerals, including mica and garnet, are common to a majority of the Turiy samples, and may show systematic variations in chemistry. Another point of interest is the great difference in modal mineralogy between extrusive melilite-bearing rocks, e.g. the olivine melilitites from the Rhinegraben of Germany (Dunworth & Wilson, submitted) and intrusive melilite-bearing rocks such as those found in Turiy.

Additionally, the question of equilibrium in these SiO_2 -poor magmas must be addressed. Isotopic and petrogenetic disequilibrium have been shown in nephelinites (Simonetti & Bell, 1993) and melilitites (Dunworth & Wilson, submitted) respectively. The issue of closed vs open-system conditions is particularly relevant in a slowly-cooled, large intrusion where processes such as gravitational settling, diffusion, wallrock-assimilation and the influx of fresh batches of magma may have been operative.

1.4 Overview of world-wide carbonatite occurrences

The most recent summary of the world-wide distribution of carbonatites was made by Woolley (1989). There are approximately 330 carbonatites known world-wide (Figure 1.2), ranging from Archean to Recent in age, with abundance inversely proportional to age. This may be a true geological feature of the Earth, or may relate to proportionally greater preservation of younger crustal material. More than one episode of carbonatitic magmatism can occur within a given area, and the time gaps between the episodes may be several hundreds of millions of years. Approximately half of the world's carbonatites are located in Africa (Woolley & Kempe, 1989).

Carbonatites occur in two main associations with silicate rocks. Firstly, they can be associated with kimberlites and/or olivine melilitites in stable cratonic areas, seemingly unrelated to lithospheric stretching (e.g. Saltpetre Kop, South Africa). Haggerty (1994) proposed that, on a world-wide scale, major episodes of kimberlite magmatism were followed 20 Ma later by major episodes of carbonatite magmatism, and related the onset of each of these cycles to major overturn events within the outer core, causing the onset of the initial kimberlite magmatism from the D'' layer at the core-mantle boundary.

The second silicate-rock - carbonatite association occurs in areas of lithospheric stretching and mantle upwelling, such as the Kaiserstuhl carbonatite in the Rhinegraben (Keller et al., 1990) and the carbonatites of the East African Rift (Woolley & Kempe, 1989). Silicate magmas associated with carbonatites in areas of lithospheric stretching include melilitites, nephelinites and phonolites and their intrusive equivalents. Pyroxenites are commonly found as mantle xenoliths in associated extrusive silicate rocks, or as cumulate(?) bodies associated with carbonatite intrusions (e.g. Jacupiranga, Huang et al., 1995). Lithospheric stretching associated with carbonatite production is

Figure 1.2 World-wide map of carbonatite occurrences. After Woolley (1989). Carbonatites also found in Antarctica (not shown) (Hall et al., 1995; Yaxley, pers comm, 1997)

normally related to uplift due to mantle plume activity. Despite lithospheric flexure due to the Alpine collision belt, Wilson et al. (1995) showed that a mantle plume component was needed to produce the isotopic signatures of the Rhinegraben melilitites, nephelinites and the Kaiserstuhl carbonatite in Tertiary-Quaternary times. The existence of a mantle plume beneath the East African Rift zone has also been postulated, on the basis of both geophysical and geochemical evidence (Bell & Simonetti, 1996). However, the existence of a mantle plume beneath the Kola Peninsula generated during the Palaeozoic is harder to prove geophysically, partly due to the fact that volcanism ceased more than 300 Ma ago, and plume uplift is no longer occurring. The sheer quantity of magma produced, and the isotopic (and trace element) signatures of the magmas produced during the Palaeozoic suggest that a plume may have been involved in the genesis of much of the alkaline and carbonatitic material.

It should be remembered that although a plume may impinge on the base of the lithosphere and invoke melting and lithospheric stretching, the final location of emplacement of the resultant magma bodies will be dependent on zones of pre-existing weakness within the lithosphere which may not be directly above the centre of the plume head. This is seen in the Kola Peninsula (Figure 1.2), in the emplacement of the majority of the Palaeozoic alkaline massifs along zones of pre-existing weakness within the Archean basement rocks and underlying lithosphere.

CHAPTER 2

GEOLOGICAL SETTING

2.1 Kola Peninsula

The Kola Peninsula, located between $\sim 66\text{--}70^\circ\text{N}$ and $\sim 30\text{--}42^\circ\text{E}$ in north-western Russia, bounded by the White Sea and Kandalaksha Graben to the south, and the Barents Sea to the north, contains the largest concentration of alkaline magmatism on Earth (Figure 2.1) (Kogarko et al., 1995). Geological investigations, focusing on tectonics, magmatism and economic ore deposits, have been carried out for more than 150 years by Russian and Western geologists.

2.1.a Precambrian

A geological map of the Kola Peninsula is shown in Figure 2.1. To simplify discussion, the units shown in Figure 2.1 have been divided into early and late Archean and early and late Proterozoic formations, based on age relationships. Magmatic units within the Archean, Proterozoic and Palaeozoic have been shown separately. A suite of rocks whose age straddles the Archean-Proterozoic boundary have been drawn as a separate category. A cross section (A-B on Figure 2.1), based on a more detailed division of the geological formations, is shown in Figure 2.2.

Archean gneisses and granite-gneisses underly most of the peninsula and outcrop along the northern and eastern parts of the peninsula. The geological map (Figure 2.1) and NNE-SSW cross-section (Figure 2.2) record deep faulting that trends NW-SE, dividing the peninsula into four or five parallel lineaments, which are subsequently cut by a small number of major N-S faults in the western part of the peninsula. Minor, early faults trend NE-SW or E-W and were subsequently cut by the NE-SW which was then followed by N-S deformation. Major folding and ductile deformation appear to be lacking throughout the peninsula; instead the faults define the boundaries of uplifted and

Figure 2.1 Simplified map of the Kola Peninsula.

Figure 2.2. Cross-section through the Kola Peninsula (line A-B in Figure 2.1.). Note that this cross-section is more detailed than the map in Figure 2.1., and the keys to the two figures do not match. The scale (above) is approximate.

rifted blocks and major collision zones. Some of these faults, and fault-intersections have acted as channels for ascending magmas.

Alkaline magmatic activity during the Precambrian was restricted to the Middle Proterozoic (2.0-1.7 Ga) (Arzamastsev, 1994). Proterozoic intrusions included ultrabasite-gabbro-norite, charnockite-granite (which surrounds the Turiy massif), granodiorite, granite, alkaline granite, syenite, and alkali gabbroids (Arzamastsev, 1994).

Mitrofanov et al. (1993) described the tectonic history of the Kola Peninsula, summarized by Arzamastsev (1994):

“The north-east part of the Fennoscandian Shield consists of three provinces: Fenno-Karelian and Murmansk Archean cratons in the south and north respectively, and an extensive Lapland-Belmorian-Kola intracratonic collision zone of Archean - Early Proterozoic age between them. The geological evolution of the collisions is notable for the repeated occurrence of high-pressure granulite to amphibolite facies metamorphic assemblages at a time interval of 2.9, 2.4, and 1.9 Ga, vast areas of ultrametagenic pegmatites, and large anorthosite bodies (2.6, 2.4 and 1.9 Ga). Almost all Precambrian and Palaeozoic alkaline massifs, dykes and pipes are located within the collision zone”.

This description of the evolution of the Kola Peninsula can be expanded by studying the map in Figure 2.1. The northern, Archean gneissic block (Murmansk craton), contains a random mixture of the three main gneissic units of the Osnovanya Formation, cut by small late Archean granites and rare early Proterozoic granite, and Palaeozoic diabase intrusions. A small Palaeozoic intrusion, Ivanovskii, has recently been found just off the northern coast of the Kola Peninsula (Arzamastsev, pers. com. 1995). A second block, to the south, shows northwest-trending, alternating, early Archean gneissic and amphibolite units in the northern part. Towards the southeastern end of the block, an elliptical arrangement of Archean granites encircles younger, lower-grade Archean metamorphic rocks which are all cut by NW-trending Proterozoic diabase dykes. To the

south and east of the granite ring, late Archean gabbro-norites and granites (seen scattered across the northern block) truncate the early Archean basement rocks which, in turn, exhibit a younging trend from NW to SE. Amphibolites are less common in the southeastern than in the northwestern end of the block. Further southwest, the third block defines a rifted Archean basin, infilled by Proterozoic sediments which appear to have been deposited synchronously with rift extension and subsidence. A thick layer of intrusive peridotite, pyroxenite and gabbro (possibly a lopolith) underlies the northern part of the sedimentary basin infill. In a fourth block, to the southwest of the Proterozoic basin, the Archean basement reappears at the surface where it is covered by a thin layer of younger Proterozoic sediments or intruded by early Proterozoic granites (now charnockites and enderbites). This fourth, and southernmost block, is dominated by the Kandalaksha graben. The northwestern end of the graben contains late Archean granite intrusions and Archean plagioclase gneisses and amphibolites, which become older towards the southwest where Archean basement outcrops have been found. The southwestern corner of the graben contains an Archean granite intrusion adjoining early Proterozoic sediments.

The western part of the Kola Peninsula is dominated by Middle Proterozoic sediments and granodiorites to the north, followed by amphibolites and sediments, and middle Archean gneisses and amphibolites with minor granite further south.

2.1.b Palaeozoic

During the Palaeozoic, the Kola Peninsula was intruded by approximately twenty-five alkaline intrusions and several dyke suites which are now exposed at the surface due to subsequent tectonic uplift and erosion (Kogarko et al., 1995). The alkaline massifs range in size from 1 - 40 km in diameter, and in composition from pyroxenite-olivinite (e.g. Afrikanda) to ijolite-melilitolite-carbonatite (e.g. Turiy, Kovdor) to ijolite-syenite-

carbonatite (e.g. Khibina).

Emplacement of the massifs was probably controlled by pre-existing fractures within the lithosphere. A line of intrusions including Khibina and Lovzerzero runs from the north-western tip of the Kandalaksha Graben in a north-easterly direction to the northern coast, with the emplacement of this suite of intrusions controlled by the deep structures of the Kontozero Graben. Other massifs such as Afrikanda, Kovdor, and Turiy are controlled by the Kandalaksha Deep Fracture Zone associated with the northwest trending Kandalaksha Graben. This fracture zone, one of the most prominent tectonic features of the Baltic Shield, influences magmatism from Archangelsk (kimberlite pipes) to Finland (Sokli carbonatite), a distance of more than 300 km. It appears that the suite of intrusions associated with the first NE-trending fracture zone are dominated by syenitic magmatism, whereas those associated with the Kandalaksha Deep Fracture Zone are dominated by the pyroxenite-ijolite-melilitolite suite of silicate rocks. Carbonatites are associated with the silicate rocks in most of these intrusions. Most of the intrusions consist of sub-vertical pipes and stocks, commonly containing ring dyke or cone sheet intrusions. The sequence of intrusion of the various units within these massifs either trends from the periphery to the centre or from the centre to the periphery. Some intrusions, including Turiy, contain mixtures of these two intrusion sequences (Kapustin, 1980).

A recent Rb-Sr isochron study of many of the Palaeozoic alkaline complexes (Kramm et al., 1993) have placed them within a 23 Ma age bracket (358 - 381 Ma). There are few other known instances of carbonatite or alkalic magmatism occurring at this time in the rest of the world, apart from carbonatites found in the eastern part of the Omineca Belt in the Canadian Cordillera (Pell, 1994).

For further information on the Paleozoic complexes of the Kola Peninsula, the reader is referred to the books by Kukharensko et al. (1965) and Kogarko et al. (1995) and

the summary papers of Gerasimovsky et al. (1974) and Kogarko (1987).

2.2 Kandalaksha Graben

Magmatic activity along the graben was controlled by the Kandalaksha Deep Fracture Zone which was active mostly during the Palaeozoic. It is also believed to have been reactivated during the Alpine Orogeny. This last episode of tectonic activity is not associated with any known magmatism. Within the graben, there are five main areas of magmatic activity. The first area, at the south-east end of the graben, is represented by diamond-bearing kimberlite pipes found at Archangelsk, across the White Sea from the Kola Peninsula (Arzamastsev, 1994). In the middle of the northern edge of the graben, the Turiy Massif and the Terskii Coast pipes occur. The latter include nephelinites, olivine melilitites and diamond-bearing kimberlites, and have been described in detail by Kalinkin et al. (1993), Arzamastsev & Arzamastseva (1993) and Arzamastsev (1994) (Figure 2.3).

There are two main kimberlite diatremes (30x40m and 100x100m) which are enriched in pyrope, chrome spinel and chrome diopside, as well as minor apatite and perovskite. Diamonds have also been found. Closer to the Turiy Massif, a suite of 19 olivine melilitite dykes and 8 pipes have been located. Nephelinites occur to the north and north-east of the massif (Figure 2.3) (Kalinkin et al., 1993). At the north-west end of the graben, around the town of Kandalaksha, there is a large swarm of lamprophyric, alkaline and carbonatite dykes, as well as the Kandaguba massif (Pilipiuk et al., 1997). The petrological characteristics of the Kandalaksha dyke-swarm have been investigated by Russian geologists for many years (Bulakh & Ivanikov, 1984; Rukhlov & Ivanikov, 1997), and a recent and detailed geochemical study was undertaken by Beard et al. (1997). The minor archipelago of small islands at the top of the Gulf of Kandalaksha contains many excellent exposures of ultramafic lamprophyres (damkjernite) and

Figure 2.3 Map of the Terskii Coast (after Kalinkin et al, 1993). Note the spatial relationship between the "kimberlites", olivine melilitites and nephelinites in terms of increasing depth of origin within the mantle away from the Turiy massif

carbonatite dykes. K-Ar ages from the lamprophyres are 368 ± 15 Ma and 360 ± 16 Ma, while ^{40}Ar - ^{39}Ar data from phlogopite and amphibole contained within glimmerite and hornblendite xenoliths from a carbonatite dyke (similar to xenoliths found within the silicate dykes) give ages of 386 ± 1.0 Ma and 395.6 ± 4.4 Ma (Beard et al., 1997). Thus it seems likely that the dykes passed through an older intrusion within the lithosphere and incorporated xenolithic material from it.

2.3 Terskii Coast

The location of the pipes along the Terskii coast, just south-east of the Turiy Massif, are shown in Figure 2.3. Their age is uncertain, although K-Ar dates on the kimberlites provide ages of 337 - 384 Ma, and ages of 465 ± 35 Ma and 640 ± 50 Ma have been obtained from the olivine melilitites (Kalinkin et al. 1993).

Arzamastsev & Arzamastsev (1993) found chrome diopside and garnet and other kimberlite-indicator minerals within the pipes. One such sample, kindly provided for this study by Andrei Arzamastsev, is described in Appendix Biv. It contains rounded megacrystic olivines, colourless needles of phlogopite with very high Mg#s, and radiating clusters of microcrystalline apatite crystals, in an altered groundmass. The olivines are almost completely serpentized. Kalinkin et al. (1993) also reported the presence of diamond in these rocks, and has confirmed their identity as kimberlites.

The bulk chemistry of the olivine melilitites show unusually low CaO contents, but very low SiO_2 , high MgO, and relatively high alkali contents (Arzamastsev & Arzamastsev, 1993). These compositions are not matched by any bulk rock analyses obtained from samples from the Turiy massif, and related dykes on the Turiy Peninsula.

It appears that the mantle source of the pipes becomes deeper with increasing distance away from the Turiy Massif (i.e. nephelinites are closest to the massif, kimberlites furthest away) which implies that the lithosphere is thinner and the geotherm

is steeper beneath the massif. This is not particularly surprising given the Turiy lies at the edge of the graben, and is in itself a local horst block (Bulakh & Ivanikov, 1996).

Whether the Kandalaksha Graben represents a passive or an active rift (lithospheric flexure vs mantle plume) is a matter for debate, as is the question of whether the two rifting mechanisms have alternated over time. Lithospheric stretching in NE Africa above the Afar mantle plume, has been shown to be due to a combination of active and passive stretching mechanisms (Stewart & Rogers, 1996). The Rhine graben in Germany bears remarkably similarity to the Kandalaksha Graben, both in terms of geographical and tectonic features, as well as the type and degree of magmatism, and the relationship between geography and magmatic activity. The Kaiserstuhl carbonatite, in the Rhinegraben is similar in its tectonic setting to the Turiy Massif within the Kandalaksha Graben, and is underlain by a mantle dome (Edel, 1975). Thus the Tertiary-Quaternary magmatism within the Rhine graben is the result of active rather than passive rifting (Schleicher et al., 1990). Whether this is a similarity also shared by the Kandalaksha Graben in general, and the Turiy Massif in particular, is an issue that will be examined further later in a later chapter.

CHAPTER 3

THE TURIY MASSIF

3.1 Introduction

The Turiy Massif, located along the south coast of the Kola Peninsula at (34° 44' E, 66° 62' N) mid-way along the north coast of the Kandalaksha graben, forms a small peninsula which juts out into the White Sea. The massif has a total area of ~50 km², most of which is covered by 30 m of glacial till and other surficial material (Ronenson et al., 1978) (Figure 3.1). The massif only outcrops along the shore-line on the western coast, and in southern coastal exposures of the dyke swarms associated with the intrusion. The probable original level of emplacement of the massif was originally 1-2 km, i.e. a confining pressure of ~5 kbar, and the current depth to the base of the massif is 3-5 km (Bulakh & Ivanikov 1984). Samples from the main massif were obtained from drill core material. A geological map of the Turiy peninsula, based on detailed studies of the drill core material by Ronenson et al. (1978) is shown in Figure 3.2.

3.2 Previous work

Turiy first came to the attention of Russian researchers more than 150 years ago, and Betlink provided the first descriptions of the rocks in 1840. Interest in the massif resumed at the turn of the century, when Federov visited the Turiy coast between 1895-1905. In 1911 and 1914 and Brenner, Ramsey and Sederholm began to establish a collection of rocks from the area. Brögger was given part of this collection and subsequently published the first famous monograph on rocks from this area, including descriptions of turjaite and the carbonatites (Brögger, 1921). Thus, the first Russian carbonatites to be recognized and described were those from the Turiy Peninsula. Kranck (1928) published a more detailed description of Ramsey's sample collection, and this became one of the most classic pieces of writing published on Turiy.

Figure 3.1 View of the Turiy Peninsula from the top of a very wobbly sentry tower, abandoned by the army. It can be seen that the peninsula is covered in glacial till and subsequent forestation, making surface exposures a great rarity. The White Sea is seen in the distance, to the south-east.

A lull in investigative activity was broken in the late 1960s and early 1970s by Dr Andrei Bulakh and his colleagues, including his PhD student, Valeriy Ivanikov, from Leningrad State University. In 1984 they wrote the most comprehensive recent publication on Turiy entitled “The problems of mineralogy and petrology of carbonatites” (in Russian), based on material from the coastal exposures and the extensive drill core material that was obtained in the early 1970s by a Russian mining company. A description of the cores was published by Ronenson et al. (1978). Professor Keith Bell (Carleton University) visited the Turiy peninsula with Dr Ivanikov in the summer of 1991 and instigated the present study. In the summer of 1995 additional samples were collected by myself. This study represents the first major study of Turiy carried out by western geologists.

Other papers on Turiy published in English include Dmitriyev et al. (1970), Mal'kov (1970), Bulakh et al. (1972), Vlodavetz (1972), Ivanikov (1973), Bulakh et al. (1973, 1975), Bulakh & Mazalov (1974), and Panina & Podgornyykh (1974).

3.3 Geological outline

Five intrusive/hypabyssal centres are found at the surface (Figure 3.2). A central complex (28 km²) is surrounded by satellite complexes of varying sizes to the north (~1 km²), south (13 km²), east (10 km²) and the Kuznavalok complex to the west (1.3 km²). The northern complex represents a suite of small, scattered bodies, as seen in Figure 3.2. Several dyke swarms occur around the south and south-west coasts of the peninsula, and were intruded in three phases, pre-, syn- and post-massif emplacement (Bulakh & Ivanikov, 1984).

A variety of rock types occur within each complex, and melilitolites are common in all five complexes. Turiy is also host to two type-locality rocks - turjaite and turjite, which are described in detail in subsequent chapters. The massif was emplaced into

Figure 3.2 Geological map of the Turiy Peninsula (after Ronenson et al., 1978).

Proterozoic charnockites overlain by Riphean sandstones along the south coast of the Turiy Peninsula (Bulakh & Ivanikov, 1984; Kogarko, 1987). Fluids emanating from the massif have created a 500-700m-wide fenitized aureole within the adjacent country rock. Bulakh & Ivanikov (1984) report the presence of microcline-aegirine, albite-rich, and feldspar-aegirine-nepheline fenites. On the basis of findings by Bulakh & Ivanikov (1984) and this study, the Turiy rocks can be divided into the following groups:

i) Pyroxenites: These rocks occur only within the central massif and were previously assumed to be cumulus rocks, although true cumulus textures (normally adcumulate) are not ubiquitous. Their mineralogy is dominated by clinopyroxene, with accessory amphibole, garnet, calcite, apatite, perovskite and magnetite. Some of the pyroxenites are ~200 Ma older than the rest of the massif, as described in a later chapter. Olivinites, reported as relicts within the pyroxenites (Ronenson et al., 1978), have not been examined in this study. Kapustin (1980) described pyroxenites from a number of Kola Peninsula massifs including Vuoriyarvi, Kovdor and Afrikanda.

ii) Ijolite-melteigite (1st series): Ijolite series rocks are particularly common components of hypabyssal carbonatite complexes in Kola and in other parts of the world (Kapustin, 1980). Detailed studies of drill core material have suggested that there are two ijolite series within Turiy (Bulakh & Ivanikov, 1984). The first series is more mafic and pyroxene-rich, and occurs both within the central massif, and as an outer ring around all the complexes (Ronenson et al., 1978). It may also occur below the high-water mark on the shore at Kusnavalok, as part of the western massif. Pyroxene and nepheline $\pm$ garnet dominate the mineralogy of these samples. Borodin (1963) proposed the following division of ijolite bodies in such alkaline massifs: a) arcuate zones around the ultrabasic rocks; b) independent massifs and c) dykes. The rocks of the first ijolite series at Turiy appear to fall primarily in category a), as is typical of hypabyssal massifs containing an ultramafic core (Kapustin, 1980). The arcuate bodies (ring dykes?) seen at Turiy are

similar to those found at Kovdor, Poschni, Vuoriyarvi and several other massifs, according to descriptions by Kapustin (1980). A porphyritic olivine melteigite pipe found in the north-western part of the southern complex is believed to be synchronous with the 1st ijoite series (Bulkah & Ivanikov, 1996).

iii) Melilitolites: One of Turiy's two type-locality rocks, turjaite, is also the most abundant rock in the massif. Turjaite was originally defined by Ramsay (1921) as an intrusive melilite-nepheline rock, with accessory phlogopite, magnetite, perovskite, apatite and calcite. A number of other accessory minerals have been found in turjaite samples examined in this study (Appendix Biii). The type-locality is at Kusnavalok on the western shore of the western massif. Turjaite makes up a significant portion of all five complexes within the Turiy Massif, and is surrounded by ijolites in all locations except the western complex. Turjaites are also abundant in the Kovdor complex where they appear to pre-date the arcuate ijolite bodies which surround them (Kapustin, 1980). Ronsenson et al. (1978) and Bulakh & Ivanikov (1996) report the presence of rare uncompahgrites (pyroxene melilitolites) and olivine-melilitolites within the central complex at Turiy, as well as melilitolites rich in perovskite and magnetite, similar to the melilitolites of the Gardiner Complex in Greenland (Nielsen, 1980). Calcite melilitolites grading into melilite-calcite carbonatites have also been found in Turiy's northern complex in this study. Bulakh & Ivanikov (1996) also report the presence of okaite in Turiy, but do not mention finding hauyne, one of the pre-requisites in the definition of type-locality okaite. No hauyne has been found in the melilitolites examined in this study. It is possible that Bulakh & Ivanikov (1996) were using the definition of okaite by Mikhailov et al. (1995) as a rock consisting only of nepheline + melilite. Kukharensky et al. (1965) described all melilitolites from the Kola Peninsula as "metasomatic", but results from this study suggest that this description is not applicable to the majority of the Turiy melilitolites.

iv) Ijolite-uritites (2nd series): Drill core studies (Ronensen et al., 1978) showed the existence of a second, more felsic ijolite series, confined to the middle of the central massif. The modal mineralogy of these samples is dominated by garnet + nepheline $\pm$ phlogopite $\pm$ pyroxene. These rocks are also described as “skarn rocks” by Bulakh & Ivanikov (1984) due to the effects of the metasomatic fluids from the younger carbonatites.

v) Phoscorites: The mineralogy of these rather unusual rocks is dominated by magnetite, phlogopite, apatite and calcite, along with pyroxene, accessory sulphides, and other minerals (Appendix Biii). Phoscorite only occurs in the central massif, and forms arcuate bodies within the central parts of the carbonatite stock (Figure 3.1). Bulakh & Ivanikov (1984) provided evidence that the phoscorites are older than the carbonatites, although the phoscorites may have intruded during a period between emplacement of the early silicate mineral - rich carbonatites and the youngest “pure” carbonatites. Most phoscorites from Kovdor, Vuoriyarvi and elsewhere are dominated by magnetite-forsterite assemblages which occur in Vuoriyarvi as vein networks or as stocks > 150m thick (Kapustin, 1980). Forsterite was not found in the samples examined in this study, although it was reported in the Turiy phoscorites by Bulakh & Ivanikov (1984). Whether the magnetite-apatite-calcite-phlogopite assemblage seen in the Turiy examples is truly magmatic, or whether it represents replacement of an original magnetite-forsterite assemblage by “apatitization” and “calcitization” processes (Kapustin, 1980), is difficult to determine. Their origin as hydrothermal veins (Kukhareenko, 1965) is unlikely at Turiy, given their isotopic signatures.

vi) Carbonatites: These rocks, extremely variable both in terms of their modal mineralogy and chemical composition, occur mainly in a veined stock-work in the middle of the central massif, although a few carbonatites also outcrop in the northern satellite body, in association with melilitolites. Most of the carbonatites are calcite carbonatites.

Bulakh & Ivanikov (1996) have divided the carbonatites into three types: forsterite-calcite, pyroxene-calcite and mica-calcite. No forsterite-calcite carbonatites were examined in this study. Common minerals found in the northern complex calcite carbonatites include pyroxene, melilite, phlogopite, magnetite, and occasional garnet and perovskite. The central complex calcite carbonatites contain $^{iv}\text{Fe}^{3+}$ -phlogopite and/or amphibole with abundant apatite and magnetite and rare pyroxene, but no melilite or garnet. Calcite-dolomite and dolomite carbonatites are found in the central complex representing the latest stages of carbonatite magmatism, and contain accessory apatite and sulphides. Accessory minerals found in the carbonatites in this study are listed in Appendix Biii, and include baddelyite, monticellite, pyrochlore and zirsinalite.

Relict xenoliths of silicate rocks from deeper within the massif are sometimes carried within the carbonatites (e.g. Figure 5.2a). Late-stage veins on the south coast of the Turiy Peninsula include ankerite-fluorite-quartz and aegirine-apatite-ankerite assemblages, typical of late hydrothermal-carbonatite activity in such massifs (Kapustin, 1980).

vii) Coastal dykes: Three periods of dyke activity are represented on the south coast of the Turiy Peninsula, totalling ~300 dykes (Bulakh & Ivanikov, 1996) which were emplaced before, during and after the intrusion of the main massif (Bell et al., 1996; Bulakh & Ivanikov, 1996). The first dyke series includes alkaline picrite, alnöite, calcite-bearing ultramafic lamprophyre (damkjermanite), carbonatite, and melanephelinite, as well as a few eruptive breccias. Spinel harzburgite, eclogite and granulite xenoliths are found in some of these dykes (Ivanikov, 1973). The second phase of dyke emplacement occurred at the same time as the emplacement of the main massif, and consists of several generations of micromelteigite-microijolite, two generations of turjite dykes and a pipe-like body of olivine melteigite found to the north-west of the southern complex. The post-massif-intrusion dykes are mostly nephelinites, melilitites and carbonatites. The

carbonatite dykes show little textural or mineralogical similarity to carbonatites associated with the main complex, and are belong to the youngest suite of dykes (Bell et al., 1996).

CHAPTER 4

PETROGRAPHY

4.1 Introduction

This chapter describes the results of a petrographic examination of the samples from the Turiy Massif used in this study, categorized according to rock type. Modal mineralogy and detailed petrographic descriptions of each of the samples studied can be found in Appendices Biii & Biv respectively. A summary of the modal mineralogy of the rock types present within the intrusion is shown in Table 4A.

Given that many of the rocks found in Turiy, including two type-locality rocks, are rare on a world-wide scale, it is important to provide comprehensive descriptions of the samples studied, above and beyond the tabulated descriptions provided in the Appendices. Thus, the principal aim of this chapter is to compare and contrast the petrography of samples within and between different rock types, in order to create a basic understanding of the factors which characterize the petrography of the wide variety of rock types found at Turiy. Such comparative descriptions provide a framework on which to base the interpretation of the mineralogical and chemical data presented in the following chapters.

The most significant feature to emerge in this study is the extreme textural variability and heterogeneity shown by all of the rocks collected from the drill core samples from across Turiy (Figure 4.1b). Thus, although an attempt has been made to sample rocks from a wide variety of locations and rock-types, these samples are, at best, only a rough guide to the petrography of the body from which they came. Heterogeneity and disequilibrium are the norm (Figure 4.2).

No specific hand-specimen descriptions are provided here. Instead, the majority of the mineralogical and textural details listed in this chapter have been gleaned from thin-section work, and hand-specimen characteristics are referred to where necessary.

Table 4.A   **Summary of the modal mineralogy
of the main rock types of the Turiy Massif**

Rock type	Major minerals	Accessory minerals*		Trace minerals**
		silicate	non-silicate	
pyroxenite	pyroxene	amphibole garnet phlogopite	calcite magnetite perovskite	apatite
melteigite-ijolite (1st series)	pyroxene nepheline	cancrinite garnet phlogopite	apatite perovskite	banalsite stronalsite sulphides titanite
melilitolite (turjaite)	melilite phlogopite nepheline magnetite	cancrinite garnet pyroxene	apatite calcite perovskite	copiapite cuspidine djerfisherite dolomite pectolite strontianite
ijolite-urtite (2nd series)	nepheline pyroxene	garnet phlogopite	apatite	banalsite calcite ilmenite pectolite titanite
phoscorite	magnetite apatite calcite	amphibole phlogopite pyroxene	dolomite sulphides	ilmenite strontianite zircon
carbonatite	calcite (dolomite)	amphibole garnet melilite monticellite phlogopite pyroxene	apatite magnetite perovskite sulphides	baddelyite barite calzirtite ilmenite strontianite titanite zircon zirsinalite
turjite	garnet phlogopite calcite	analcime celsian	magnetite melilite?	apatite ilmenite perovskite pyroxene strontianite titanite

* modal abundance of > 1% in at least one sample

** modal abundance of < 1% in at least one sample

Figure 4.1a A curator's nightmare... Field-work, Turiy -style! Alexy Ruklov takes a break from supper preparations (mushroom picking) to ponder exactly which sample to choose from the melilitolite section of the drill core site, Turiy Massif. Beneath the surficial layer of unlabelled cores, formed as a result of gravitational sorting, labelled boxes of undisturbed material occur.

Figure 4.1b Undisturbed drill core box from the drill core site, Turiy Massif, exposing about 12 m of sequential core samples. Note the extreme compositional and textural heterogeneity within such a small section of core material. This heterogeneity was common to virtually all the boxes examined in the summer of 1995.

Figure 4.2a Drill core sample from the central massif. It contains brecciated urtite, veined by carbonatite with the contact between the two rock-types lined by a thick band of phlogopite (glimmerite). Grey circle (bottom right) = 3 cm diameter.

Figure 4.2b Sample C.190.x from the north-east edge of the central massif shows a contact between the chilled margin of an ijolite (dark brown band, right hand end of the sample) and the surrounding country rock (majority of the sample). Notice the elongate crystals of dark green aegirine, many of which are oriented perpendicular to the contact between the two rock types, and whose size increases away from the contact. Grey circle (bottom) = 3 cm diameter.

Transmitted and reflected light microscopy and cathodoluminescence spectroscopy were used to identify the minerals and characterise the textures within the samples. EDS spectra (and analytical data) from electron-microprobe analyses were used to identify minerals too small to be identified using optical microscopy techniques. The analytical procedures for the techniques listed above are described in detail in Appendix D. The terms “fine-grained”, “medium-grained”, and “coarse-grained” are used to describe minerals which are < 1mm, 1-5mm, and > 5mm in diameter, respectively, throughout this chapter.

The rocks have been named and classified using the igneous rock classification scheme of Le Maitre (1989), except for the melilite-bearing rocks which have been classified using the scheme outlined in Appendix C. The sections below are listed by rock-type, and related to the presumed order of intrusion, from oldest to youngest (Ronenson et al., 1978; Bulakh & Ivanikov, 1984, 1996; Bell et al., 1996). These interpretations of the age relationships are based on drill-core field evidence, which itself exhibits many contradictory cross-cutting relationships, and hence provides somewhat ambiguous results regarding the sequence of magmatic activity. The description of the olivine melteigite pipe, whose age is uncertain, is given prior to the ijolite section, because it bears the greatest petrological resemblance to the ijolite series rocks.

4.2 Pyroxenites

Two samples of pyroxenite [C.AG.347 and C.90.29] from the Central massif contain pale yellow/green, medium- to coarse-grained, equigranular pyroxenes, mostly anhedral-subhedral in shape. Interstitial accessory minerals include amphibole, apatite, garnet, magnetite, nepheline and perovskite. The cores of many pyroxenes from sample C.AG.347 contain euhedral exsolution lamellae of chrome spinel (Figure 4.3b). These exsolution rods are not developed in the diopsidic overgrowths around the cores, and the

Figure 4.3a Clinopyroxene grain from the pyroxenite sample C.AG.347, showing exsolution lamellae of chrome spinel, rounded inclusions of magnetite, and a diopsidic overgrowth rim (top of the picture) free of inclusions or exsolution lamellae. Photograph taken under crossed polars, field of view = 0.75 mm.

Figure 4.3b Pyroxenite sample C.AG.347 showing a cluster of clinopyroxene grains with interstitial magnetite. Notice the variation in birefringence between the core and the rim of the clinopyroxene grains, indicating changes in composition. The cores of several of the grains also contain chrome spinel exsolution lamellae and magnetite is also enclosed within several of the grains. Photograph taken under crossed polars, field of view = 2 mm.

optical extinction patterns of the pyroxenes are discontinuous between core and rim. The cores are mostly anhedral, and may have been partially resorbed by the enclosing magma prior to the crystallization of the overgrowths (Figure 4.3b). Interstitial minerals in this sample include Cr-poor magnetite which also occurs as blebs within pyroxenes, pargasitic amphibole, perovskite, calcite and apatite. The latter four minerals commonly occur in clusters.

Kapustin (1980) discussed the origin of perovskite in pyroxenites in Kola. Perovskite was assumed to either be part of the primary pyroxenite mineral assemblage (Kukharensky et al., 1965; Volotovskaya, 1960), or related to alkali metasomatism (Borodin, 1963). If the origin of perovskite in the Turiy pyroxenite samples is due to alkali metasomatism, along with the formation of magnetite, calcite and amphibole, then isochron evidence from this study suggests that such metasomatism was contemporaneous with crystallization. Ronenson et al. (1978) report the presence of interstitial melilite in the “pyroxene rocks”, grading into almost monomineralic melilite areas, with rare nepheline. The presence of rare melilite in pyroxenites was also noted by Larsen (1942) and Nash (1972) at the Iron Hill Complex in Colorado, and in pyroxenite xenoliths in the Oldonyo L’Engai magmas (Donaldson & Dawson, 1978).

It is possible that the cores of the pyroxenes represent early cumulus material crystallized within an underlying magma chamber, while the rims represent material crystallized during ascent and emplacement (i.e. adcumulates). The origin of the Cr-spinel lamellae is presumed to relate to high-pressure exsolution early in the history of the massif (Ercit, pers. comm. 1994). However, it seems unlikely that the underlying magma chamber would have erupted cold, dense, pyroxene-rich cumulus masses from its base as the first “magma” to be intruded up into the crust and form the initial stages of the emplacement of the Turiy Massif. An alternative explanation may lie in the work of Tilley & Yoder (1968) who converted two nephelinites and a natural melilite-bearing

rock, in the presence of volatiles, at 2-10 kb pressure, to hornblende pyroxenites. According to the geological map of Bulakh and Ivanikov (1984), the pyroxenites at Turiy were intruded into the central complex as elongate bodies whose long axes lie approximately parallel to the nearby north-east edge of the Kandalaksha Graben. Rb-Sr and Sm-Nd isochron data from Chapter 6 also indicate that at least some of the pyroxenites are considerably older than the rest of the intrusion, which adds further complexities to the problem of their origin.

4.3 Olivine melteigite

A small body of hypabyssal olivine melteigite [C.AG.48A] is found on the eastern edge of the central complex which Bulakh & Ivanikov (1984) considered to be of similar age to the early ijolites.

The melteigite contains abundant coarse-grained, rounded olivines and inclusion-rich, zoned, green-brown augite phenocrysts (Figure 4.4a). The origin of the olivines is uncertain as they show reaction rims and magnetite overgrowths, and contain abundant inclusions. They may be cumulate or possibly xenocrystic in origin. Nepheline forms most of the groundmass, poikilitically enclosing rare, euhedral and fine-grained aegirine crystals. A similar crystallization sequence is seen in melilitite magmas from Germany (Dunworth and Wilson, submitted). One inclusion from the core of a pyroxene crystal contained magnesio-hastingsite amphibole, similar to the pargasitic amphibole found in the pyroxenite sample, C.AG.347

4.4 Ijolite series

It has been postulated that there are two phases of ijolite intrusion at Turiy (Bulakh & Ivanikov, 1984). The outer rims of the central, southern, eastern, and part of the northern complexes consist of rocks of the "first ijolite series" that may represent the

Figure 4.4a Olivine melteigite sample C.AG.48A showing rounded, coarse-grained olivine (top left) with irregular rims and an abundance of magnetite around the rims, and abundant augite with grain size ranging from <1mm to 4mm and granularity ranging from anhedral to euhedral. Interstitial nepheline (top right) contains rare crystals of deep-green aegirine-augite (centre right). Rare mica can also be seen (bottom left). Photograph taken in plain polarized light, field of view = 2 mm

Figure 4.4b Urtite sample C.IJ.14 showing garnet (dark brown), $^{iv}\text{Fe}^{3+}$ -phlogopite (orange) enclosing and replacing clinopyroxene (green), all contained within abundant nepheline (colourless). Significant quantities of primary pectolite also occur in this specimen (not shown). Photograph taken in plain polarized light, field of view = 2 mm

first intrusive phase of the main intrusion (Bulakh & Ivanikov, 1984). The “second ijolite series”, intruded after the melilitolites, is primarily seen within the central complex surrounding the central carbonatite stock. Contradictions appear between the maps of Bulakh and Ivanikov (1984) and the drill-core maps published by Ronenson et al. (1978) particularly in relation to the location of the second ijolite series. Examination of drill-core material at the drill-core sites on the Turiy Massif by the author in the summer of 1995 provided inconclusive evidence as to the true nature of the intrusion locations, due to the incomplete nature of the remaining core samples and the inadequate identification of the original location of the majority of the cores. What does appear to be clear is that the ijolitic material immediately adjacent to the central carbonatite stock has been metasomatized and altered by fluids emanating from the younger carbonatites. The nature of the original mineralogy of the central massif ijolites, and associated rocks, is thus not well preserved.

Most samples (except WKM) are medium- to coarse grained and are normally texturally homogeneous within the scale of the samples collected. Typically, the ijolites are rich in pale green diopsidic pyroxenes, nepheline and melanite garnet. Garnet can become more abundant than pyroxene, especially in the central complex samples. Phlogopite is less common and varies from Ti-poor Al-rich (pale green) to $^{iv}\text{Fe}^{3+}$ -rich Al-poor. The latter composition exhibits orange pleochroism and reverse absorption. Apatite, magnetite, ilmenite, perovskite and titanite also occur. Calcite, banalsite, stronalsite, analcime, pectolite, sodalite, wollastonite occur as accessory minerals in some samples, and zeolites and cancrinite are common alteration products.

The “first ijolite series” is represented by samples C.AG.122 which is a melteigite from the outer ring of the Central complex, E.123.106 which is an ijolite-urtite from the southern edge of the Eastern complex, and S.162.300 which is an ijolite-melilitolite contact sample from the southern massif. Sample C.AG.122 is coarse-grained and

contains the most abundant pyroxene and least nepheline of all samples studied. Only two of these samples contain iron oxide minerals, and phlogopite is also rare. Sample E.123.106 contains the most abundant nepheline of all the samples studied, and is much closer in composition to the “second ijolite series” urtites from the central massif which are described below.

A small fine-, to medium-grained, homogeneous micro-ijolite body below the high-water mark on the western shore [WKM] is difficult to relate to any of the other ijolites. The western complex is the only part of the complex not to have an ijolite “ring” between the melilitolites and the country rock according to the map of Bulakh & Ivanikov (1984). However, this is not in agreement with the map of Ronensen et al. (1978) which also shows an ijolite ring around the west coast melilitolites. Sample WKM bears little resemblance, in terms of modal mineralogy, to any of the other ijolite samples, and its ratios of [pyroxene + garnet + mica] : nepheline, and garnet : mica are both approximately 1:1.

The ijolites enclosing the carbonatite stock in the middle of the central massif are believed to be younger than the melilitolite intrusions, and are commonly described in the Russian literature as “skarns”, referring to the intense metasomatic features which they sometimes exhibit (Bulakh & Ivanikov, 1984). Samples from this ijolite body include C.ND.500 and C.IJ.14 (locations unknown), C.170.38 (eastern side of the carbonatite stock) and C.90.y (south of the carbonatite stock). All four samples contain about 50% nepheline (or related alteration products), most contain magnetite and phlogopite, and the combined modal percentage of pyroxene + garnet + phlogopite is between 32-42%. Garnet is most abundant in samples C.IJ.14 (Figure 4.4b) and C.170.38. The former sample shows many similarities to turjite, as will be discussed in Section 4.8. Sample C.ND.500 represents an ijolite from an ijolite-melilitolite contact zone.

Also of note are the ijolite-fenite contact samples discussed in Section 4.10, one of which comes from the NNW side of the Eastern stock. The ijolitic parts of these samples are extremely similar to ijolites C.ND.500 and E.123.106, in terms of their petrology. Ronenson et al. (1978) reported the inclusion of relict fenites within the outer ijolite rings around the complexes.

Sample S.162.300 is from the southern side of the Southern stock, and appears to be representative of a contact between the ijolite and melilitolite bodies. The boundary between the ijolite and melilitolite parts of the sample is not well-defined in terms of textures, mineral associations and compositions. Medium-grained wollastonite is present although its origin is unclear. Roughly equal amounts of pale green pyroxene and nepheline are present in the ijolite part of the sample. The melilite contains small inclusions of ${}^{\text{iv}}\text{Fe}^{3+}$ -phlogopite, not found in any of the other melilitolites in Turiy. Ronenson et al. (1978) reported the melilitolites and outer-ring turjaite as being “in equilibrium” along the contacts between the two rock-types, and hence suggested that the two rocks were intruded at the same time. There is certainly no sign of a chilled margin in sample S.162.300, a characteristic common to many melilitolite - ijolite contacts within the massif (Bulakh, pers. comm. 1994).

One remaining problem is whether the outer ijolite rings in the eastern and southern complexes were intruded earlier or later than the melilitolites which they enclose. If the magmas were intruded as balloon-like masses, then presumably the outer ijolite “rings” were intruded first, followed by the melilitolites, with each new magma pulse intruding inside the previous one (a mechanism seen in many granite bodies). If this is the case, then the outer ijolite bodies should show structural deformation features consistent with the subsequent intrusion of the large melilitolite bodies into their midst. On the other hand, if the melilitolites were intruded first (i.e. directly into the country rock, as is suggested by the isotope data), then the resulting contact between the

melilitolites and the surrounding country rock would provide a zone of weakness along which an ijolite ring-dyke could readily intrude at a later time. If the latter model is correct, then this implies that there may only have been one phase of ijolite magmatism at Turiy (post-melilitolite, pre-phoscorite). Alternatively, the gradual contact between the melilitolites and the outer ijolite ring in the southern massif seen in sample S.162.300 may, instead, be indicative of magmatic differentiation. Treiman & Essene (1985) encountered similar problems in interpreting inconclusive field evidence from the Oka complex, and suggested that ijolite could form from uncomphagrite (pyroxene melilitolite) by the replacement of magnetite by garnet + perovskite, and by the replacement of melilite by diopside + nepheline $\pm$ wollastonite $\pm$ calcite, as textures in the Oka rocks preserve such reactions frozen in progress.

In summary, the ijolite series rocks (melteigite-ijolite-urtite) contain nepheline and pyroxene as essential and variably abundant minerals (Le Maitre, 1989). Commonly, the place of pyroxene is taken by Ti-rich garnet, especially in the central complex samples. Iron oxides and micas are not common, but become more significant as the percentage of pyroxene and garnet decreases. The issue of the number of phases of ijolite intrusion at Turiy remains unresolved. The modal mineralogy suggests a division between pyroxene-rich ijolites occurring as rings around the complexes (1st series), and garnet-rich ijolites around the central carbonatite stock (2nd series).

4.5 Melilitolites

The presence of melilite in many of the intrusive rocks in Turiy is a distinctive feature of the massif, which contrasts with carbonatite-syenite intrusions such as Khibina to the north where melilitolites are extremely rare. Other melilitolite intrusions in the world include the Kovdor Massif (Kola), the Gardiner Complex (Greenland), the Oka Complex (Quebec), the Rangwe Massif (Tanzania), the Iron Hill Complex (Colorado)

and several intrusions (e.g. Kugda) from the Maimecha-Kotui province (Siberia). For further details, the reader is referred to Dunworth & Bell (submitted).

4.5a Melilitolite classification

Turiy is host to two type-locality rocks, turjiate and turjite. Turjaite is a melilitolite containing essential melilite and nepheline, along with phlogopite, magnetite, apatite, calcite and garnet. Perovskite is a common mineral, but constitutes < 5% of the modal mineralogy of all the samples investigated. Olivine is virtually absent from the melilite-bearing rocks at Turiy, monticellite is rare, and pyroxene occurs in greater than trace quantities in only 2 out of the 10 samples investigated. The most recent classification scheme for intrusive melilite-bearing rocks proposed by the IUGS (Woolley et al., 1996) is based on a pyroxene-olivine-melilite ternary plot, and is clearly not applicable to these samples. An alternative scheme has been created, based on the modal mineralogy of melilitolite intrusions from around the world (Dunworth & Bell, submitted). The scheme will be used to classify all samples described in this study and is reproduced in Appendix C. The relationship between modal mineralogy and proposed rock names in this scheme is shown in Table 4.B.

4.5b Turiy melilitolites

Samples WKT, TUR I and TUR II all come from Kusnavalok on the west coast, which is the type-locality for turjaite, a rock type proposed and defined by Krank (1928). However, there is large variation in the modal mineralogy of the three samples. TUR II contains 65% melilite, along with nepheline, pale green phlogopite and two generations of both perovskite and magnetite. Apatite, calcite and garnet are present in trace quantities only. The later generation of colourless perovskite occurs as thin (average < 10 μm wide) reaction rims, completely enclosing the early-formed, coarse-grained

Table 4.B   Mineral assemblages in melilitolites and definition of specific root names

Group	Dominant mineralogy	Defining mineral (>10%)	Specific root-name
A	silicate	mafic	"copiamelilite" uncompahgrite kugdite
		felsic	turjaite okaite
B	oxide	perovskite	afrikandite
C	carbonate (primary)	carbonate < 50%	<i>e.g. calcite-melilitolite*</i>

*No specific root-name is provided for samples containing > 10% and < 50% primary carbonate

magnetite. There are overgrowths of pale green, Al-rich phlogopite around the perovskite-covered magnetite crystals. The pale perovskite rims found around the magnetite are not found around the earlier generation of medium-grained, deep brown perovskite, which suggests that the pale perovskite rims around the early magnetite are the product of a reaction. Fine-grained, anhedral magnetite grains, usually located at silicate grain corners, do not exhibit perovskite rims, and appear to have crystallized later in the cooling sequence. Thus, the earlier generation of deep brown, inclusion-rich perovskite crystallized prior to, or at the same time as the coarse-grained magnetite. Reaction of the coarse-grained magnetite with the host melt produced perovskite rims around the magnetite, followed by the crystallization of fine-grained, late-stage anhedral magnetite.

Sample WKT contains some irregular layering, which is seen more clearly in nearby exposures (Figure 4.5). Within the bands, sample WKT is a fine-grained, homogeneous and unaltered turjaite and shows the greatest enrichment in magnetite of all the samples studied. A plot of mica vs magnetite (Figure 4.6) suggests that there may be an antipathetic relationship between mica and magnetite in all the melilite-bearing samples studied, except WKT, indicating that the enrichment in sample WKT is unusual and fairly extreme. According to the classification given in Appendix C, this sample is a magnetite turjaite.

Sample TUR I is almost pegmatitic, containing abundant phlogopite, primary cancrinite (instead of nepheline), and rare anhedral diopside, commonly associated with calcite or enclosed within melilite. It can be considered as a mica turjaite (Figure 4.7a). The largest, inclusion-free apatites yet obtained from Turiy, were also found in this sample.

An interesting relationship between andradite garnet and melilite is seen in sample TUR I. Although a few deep brown, anhedral melanite garnet crystals are found in TUR

Figure 4.5 (left) Layered outcrop of turjaite at the type-locality, Kusnavalok, on the west coast of the Turiy Peninsula. Homogeneous outcrops also occur.

Figure 4.6 (above) Plot of mica vs magnetite modal abundances in the melilite-bearing rocks. Notice an antipathetic trend between the two variables outlined by all samples except WKT (12, 30) which is unusually enriched in magnetite. Errors estimated at $\pm 3\%$.

I, garnet is also found as colourless, anisotropic, fine (5-10 μm) rims around melilite crystals, obviously part of a replacement/alteration reaction. These two morphological forms of garnet may be continuously linked. Garnet rims around melilite have previously been documented from the Oka Complex in Canada (Watkinson, 1971). Garnet overgrowths are also found around the perovskite rims around magnetite in this sample.

The two melilitolites from the Central massif [C.5.118, C.190.105] are a medium- to fine-grained, banded mica turjaite (Figure 4.7b), and a fine to medium-grained turjaite, respectively. Both show little alteration. Phlogopite exhibits deep brown to beige pleochroism in both samples. Ronenson et al. (1978) report the presence of olivine enclosed in melilite in some of the central massif turjaites, a finding which could not be corroborated in this study.

The turjaite from the Eastern complex [E.133.30] is highly altered and contains diopside, similar to that found in sample TUR I. Two melilitolite samples were also investigated from the southern complex, one of which, sample S.78.15, shows extensive alteration, mostly in the form of unidentified hydrous alkali-rich phases, and the other is an ijolite-melilitolite contact sample. Both are described in Section 4.10.

Sample N.64.18m is a calcite-magnetite-melilitolite. Similar samples with up to 50% calcite are also reported from the Oka complex (Treiman & Essene, 1985). Fluid metasomatism and/or subsolidus breakdown of melilite in this sample, near to the contact with carbonatite sample N.64.18c, has produced diopside and vesuvianite and a variety of unidentified hydrous minerals. Sample N.64.18m was photographed using cathodoluminescence spectroscopy, as shown in Figures 4.8[a&b]. A band of magnetite crystals marks the contact between the two rocks. Such concentration of Fe oxides at contact zones involving carbonatite was noted by Pozharitzkaya (1966). About 10 cm away from the contact, the effects of hydrous metasomatism within N.64.18m are less obvious but a reaction involving melilite, magnetite, perovskite, calcite and monticellite

Figure 4.7a Melilitolite sample TUR I exhibiting two grains of phlogopite showing varying zoning patterns due to different optical orientation (centre and top right), a large apatite grain (bottom right), with interstitial cancrinite (dark blue) and melilite (top left). The section has been cut slightly thick, hence the high birefringence colours for the melilite and apatite. Photograph taken under crossed polars, field of view ~ 2mm.

Figure 4.7b Melilitolite sample C.5.118 exhibiting anhedral melilite (blue and yellow interference colours) with phlogopite, magnetite and nepheline. Photograph taken under crossed polars, field of view ~ 2mm.

Figure 4.8a Melilite-calcite carbonatite N.64.18m, < 5mm from the contact with the calcite carbonatite N.64.18. The field of view represents a cathodoluminescence picture of a metasomatized crystal of melilite surrounded by calcite, which exhibits orange luminescence due to Mn^{2+} activation (Marshall, 1988). Many of the minerals which have replaced the melilite are too small to be identified. However, most of the brown material just inside the rim is garnet, the red crystals are pure diopside, and the brown material intergrown with the diopside is vesuvianite. The rest of the material is mainly hydrous alkaline earth and alkali silicates. It is possible that the cluster of small, yellow crystals (upper part of photograph, right of centre) are pectolite or wollastonite. Field of view 1.5mm.

Figure 4.8b Melilite-calcite carbonatite N.64.18m, < 3mm from the contact with the calcite carbonatite N.64.18. The field of view represents a cathodoluminescence picture of diopside (red) intergrown with vesuvianite (brown). The bright yellow crystals may be wollastonite or pectolite. Photograph taken using cathodoluminescence microscopy. Field of view = 0.5mm.

appears to be in frozen in progress around a number of magnetite crystals, creating a “postage-stamp” texture between calcite and melilite and/or magnetite (Figure 4.9a). A cartoon of the minerals and textures found in this reaction is shown in Figure 4.9b.

Huckenholz et al. (1976) investigated the stability of Ti-bearing andradite with respect to temperature and oxygen fugacity. Breakdown products in the experiments included wollastonite, kirschenite (Fe-monticellite) and magnetite. A possible invariant point involving garnet, wollastonite, kirschenite, magnetite, liquid and vapour occurred at 940-1020°C and $\log fO_2$ -8.5 (precise temperature and oxygen fugacity dependent on original garnet composition) which is in between the MH and NNO oxygen buffers. The cartoon in Figure 4.9b shows a band of monticellite, but this band may also include some wollastonite. The presence of perovskite, seen in Figure 4.9b, is not accounted for in the experimental work of Huckenholz et al. (1976). However, if the thin perovskite rims are found around magnetite in all the melilite-bearing rocks in Turiy, then the perovskite blebs seen in Figure 4.9b may represent the original rim around the magnetite grain which has since been partially resorbed and involved in the formation of garnet. If this is the case, then it appears that the assemblage in Figure 4.9b may represent the assemblage present at the invariant point in the experiments of Huckenholz et al. (1976) (listed above), or that the magnetite is being used up in one of the reactions leading to the invariant point. The results of this reaction are also seen in sample TUR I where garnet is found enclosing the perovskite rims around magnetite. Garnet was also reported as rims around perovskite or magnetite in the uncomphagrite rocks at Iron Hill (Nash, 1972). Thus, the experimental reactions examined by Huckenholz et al. (1976) are also seen in natural samples taken from the Turiy melilitolites.

The issue of whether cancrinite was primary or secondary in these samples was discussed by Bell et al. (1996). Given that melilite is generally very sensitive to alteration in many of the melilitolite, the fresh nature of this mineral in the Turiy

Figure 4.9a Melilite-calcite-carbonatite (sample N.64.18a) with “selvage” rims around melilite (at extinction) and calcite. Photograph taken under crossed polars. Field of view = 2 mm

Figure 4.9b Cartoon showing a reaction relationship around magnetite grains in the melilite-calcite-carbonatite N.64.18m. The identification of monticellite is uncertain, since the monticellite(?) band was too small to analyse. Width of the reaction area from the edge of the magnetite to the edge of the melilite is highly variable, but the [andradite + monticellite] bands are 15µm wide.

← 2 mm →

magnetite

andradite

perovskite

monticellite(?)

calcite

melilite

melilitolites (especially TUR I) suggests that it is unlikely that cancrinite is secondary after nepheline. The presence of cancrinite in the turjaite (and in some of the ijolite rocks, as listed in Appendix Biii) is probably the result of the natural reaction between nepheline + calcite, as documented by Watkinson & Wylie (1971), and corroborated by Nielsen (1980) in samples from the Gardiner intrusion.

Panina & Podgornyykh (1974) reported homogenisation temperatures of inclusions containing glass, “ore” and volatiles, in a variety of turjaite minerals including pyroxene (1260°C), melilite (1210-1230°C) and nepheline (1100-1130°C). Secondary inclusions in melilite homogenized at 830-950°C. Homogenisation temperatures for inclusions in the melilites from the Gardiner Complex ranged from 1030 - 1060°C (Nielsen et al., 1997). This strongly suggests that melilite is not a low-temperature, metasomatic mineral, as suggested by Kukhareenko et al. (1965).

The association of diopside and melilite in the turjaite is interesting, given the reaction:

and as seen in TUR I, diopside and calcite are commonly in close proximity to each other.

Studies by Velde & Yoder (1976) showed, however, that the reaction:

“Tschermak”   diopside   alkalis

akermanite   soda-melilite   forsterite   nepheline

could leave excess diopside in association with melilite once the reaction had gone to completion. Given the lack of olivine in the Turiy samples and the presence of calcite associated with the melilite and diopside, it is probable that reaction (i) better characterises samples TUR I and E.133.30.

Garnet in the turjaite normally makes up < 5% of the mode, and this, combined with the lack of pyroxene, and presence of perovskite and magnetite, contrasts with the Fe^{3+} and Ti-bearing pyroxene and garnet, with minor iron oxides in the ijolites. A general comparison of modal abundances of mafic minerals between the melilitolites and ijolite series rocks, according to the modal mineralogy in Appendix Biii, demonstrates:

This comparison of overall modal abundances is interesting in terms of its implications for oxygen fugacity ($f\text{O}_2$), oxide abundances, $a\text{SiO}_2$, and melt structure, in the two rock types. These are topics which will be discussed further in later chapters.

In summary, the Turiy turjaite show great variations in texture and mineralogy, including extensive banding and samples rich in volatile- and trace element-bearing minerals. Cancrinite, where it occurs instead of nepheline, is believed to be primary.

4.6 Phoscorites

Phoscorites are traditionally “magnetite-apatite-forsterite” rocks, rare in occurrence, containing additional calcite and phlogopite (normally $^{iv}\text{Fe}^{3+}$ rich), with less common sulphides, pyroxene and rare trace minerals such as zircon and amphibole. Phoscorites are also found in the Kovdor, Vuoriyarvi, Fen and Oka complexes (Kapustin, 1980; Andersen, 1984; Gold et al., 1986). The history of the term “phoscorite” is summarized by Zaitsev & Bell (1995). The origin of phoscorites is fascinating - the origin of carbonatites can be linked to CO_2 reactions with calcium-rich minerals, and the notion of carbonate-silicate immiscibility is now a commonly accepted phenomenon, but the segregation of large quantities of iron oxide and calcium phosphate is still puzzling. Kapustin (1980) refers to “apatitization” in magnetite-forsterite rocks (phoscorites), associated with the intrusion of carbonatite veins, prior to the main intrusion of

carbonatite bodies, indicating that the apatite in these rocks may be secondary.

All of the phoscorites (“ore rocks”) within Turiy are found in the central carbonatite stock of the central complex. Previous field interpretations (Bulakh and Ivanikov, 1984) suggest that they post-date the late ijolites, but precede carbonatite intrusion. The Turiy phoscorites are dominated by coarse-grained magnetite, veins of fine- to medium-grained apatite, coarse-grained phlogopite, and calcite. Clinopyroxene occurs in significant quantities in some samples. Accessory minerals include amphibole, chlorite, dolomite, strontianite, sulphides, and zircon.

Although phoscorites are traditionally “magnetite-apatite-forsterite” rocks (Egorov, 1993; Zaitsev & Bell, 1995), it can be seen from Appendix Biii that the apatite content is low in the thin sections examined where point counting was employed and forsterite is absent. The modal mineralogy suggests that “magnetite-mica-(pyroxene)-phoscorite” might be a more appropriate name for these samples. However, as they do not contain enough carbonate to be classified as carbonatites, and the hand-specimens commonly contain thick apatite veins, the term phoscorite will be retained here. A notable feature of the phoscorites is their very coarse-grained nature, along with distinctive segregations of clusters of grains of individual minerals, and hence it is difficult to estimate their modal mineralogy. Interpretation of the modal mineralogy tables in Appendix Biii should be carried out with caution.

All phlogopites from the phoscorite samples [C.K.7, C.3.93, C.18.30, C.19.63] contain large amounts of $^{\text{iv}}\text{Fe}^{3+}$, (“tetraferriphlogopite” - terminology not yet approved by the IMA). This imparts a vivid orange colour, and reverse pleochroism (deep orange to pale orange). Concentric zoning and resorption effects within the phlogopites are common. Two out of the four phoscorite samples studied also contain pyroxene, and calcite is a common groundmass component.

Most magnetite-apatite rich rocks at Turiy (i.e. Fe^{3+} -Ti-P-rich phoscorites and some carbonatites) exhibit an interesting textural relationship along the grain-boundaries between magnetite and apatite crystals. Such samples contain euhedral coarse-grained magnetite crystals which exhibit irregular contacts and anhedral intergrowth textures with neighbouring apatite crystals. This unusual texture is specific to contacts between apatite and magnetite, and is presumed to be due to structural or kinetic effects during crystallization. Fe and P are known to form close associations in SiO_2 -poor melts (Mysen, 1992).

4.7 Carbonatites

Turiy is one of the most significant alkaline intrusions on Earth due to the presence of well-preserved, intrusive melilite-bearing rocks, and the presence of pristine carbonatites (Figure 4.10a). Turiy was one of the first places in Russia where carbonatites were first recognized as true magmatic rocks (Brögger, 1921). The majority of the Turiy carbonatites occur in the stock in the middle of the central complex, although two small satellite bodies in the north also contain carbonatites.

Following Le Maitre (1989), carbonatites are defined as igneous rocks with > 50% carbonate minerals. Given the textural heterogeneity of the Turiy carbonatite samples, a little latitude has been allowed in the definition in this study, and some samples with > 40% carbonate minerals have been included within the carbonatite group. One of the other significant textural differences between the phoscorites and carbonatites at Turiy is that the minerals which are common to both rock types tend to be much coarser-grained in the phoscorites, and that textural heterogeneity within the phoscorites is also much more pronounced. In addition, the phlogopite-magnetite-apatite calcite-carbonatites commonly contain dolomite (e.g. C.49.40), which is rarely found in the phoscorites.

An interesting relationship is seen between a sample of central complex ijolite cut by calcite-carbonatite veins showing distinct vertical flow textures (Figure 4.2a). The contacts between the fragmented ijolite and the carbonatite are lined with phlogopite (black bands). Phlogopite also lines the contact zones between ijolites and carbonatites in Oka (Treiman & Essene, 1985). One of the implications of the textures seen in Figure 4.2a is that the ijolites were already crystallized prior to the intrusion of the carbonatite veins, indicating the need for a time gap between the emplacement of the different rock types (this contrasts with the gradational boundaries seen between many of the silicate rock types in this massif).

Textural heterogeneities in the carbonatites include banding, veining, and variations in grain-size both within and between samples. Sample C.6.15 is an excellent example of such textural heterogeneity, with bands of apatite-magnetite bands of different thicknesses and different average grain sizes, separated by bands of calcite and dolomite (Figure 4.10b). One magnetite-apatite band in this sample exhibits baddelyite rims around the apatite grains, a feature that is not seen in other magnetite-apatite veins in the thin-section. S-shaped veins of apatite are enclosed within natrolite in sample C.54.x, a late-stage calcite - natrolite carbonatite vein. Sample C.54.x is the only sample to contain euhedral, rhombohedral calcite, which is also unique in Turiy due to its display of concentric zoning under cathodoluminescence (Figure 4.10c). Such rhombohedral, zoned calcite is common in carbonate-zeolite veins from the Khibina massif (Zaitsev, 1996), commonly believed to be explosive in nature (Bulakh & Ivanikov, 1984). Sample N.187.134 exhibits rare, rounded calcite phenocrysts, with strained extinction, about 3mm in diameter, enclosed in a finer-grained, interlocking, calcite matrix (Figure 4.10d). This sample is a good example of the preservation of primary textures within the carbonatites.

Figure 4.10a Calcite carbonatite from the central massif with clusters of euhedral, inclusion-free apatite crystals and zoned $^{iv}\text{Fe}^{3+}$ -phlogopite enclosed in an interlocking framework of calcite grains. Photograph taken under crossed polars. Field of view = 2mm.

Figure 4.10b Sample C.6.15 contains bands of associated magnetite + apatite, interspersed with bands of calcite and dolomite. This picture shows a coarser band of magnetite + apatite with interstitial calcite (left), a band of anhedral calcite with interlocking grain boundaries (centre), and two bands of fine-grained apatite associated with relatively coarse magnetite, interspersed with calcite. The greatest variation in grain size is seen in the apatite. Baddelyite forms 10 μm thick rims around the apatite crystals just to the left of the central calcite band (not visible). Photograph taken under crossed polars. Field of view = 2mm.

Figure 4.10c Concentrically zoned subhedral calcite crystals in carbonatite C.54.x. Photograph taken using cathodoluminescence microscopy. Field of view = 0.5mm.

Figure 4.10d Sample N.187.100 showing a rounded calcite phenocryst (right) with strained extinction (not visible), in a finer-grained calcite groundmass. The hexagonal mineral on the left is phlogopite (near-perfect optic axis orientation, hence low interference colour). Photograph taken under crossed polars. Field of view = 2mm.

The northern carbonatites [N.19.63, N.63.4, N.64.18, N.187.100, N.187.134] commonly have low apatite and magnetite abundances and contain no amphibole, but several of these samples contain significant quantities of garnet, phlogopite and pyroxene. Several of the northern carbonatites which lack pyroxene are melilite-bearing, and one contains significant monticellite. Monticellite is usually considered to be a high-temperature mineral, but can occur in equilibrium with calcite at temperatures as low as 600°C (Frantz & Wyllie, 1967). Kapustin (1980) described monticellite-melilite-calcite rocks as being formed by “replacement” without decomposition of melilite, where carbonatites intrude into melilitolites. Such textures and mineral assemblages were found in Oka (Maurice, 1956), but were described as magmatic assemblages by Treiman & Essene (1985).

The central carbonatites tend to have a more varied mineralogy than the northern carbonatites, and include amphibole-rich varieties, dolomite-calcite-carbonatites, and contain variable amounts of sulphide. The abundance of sulphide is inversely proportional to the abundance of silicate minerals and roughly proportional to the abundance of dolomite. No melilite or garnet are found in any of the central massif carbonatites, except C.54.x, a late-stage vein in the north of the central massif, which contains garnet. Apatite, magnetite and $^{iv}\text{Fe}^{3+}$ -phlogopite are abundant, with some of the central massif carbonatites almost grading into phoscorites.

Most of the carbonatites in Turiy are calcite-carbonatites, with occasional dolomite, and rare blebs of strontianite. Dolomite is more abundant in the younger carbonatites (Figure 4.11), and the calcite-dolomite-carbonatite [C.DC], containing equal amounts of calcite and dolomite, is the most dolomite-rich carbonatite examined in this study. Ankerite is only found in the two late-stage vein samples, TD 1 and TD 2. Thus, a general sequence of carbonatite evolution can be seen in the gradation from melilite-calcite-carbonatites and pyroxene-calcite-carbonatites, to amphibole calcite-carbonatites,

Figure 4.11 Cathodoluminescence picture of a calcite (orange) carbonatite, veined by zoned dolomite (red) and concentrically zoned apatite (blue/purple) in sample C.49.40. The apatite appears to be concentrated only on one side of each dolomite band. Field of view = 0.5mm.

Figure 4.12 Concentrically zoned phlogopite from sample C.49.40. The phlogopite alternates between reverse pleochroic $^{iv}\text{Fe}^{3+}$ -phlogopite (bright orange in the central crystal) and normal phlogopite (pale orange in the central crystal). The groundmass in this photograph consists mostly of fluid-inclusion-rich apatite. Photograph taken in plane-polarized light. Field of view = 2mm.

to apatite-phlogopite-magnetite-calcite-carbonatites, and to dolomite-calcite-carbonatites. It is presumed that the ankerite-fluorite-quartz and ankerite-aegirine apatite veins (TD 1 and TD 2 respectively) seen on the south coast represent the latest phase of all. This pattern follows the carbonatite classification scheme suggested by Pozharitzkaya & Epshtein (1963, 1964), who divided carbonatites into four stages of intrusion: early and late calcite-carbonatites, dolomite-calcite-carbonatites and ankerite-carbonatites. More than 90% of all carbonatites fall into the first two stages (Kapustin, 1980), and, because of this, Kapustin (1980) divided carbonatites into two groups “early” and “late” and claimed that dolomite preceded calcite formation in all massifs. However, where dolomite appeared after calcite, it was claimed to be secondary, replacing calcite. The carbonatite samples seen in Turiy almost certainly follow the scheme of Pozharitzkaya & Epshtein (1963, 1964) rather than that of Kapustin (1980), as will be discussed further in later chapters.

Trace minerals in the carbonatites include amphibole, normally richterite or tremolite. The former occur as radiating clusters of elongate crystals, the latter as clusters or solitary crystals in the later carbonatites. Baddelyite, barite, calzirtite, chlorite (alteration), garnet, ilmenite (micro-intergrowth features in magnetite), quartz, perovskite, pyrochlore, sulphides, titanite, zeolites, zircon and zirsinalite have also been found in a few samples. Diopside (up to 50% modal abundance) occurs in several of the northern carbonatites, and makes up to 48% of the rock. Kapustin (1980) considers diopside to be typical in early carbonatites. His description of aegirine as typical of later carbonatites is not borne out in the central complex samples. The closest representative is an aegirine-augite-apatite-ankerite vein on the south coast [TD 2].

The phlogopites in the carbonatites show an interesting range of textures and compositions. Kapustin (1980) identified four generations of mica growth within carbonatites. The first and third generations are represented in Turiy. The first

generation is represented by the pale green phlogopites in the northern complex carbonatites, while the third generation is very clearly represented by the $^{iv}\text{Fe}^{3+}$ -phlogopites which occur in almost all the carbonatites within the central complex, especially the apatite-magnetite-phlogopite calcite-carbonatites. Phlogopite in sample C.49.40 exhibits “picture-frame” concentric zoning involving repeated alternation between normal and reverse absorption schemes and orange pleochroism (Figure 4.12). Sample C.49.40 also contains small veins of dolomite, reminiscent of the dolomite veins and “micatization” documented by Frolov (1969) and Kapustin (1980) in the cataclastic carbonatite zones of Vuoriyarvi. Such late-stage metasomatism is also thought to have affected some of the Turiy carbonatites (Bulakh & Ivanikov, 1984). However, it should be noted that there is little evidence (as seen in cathodoluminescence photographs) for replacement textures in most carbonatites seen at Turiy. A large proportion of $^{iv}\text{Fe}^{3+}$ -phlogopite in the carbonatites, commonly zoned, coupled with a general lack of alteration of other minerals and lack of associated replacement features, suggests that much of the phlogopite is primary. Nash (1972) reported the development of $^{iv}\text{Fe}^{3+}$ -phlogopite rims around cores of Mg-rich “normal” phlogopite in the rauhaugites at Iron Hill, a characteristic also found in the calcite-dolomite-carbonatites at Turiy.

“Richterisation” is believed to have affected some of the Turiy carbonatites (Bulakh & Ivanikov, 1984). Richterite forms radiating clusters of needles in a few carbonatite samples at Turiy, and these may indeed be a late-stage feature.

Ti-poor magnetite is ubiquitous in the carbonatites, although its abundance is very varied. It is reported to have an inverse relationship with biotite (Kapustin, 1980), as seen in the Turiy melilolites (Figure 4.5). However, in the Turiy carbonatites, magnetite is very abundant in rocks containing Mg-rich phlogopite (and $^{iv}\text{Fe}^{3+}$ -phlogopite), and the inverse magnetite-mica relationship does not appear to occur. At Turiy, magnetite is most commonly found in the carbonatites that have greatest similarity in modal

mineralogy to the phoscorites, and in some of the melilite-bearing carbonatites from the northern complex. Sample N.64.18 represents a contact between a pure calcite-carbonatite and a melilite-calcite-carbonatite. The boundary between the two rocks is delineated by a continuous band of magnetite which is also a common component of the melilite-calcite-carbonatite. In general, magnetite is less abundant in the carbonatites than in the phoscorites.

Apatite is typically an abundant phase in the Turiy carbonatites, and can reach up to 30%. A large variety of textures are associated with the apatite crystals. Textural equilibrium is in evidence in many of the Turiy carbonatites between calcite and apatite (an important consideration when interpreting isotope data from mineral separates from these samples), although the modal abundances may be heterogeneous. Apatite can occur as solitary, rounded grains associated with rounded, fine- to coarse grained calcite. Some samples contain apatite grains at the calcite grain corners (Figure 4.13a), typical of early carbonatites (Kapustin, 1980). Apatite also occurs as veins and/or clusters of tightly-packed grains with hexagonal cross-sections (Figure 4.13 b&c), and a variety of zoning patterns, textures and inclusions, in terms of their sizes, types and abundances. Those apatite crystals which contain large numbers of very small, tightly clustered inclusions, mostly contain volatile and fluid-rich inclusions. Those crystals which contain few, but larger inclusions, commonly contain carbonate (Figure 4.13a) or occasionally zircon inclusions. Some of the apatite crystals contain elongate inclusions, which parallel the *c* crystallographic axis of the crystals (Figure 4.13d), commonly emphasizing the random orientation of neighbouring grains. Figure 4.13d shows one elongate apatite in the centre of the picture with two inclusion-rich areas which may represent two phenocrystic cores linked together by a later overgrowth. It is likely that at least the cores of the apatite crystals which contain multiple inclusions crystallized prior to emplacement, and that the rims of the crystals were formed *in situ* during the final stage of magma emplacement and

Figure 4.13a Cathodoluminescence picture of apatite (blue) showing very mottled zonation throughout the crystal. The cause of this zonation is not known. The apatite grain is surrounded by calcite crystals (orange) with rare interstitial blebs of dolomite (red). The apatite crystal contains a small train of fluid or calcite inclusions (left to right) and larger solitary inclusions, mostly filled with calcite. It can be seen that the two large calcite inclusions near the top of the apatite crystal occupy only part of the inclusion space in which they sit, indicating that the calcite is coexisting with a fluid or other volatile phase. Field of view < 0.5mm.

Figure 4.13b Cathodoluminescence picture of a cluster of euhedral, concentrically-zoned apatite crystals, surrounded by silicate minerals (garnet and pyroxene, black) and calcite (orange). Field of view 1.5 mm.

Figure 4.13c Close-up of the concentric zonation in the apatite crystals in Figure 4.10b, exhibiting at least 8 stages of growth. Conentric zonation can also be seen in the calcite crystals adjacent to the apatite. Photograph taken using cathodoluminescence microscopy. Field of view < 0.5mm

Figure 4.13d Apatite crystals showing large numbers of small fluid(?) inclusions within the cores of the crystals that are overgrown by inclusion-free rims. A few large, solitary inclusions can also be seen. The central apatite crystal appears to contain two cores which have been joined together by a later overgrowth. Surrounding material includes calcite and phlogopite. Photograph taken in plane polarized light. Field of view = 0.5 mm.

crystallization. Samson et al. (1995) used this two-stage crystallization model in an attempt to explain overpressured inclusions in apatite, monticellite and calcite within carbonatites from the Oka complex. The Oka inclusions crystallized at 4-8 kbar and 675-720°C, presumably indicating the carbonatite crystallization conditions, but the Oka complex is known to have been emplaced at a confining pressure of < 1 kbar, supporting the idea of early crystallization of the inclusion-rich apatite cores. In Turiy, rounded apatite crystals which occur at calcite grain corners rarely contain inclusions, and those inclusions which are found are commonly large, solitary calcite inclusions (Figure 4.13a). These calcite-apatite textures may suggest that the two minerals crystallized concomitantly and in equilibrium with each other in these samples.

Calcite crystals in the carbonatites tend to exhibit four main textural styles:

- i) medium-coarse grained, rhombohedral crystals with extensive concentric zoning (Figure 4.10c). With rare, and poorly defined exceptions, these latter two features are only seen in C.54.x.
- ii) rounded crystals, fine-coarse grained, occasionally with strained extinction. Some samples contain apatite crystals at the calcite grain corners. Many of the early central carbonatites, and some of the northern melilite-bearing carbonatites exhibit these features. Similar features seen within the Kovdor carbonatites were assumed to be indicative of primary textures (Zaitsev & Bell, 1995).
- iii) fine-medium grained, uniform extinction, and irregular interlocking grains. This texture is most commonly seen in areas of a thin-section containing 100% calcite, or with small crystals of apatite located at the calcite grain corners (Figure 4.10a), which may indicate that the rock crystallized in chemical and physical equilibrium.
- iv) a very mottled texture within anhedral and interstitial carbonates, representing a mixture of calcite, dolomite and ankerite. This texture is probably metasomatic in origin, and has only been seen so far in two samples which are not carbonatites *per se* - one a

calcite-fluorite vein [TD 1] and the other an aegerine-augite-apatite-ankerite rock [TD 2] (Figure 4.15 a&b).

The origin of these is uncertain, but some constraints can be suggested. The first, occurring within a hydrothermal vein [C.54.x] is thought to indicate rapid cooling under rapidly changing fluid conditions. This vein is also the only central carbonatite which has mineralogical composition links with the melilite-bearing rocks (as described in Chapter 5 and discussed in Chapter 8).

The second calcite texture described above is indicative of a carbonate-phosphate melt cooling in equilibrium (as opposed to apatite veins cutting through carbonate-rich sequences, as seen in the phoscorites), and the third likely represents cooling of a solitary carbonate phase. The fourth, as already stated, is probably metasomatic in origin.

4.8 Turjite

The coastal exposures around the Turiy Peninsula include turjite, the second type-locality rock found in this area. It was first defined by Beliankin & Kupletskii (1924) and described by Le Maitre (1989) as a “melanocratic lamprophyre rich in biotite, analcime, calcite and melanite garnet”. Bell et al. (1996) gave an extensive description of two type-locality samples [TU 119, TU120A]. This rock-type is a peculiar hybrid between the melilitolites, ijolites and carbonatites. It occurs as two dykes on the southern coast of the massif, and may be part of the syn-massif or post-massif intrusion suite of dykes in this area (Bulakh & Ivanikov, 1984).

Three varied examples of turjite were examined in this study [TU 119, TU 120A, TQ]. The medium- to coarse-grained sample TU 119 bears resemblance to sample C.IJ.14, a phlogopite garnet urtite, described earlier. Sample TU 119, however, contains more garnet and mica than sample C.IJ.14, and analcime dominates the groundmass. The replacement of pyroxene by phlogopite, seen in sample C.IJ.14 appears to have gone to

completion in sample TU 119. Replacement of pyroxene by mica is also considered common in the centre of many carbonatite bodies (Kapustin, 1980). Sample TU 119 is striking in thin-section, containing more than 50% anhedral- to euhedral zoned medium- to coarse-grained garnet (Figure 4.14a) and coarse-grained, bright orange, zoned, reverse-pleochroic $^{iv}\text{Fe}^{3+}$ -phlogopite (Figure 4.14b). The garnet ranges in composition from andradite (pale yellow/green, weakly anisotropic) to schorlomite (deep brown, almost opaque and isotropic). Euhedral, concentrically zoned crystals alternate in composition between andradite and melanite (Ti-rich andradite), whilst schorlomite forms irregular, anhedral veins. One fine-grained, euhedral andradite crystal, completely enclosed by interstitial analcime, appears anisotropic (1^{st} order grey interference colours) and exhibits multiple concentric and sector growth zones under crossed polars. The anisotropy in andradite is believed to be due to a small hibschite or katoite solid solution component within the andradite. At least five major zones of growth can be identified in the euhedral garnet crystals in sample TU 119. The phlogopites show mostly core-rim variation, with the proportion of $^{iv}\text{Fe}^{3+}$ -phlogopite increasing in the rims. In one location, a small patch of annite was found as a rim on one of the phlogopite crystals. A few patches of interstitial calcite make up the rest of the modal mineralogy, along with trace quantities of fine-grained apatite, magnetite, perovskite and strontianite. One small aegirine crystal was found enclosed within phlogopite.

Sample TU 120A is more fine-grained and equigranular than sample TU 119, with a garnet : phlogopite ratio of $\sim 1:1$, as opposed to 5:1 in sample TU 119, and it is cut by a small, porphyritic nephelinite vein (< 0.5 cm). Turjite sample TU 120A is also altered, and evidence for hydrothermal activity includes the presence of Ba-Mn rich minerals such as euhedral celsian and rare Mn-rich ilmenite in the groundmass. Calcite is the main groundmass mineral in sample TU 120A, as opposed to the analcime seen in sample TU 119. The pleochroic scheme of the fine- to medium-grained zoned phlogopite

Figure 4.14a Turjite sample TU 119 showing euhedral and concentrically zoned (pale yellow - dark brown) andradite - melanite garnet. There is also an anhedral area of schorlomite in the top left-hand corner of the picture. Photograph taken in plane polarized light. Field of view = 2mm.

Figure 4.14b Turjite sample TU 119 showing three grains of zoned phlogopite, with variations in pleochroic colours dependent on crystal orientation. Groundmass consists of garnet and analcite. Photograph taken in plane polarized light. Field of view = 2mm.

in sample TU 120A varies in pleochroism between pale green to pale orange, and pale orange to bright orange due to alternation between phlogopite and ${}^{\text{iv}}\text{Fe}^{3+}$ -phlogopite respectively. Phlogopite is also subophitically enclosed by garnet. The fine- to medium-grained garnet in this sample is concentrically zoned between andradite and melanite, as seen in sample TU 119, and is commonly anisotropic. The granularity of the garnet crystals depends on the neighbouring minerals - the garnet crystal faces next to calcite are euhedral, those which sub-ophitically enclose phlogopite, or are adjacent to groundmass alteration products, are anhedral. The textural relationship between garnet and phlogopite suggests that both minerals crystallized at the same time, with garnet continuing to crystallize with calcite after the cessation of phlogopite formation.

The final turjite sample examined [TQ] comes from the edge of the main turjite dyke, and represents the chilled margin. This sample is highly altered (> 30%), and shows abundant relict laths (37%), believed to have been melilite (or alternatively, calcite), cross-cutting the sample. Regardless of composition, the manner in which the laths dominate the texture of this sample suggests that they are a product of rapid crystallization. The remainder of the minerals consist of equigranular and texturally homogeneous fine-grained garnet, bright orange ${}^{\text{iv}}\text{Fe}^{3+}$ -phlogopite, and interstitial calcite.

The petrogenesis of the turjite is hard to ascertain. Turji is the type-locality for this rock which is unknown elsewhere on Earth. Bell et al. (1996) proposed that turjite is the product of a reaction between carbonatite and ijolite, using evidence from the experimental work of Verwoerd (1978) which showed that andradite garnet is the initial crystallizing phase on the liquidus. Sample C.IJ.14 would be a suitable ijolite end-member, or may represent the initial stages of the reaction. Sample C.IJ.14 shows significant replacement of pyroxene by ${}^{\text{iv}}\text{Fe}^{3+}$ -phlogopite (a reaction which is virtually complete in TU 119), and there is an increase in the abundance of garnet from sample C.IJ.14 to the turjite. The model of Verwoerd (1978) calls for a natrocarbonatite-calcite-

carbonatite mixture for the “carbonatite” end-member in the reaction. The nature of the carbonatite involved in producing the turjite cannot be constrained on petrography alone. However, given the presence of calcite in the groundmass of all samples, it is likely that calcite-carbonatite was the dominant carbonatite in the reaction.

4.9 Late-stage veins

The southern coastal exposures around the Turiy Peninsula are predominantly made up of gently fenitized Riphean sandstones, alkaline dykes (nephelinite, melilitite, turjite, carbonatite) and late-stage veins, believed to have originated from fluids emanating from the main massif. These late-stage, or possibly metasomatic, veins have different mineralogies. Included among these are pectolite-wollastonite, calcite-quartz-fluorite and aegirine-apatite veins. Samples TD 1 and TD 2 are representative of the calcite-quartz-fluorite and aegirine-augite-apatite-ankerite veins respectively.

Sample TD 1 is coarse- to fine-grained and contains 30% purple fluorite with calcite and ankerite (35% and 20% respectively), making this, in essence, a carbonatite. However, in other parts of the vein, fluorite constitutes > 50% of the mode, where carbonatite becomes inappropriate terminology. Quartz and apatite are also present, along with trace quantities of sulphides. Nb-rich rutile inclusions are contained within the quartz. Cathodoluminescence studies show concentric zonation in the fluorite (Figure 4.15a), especially in the rims, and mottled calcite and ankerite in the carbonate groundmass. The apatites are fine-grained, inclusion free and euhedral, and occur throughout the section, commonly enclosed by carbonate or fluorite. They occur, less commonly, within patches of quartz. The centres of the fluorite crystals exhibit irregular zonation and contain inclusions of carbonate and apatite.

Sample TD 2 was collected from an aegirine-augite - apatite - ankerite vein given the pseudo-mnemonic name “aegiapite” by Beliankin and Vlodavets (1932) to

Figure 4.15a Sample TD 1 exhibiting concentric zonation in fluorite crystals. Small, euhedral blue/purple crystals also showing zonation are apatite. Groundmass contains calcite and ankerite. Photograph taken using cathodoluminescence microscopy. Field of view < 2mm

Figure 4.15b Sample TD 2 exhibiting two varieties of apatite. In the left of the photograph, a large crystal with irregular zoning, a variety of small inclusions, and a very regular overgrowth rim can be seen, representing the first variety of apatite. Nearby, are a small number of euhedral, zoned apatite crystals representing the second variety of apatite, within a groundmass of aegirine (black), calcite (orange) and ankerite (deep orange). Most of the small groundmass crystals are oriented so that the c-axis is perpendicular to the plane of the photograph. In other parts of the section, these groundmass crystals are seen to be acicular and to radiate out from the edges of the coarse apatite crystals. Field of view 1.5mm.

characterise the dominant aegirine-augite - apatite mineralogy. Pyroxene and apatite account for 70% of the modal mineralogy, in roughly equal proportions, with carbonates accounting for 20% and K-feldspar the remaining 10%. Aegirine in this sample is deep green, faintly pleochroic (to paler green), subhedral- to euhedral, and mostly medium-grained. Two generations of apatite can be distinguished - the first generation is medium-grained with abundant inclusions in the cores of the crystals, and euhedral, inclusion-free rims, producing extremely well-defined crystal edges and apices. The second generation of apatite consists of fine-grained, inclusion free, acicular crystals that commonly grow perpendicular to the crystal faces of the first generation of apatite, or occur within the groundmass (Figure 4.15b). Feldspar and carbonate minerals make up the remainder of the groundmass. The mottled calcite-ankerite textures are similar to those of TD 1, as seen using cathodoluminescence techniques. Trace amounts of Nb-rich rutile were also found in sample TD 2, along with quantities of strontianite, barite and mica similar to those found in sample TD1.

The origin of these veins, as with all late-stage, volatile-rich material, is heavily dependent on fluid and volatile composition. Neither sample contains any minerals with significant quantities of H₂O or Cl. F-bearing phases and a variety of carbonate minerals are present in both samples. Their modal mineralogies reflect variations in their major-element chemistry. Sample TD 1 dominated by alkaline earth elements, Fe²⁺ and Si, while sample TD 2 contains a significant proportion of alkali metals and the network forming cations Fe³⁺, P, and Si. The first generation of carbonatites in the Khibina Massif are biotite-aegirine-apatite-carbonate rocks with carbonate < 50 modal %, similar in composition to sample TD 2, which are believed to be petrogenetically associated with the phoscorites (Zaitsev, 1996).

4.10 Fenites

Two samples of contact ijolite-fenite and one sample taken several hundred metres from the massif were examined in an attempt to study the composition of the country rock into which the massif intruded, and the effects of fenitization due to fluids emanating from the massif.

The Turiy massif intruded into Proterozoic charnockite rocks and granitoids (Ronenson et al., 1978; Bulakh & Ivanikov, 1984; Kogarko, 1987; Arzamastsev, 1994). Sample NC.153.82 is a fenitized version of these country rocks taken from a location to the north of the central complex. Consisting primarily of potassium feldspar and pure albite with sodic pyroxene and alkali amphibole, the rock could be classified as a syenite (Le Maitre et al., 1989). It is similar in petrography and composition to many syenites, especially those of the Chilwa Island complex of Malawi (Woolley & Jones, 1987). The rock contains 10% alteration products. Trace quantities of apatite, barite, calcite, fluorite, sulphides, titanite, wollastonite, zircon and an unknown mineral were also found. Perthite exsolution lamellae in the coarse-grained, anhedral, alkali feldspar commonly contain wollastonite (or occasionally calcite), which is only visible using cathodoluminescence (yellow in Figure 4.16). This combined mineralogical and textural association has not been described before. Deep green aegirine-augite and less common, very dark green, pleochroic amphibole (ferroan pargasitic hornblende) account for the mafic phases. In Vuoriyarvi, the gneisses surrounding the massif were “converted into massifs of syenitic composition” according to Kapustin (1980).

Possible variations in the degree of fenitization surrounding different parts of the massif, cannot be assessed using the limited sample set currently available to the author from Turiy. However, it is likely that halogen-rich fluids played a significant role in the development of the fenites. Fluorite is found, along with Cl-rich amphibole, in sample NC.153.82.

Figure 4.16 Sample NC.153.82 (fenite). In the top right-hand corner the perthite lamellae within an alkali feldspar are filled with wollastonite (yellow). Other minerals present include pyroxene (black, replaced by calcite). Photograph taken using cathodoluminescence microscopy. Field of view < 2mm

Samples C.190.x and “ij-fen” were collected from ijolite-fenite contact zones, the former from the NE side of the Eastern stock, and the latter from an unidentified drill-core box on the main Turiy drill-core site. The ijolite part of both samples closely resembles sample E.123.106, the eastern ijolite sample described earlier, and it consists of medium-grained, subhedral diopside and interstitial nepheline. The contact between the ijolite and fenite in both samples is marked by the development of fine-grained aegirine and feldspar in the fenite part of the sample, and medium-grained diopside crystals with aegirine rims in the ijolite side which partly intrude into the fine-grained contact.

The fenite from both samples contains potassium feldspar, albite, aegirine-augite/hedenbergite and sodalite as the primary minerals. Sample C.190.x includes 20% unidentified alteration products within the fenite. Barite, strontianite, tausonite (Sr-perovskite), titanite and wollastonite occur in trace amounts. Close to the contact, medium- to coarse-grained feldspar and euhedral aegirine with interstitial sodalite are dominant. About 2 cm from the contact, skeletal acicular crystals of aegirine are preferentially oriented perpendicular to the contact between the two rock types.

Unfortunately, it was not possible to obtain a sample of the country rock from a location unaffected by fenitization. It is possible, although unlikely, that the two contact samples might represent late-stage syenites rather than fenites. The mineralogies of the supposedly fenitized country rock and the syenites are similar. The lack of nepheline in any of the samples, however, indicates that these samples do not represent late-stage nepheline syenites, such as those found in the melilitolite-carbonatite Kovdor Massif (Arzamastsev, 1994).

CHAPTER 5

MINERALOGY

5.1 Introduction

The aim of this chapter is to outline, describe and interpret the compositions of the minerals found in the Turiy samples. Of particular interest are the issues of solid-solution variability within each mineral, whether the compositions of each mineral vary systematically with rock type, and how such variations can help to interpret the petrogenesis of these rocks.

Because of the textural heterogeneity of many of the samples, it is difficult to obtain any meaningful data on element partitioning between any one mineral grain and its host magma. The only clue to factors governing changes in magma composition lies in comparisons between varying compositions of minerals common to a variety of rock types, although it must be borne in mind that changes in pressure and temperature conditions will also play a role in magma chemistry and mineral-melt partitioning.

The electron microprobe at the Canadian Museum of Nature, along with the SEM in the Department of Earth Sciences at Carleton University, were used to obtain the mineral analyses. Analytical techniques, including procedures for the calculation of standard deviation values for both weight % oxide and cation values, and bulk data reduction methods, for the electron-microprobe and SEM analyses are described in Appendix D. The precision of the weight % oxides, based on standard deviation values, are a linear function of the weight % values, and have a range of ± 0.02 - 0.17 wt.%. The (instrumental) standard deviation values for each cation value is governed by a complex relationship between all the elements within an individual analysis. Calculated values for the elements present in a garnet analysis from carbonatite sample N.187.100, provided as an example in Appendix D, have a range of ± 0.011 - 0.030 apfu (atoms per formula unit). Mineral formulae for all the minerals discussed within this chapter are listed in Appendix

I, and selected microprobe analyses are tabulated in Appendix J. Textural and petrographical relationships within individual samples are described and discussed in Chapter 4. Mg# is defined as $100[\text{Mg}/(\text{Mg}+\text{Fe}^{2+})]$ throughout this chapter, unless otherwise stated. Calculated Fe^{3+} (where applicable), is not included in the Fe component used in the Mg# definition, and Fe is assumed to be present as Fe^{2+} at all other times.

The major rock-forming minerals are listed below in order of increasing complexity of silicate structure, followed by the major non-silicate groups. An additional section lists hydrous alkali-silicates and uncommon minerals.

5.2 Silicate minerals

5.2a Monomeric silicates

5.2ai *Olivine & Monticellite Groups*

Olivine has only been found in the olivine melteigite sample C.AG.48A. The Mg# of the olivine varies from 81.7 - 85.4. The CaO content remains constant around 0.5 wt.% while the MnO concentration varies from 0.22-0.42 wt.%. No significant core-to-rim compositional variations are seen.

The origin of these coarse-grained olivine crystals is uncertain. One possibility is that they represent mantle xenocrysts, but they are too abundant and their Mg#s are relatively low for xenocrystic mantle olivine (Dunworth & Wilson, submitted). Alternatively, they may represent an accumulation of phenocrysts within the magma. Their rounded nature suggests that they have undergone some resorption *en route* to the surface. Turiy is reported to contain olivinite relics within the pyroxenites and rare olivine in the melilitolites (Ronenson et al., 1978), but none were found in this study of the complex. The olivinite from the Afrikanda Massif to the northwest of Turiy, contains

olivines with approximately constant Mg# $\sim$ 86 (this study), slightly higher than the values obtained from the olivine melteigite.

It is unusual not to find forsteritic olivine in carbonatites, and particularly phoscorites (Kapustin, 1980), but no olivine has been found in either the phoscorite or carbonatite samples investigated in this study. Monticellite occurs in two carbonatites from the northern satellite body, both of which are melilite-bearing. In sample N.64.18m, a melilite-calcite-carbonatite adjacent to a calcite-carbonatite, monticellite forms part of a suite of minerals that includes diopside, vesuvianite, and a variety of hydrous alkali-silicate minerals which have replaced melilite. The Mg# of the monticellite falls between 67.4-78.1, indicating a significant kirschenite (CaFeSiO_4) component. MnO contents vary from 1.3-1.8 wt.%. Medium-grained, anhedral, interstitial monticellite also occurs in a melilite-calcite carbonatite N.64.19 in association with garnet and phlogopite. Such mineral associations are typical of early carbonatites (Kapustin, 1980). The monticellite in this sample has an anhedral, interstitial habit, and its birefringence and 2V are very similar to those of olivine (Nesse, 1991). The Mg# of the monticellite is constant at 70, with no zonation apparent. The significant Fe content (30% kirschenite component) has had no apparent effect on the birefringence or 2V, in comparison with the optical properties of pure monticellite (Nesse, 1991). MnO contents in monticellite from sample N.64.19 range from 2.11-2.25 wt.%, and $\sim$ 0.15 wt.% SrO is also found. One small grain of monticellite was found along a grain boundary in the melilitolite sample TUR II with a Mg# of 74.

It is fairly unusual (Deer et al., 1992) to find such low Mg#s in natural monticellite from igneous rocks. The low Mg#s seen here may reflect late-stage magmatic differentiation, particularly in sample N.64.19, where the monticellite is interstitial to, and appears to have crystallized after neighbouring minerals with higher Mg#s. Monticellite is also reported from melilite-bearing rocks in the Gardiner intrusion,

Greenland (Nielsen, 1980), Kovdor (Arzamastsev, 1994) and Oka (Gold et al., 1986), and as rims around olivine in kimberlites (Kampata et al., 1994).

5.2a *Garnet Group*

The garnet group is represented in Turiy by the $\text{Ca}_3(\text{Fe}^{3+}, \text{Ti}^{4+}, \text{Al})_2(\text{Si}, \text{Al}, \text{Fe}^{3+})_3\text{O}_{12}$ solid solution (titangrandite) series with end-members andradite ($\text{Ca}_3\text{Fe}^{3+}_2\text{Si}_3\text{O}_{12}$), schorlomite ($\text{Ca}_3[\text{Ti}, \text{Fe}]_2[\text{Si}, \text{Fe}]_3\text{O}_{12}$) and grossular ($\text{Ca}_3\text{Al}_2\text{Si}_3\text{O}_{12}$). It is possible that some of the Si-rich melilitolite and turjite garnet analyses also contain a small amount of the hibschite or katoite component ($\text{Ca}_3\text{Al}_2[(\text{SiO}_4), (\text{OH})_4]_3$). The presence of such a component causes anisotropy in garnet, which is seen in the andradite-rich garnet from the melilitolites and some of the turjite samples. Garnet occurs in almost all rock types in the complex, with varying degrees of abundance, and also occurs as monomineralic veins within some of the ijolite series rocks. Garnet is one of the minerals that shows significant variations in composition with rock type, and this factor, combined with its widespread occurrence and abundance, has made it a critical mineral in the interpretation of the chemistry of the rocks at Turiy.

The formula $\text{A}_3\text{B}_2\text{T}_3\text{O}_{12}$ allows for a variety of (commonly coupled) substitutions in all cation sites, although substitutions in the Turiy garnets appear to be mostly limited to the B and T sites. Much work has been done over the last 20 years into the nature of cation substitutions in the tetrahedral site of minerals such as garnet (Locock et al., 1995) and mica (Abrecht & Hewitt, 1988; Foley, 1989, 1990; Zhang et al., 1993). Possible tetrahedral cations include Si, Al, Fe^{3+} , Ti^{4+} , Ti^{3+} , Fe^{2+} , Mg. That Si is the dominant occupant of the tetrahedral site is not in doubt, but the role of Ti has caused some debate, particularly in the light of its variable valency. Mn substitution is also a problem in this respect, but its abundance in these samples is not significant enough to warrant inclusion in this discussion of multiple valencies.

The redox equation:

provides a means of stabilizing the reduced Ti^{3+} ion, under apparently oxidising (Fe^{3+}) conditions, using IVCT (inter-valence-charge-transfer) (Sherman, 1987; Burns, 1993). The IVCT stabilization energies vary depending on the octahedral (M-site) or tetrahedral (T-site) co-ordination of the relevant cations. Experimental studies have concluded that despite the small size of the Ti^{4+} ion, it does not enter the T-site in mica or garnet (Abrecht & Hewitt, 1988; Locock et al., 1995). Al and Fe^{3+} are able to substitute for both T and B ions. It is unlikely that Mg is a viable candidate for tetrahedral co-ordination in garnet, given its large size, although Foley (1989, 1990) claimed to have found Mg in tetrahedral co-ordination in lamproitic micas. One of the latest spectroscopic studies on garnet by Locock et al. (1995) proposed tetrahedral Fe^{2+} preference after Si, ahead of Al and Fe^{3+} . While recognising that work in this area is currently in a state of flux, and that co-ordination preferences can vary depending on magma chemistry and oxygen fugacity, a standard formula was used for garnet, mica and pyroxene in this study with $\text{Si} > \text{Al} > \text{Fe}^{3+}$ preferences for occupancy of the tetrahedral site and exclusion of Ti^{4+} , Fe^{2+} and Mg from tetrahedral co-ordination. Garnet formulae were calculated with 16 cations : 24 anions, while realising that the $\text{Fe}^{2+} \rightarrow \text{Fe}^{3+}$ conversion enabled by this valence calculation makes no allowance for the varying valency of Ti or Mn, for any site vacancies, or for any hydrogrossular component.

In all analyses of garnet from all rock types in Turiy, Ca dominates the distorted cubic A site, never dropping below 96.5% occupancy. In the sample with the lowest Ca content (eastern ijolite, E.123.106), Na is uncommonly abundant ($\text{Na}_2\text{O} < 0.6$ wt.%). Thus, substitutions in the A-site are minimal and include Na and occasionally Sr, along with unspecified members of the (Mg, Fe^{2+} , Mn) group where appropriate.

Following the guidelines outlined above, the incorporation of Ti^{4+} into the M site causes the need for coupled substitutions such as:

It should be emphasized that an inverse relationship between Ti and Si concentrations is in agreement with the above substitution scheme, and does not imply direct Ti-Si substitution in the tetrahedral site. A possible M-site coupled substitution, which does not relate to ${}^{\text{iv}}\text{Si}$ content is:

A plot of ${}^{\text{vi}}\text{Ti}$ vs ${}^{\text{iv}}\text{Si}$ for this scheme would show no correlation. Samples rich in Ti appear to follow scheme (ii) until they reach $\text{Ti} = 1/2$ M-site occupancy (schorlormite, ~15 wt.% TiO_2), followed by substitution dominated by scheme (iii).

The abundance of garnet is greatest in the ijolite-urtite series, where garnet and phlogopite increase in abundance relative to pyroxene as the proportion of nepheline increases. Garnet in these rocks is rich in Fe^{3+} and Ti ($\pm$ Zr) and commonly grades into schorlomite-rich compositions (Al-poor), with Fe^{3+} present in both tetrahedral and octahedral co-ordination. Concentric zonation is common between andradite (pale green) and Ti-rich (dark brown) bands.

The melilite-bearing rocks contain rare Al-rich andradite (Al-poor grossular) that is common, but not abundant, in these samples. Al-rich andradite probably occurs as a late-stage reaction product of melilite with residual magmatic fluids or due to melilite break-down (emphasized by the lack of Ti in the garnet as Ti is also absent from melilite), and forms veins and rims along grain-boundaries. Only in melilitolites such as sample TUR I, a pegmatitic variety of melilitolite, does garnet form fine-medium-grained brown anhedral crystals, relatively rich in Ti, which may grade into colourless veins of andradite around melilite, or magnetite and perovskite. (Bell et al., 1996). Grossular usually occurs in metamorphosed calcareous rocks, or rocks subject to calcic

metasomatism (Deer et al., 1992), and thus the presence of $^{\text{vi}}\text{Al}$ -rich garnet in the melilitolites is consistent with the calcium-rich nature of these samples.

There are two points of note here. The first point is that garnet can occur as late-stage veins in igneous rocks (also documented from the Oka complex by Watkinson, 1971), contrary to the normal impression of garnet as a dominant granular silicate component of mantle xenoliths or eclogite. The second point is the variation in $(\text{Ti}, \text{Fe}^{3+})$ - Al abundance. As noted above, the ijolites are rich in $(\text{Ti}, \text{Fe}^{3+})$ -bearing garnet, while the melilitolites contain $(\text{Al}, \text{Fe}^{3+})$ varieties. According to the tetrahedral site preference scheme $[\text{Si} > \text{Al} > \text{Fe}^{3+}]$, a clear distinction between garnets from the ijolite and melilitolite series can be drawn (Figure 5.1). All garnet analyses from the melilitolites have $(\text{Si} + \text{Al}) > 8$ cations, creating $^{\text{vi}}\text{Al}$ and an absence of $^{\text{iv}}\text{Fe}^{3+}$. Conversely, garnets from the ijolite series contain little $^{\text{vi}}\text{Al}$ but do contain $^{\text{iv}}\text{Fe}^{3+}$. This is in contrast with their bulk-rock compositions, discussed in a later chapter, where SiO_2 (ijolite) $>$ SiO_2 (melilitolite). Carmichael et al. (1970) pointed out that low SiO_2 activity in melilitite magmas would promote Al activity. Figure 5.1 shows these relationships clearly. It can be seen that the melilitolite garnets fall exclusively on the $^{\text{vi}}\text{Al}$ side, with low Ti, while the ijolite garnets show increasing Ti abundance with increase in $^{\text{iv}}\text{Fe}^{3+}$. It is important to note that the Si content of the melilitolite garnets is normally close to 6 apfu, whereas the ijolite garnets contain > 4.6 apfu. Thus, Figure 5.1 should not just be interpreted in terms of Al abundance, but rather Al (and Fe^{3+}) site allocation and cation co-ordination. The cluster of points around [0,0.5] is dominated by analyses from the pyroxenites and the calcite-melilitolite sample N.64.18. The cluster of melilitolite points around [0,1.0] represent analyses from the pegmatitic turjaite TUR I, which contains rare brown crystals of schorlomite, while the data points around [1.8, 0] represent analyses from melilitolite samples E.133.30, N.64.18, and TUR I.

Figure 5.1 Garnet compositions from the Turiy rocks. All analyses normalized with 16 cations : 24 anions, assuming $\text{Si} > \text{Al} > \text{Fe}^{3+}$ occupancy preference within the tetrahedral site. The analyses from the ijolites contain Fe^{3+} in the tetrahedral site while the analyses from the melilitolites contain octahedrally-co-ordinated Al. Analyses from the turjite and carbonatite rocks fall on both sides of ${}^{\text{iv}}\text{Fe}^{3+} - {}^{\text{vi}}\text{Al}$ dividing line. X and Y axis units = apfu.

The two other rock types which contain garnet are turjite and carbonatite, and their garnet compositions are also plotted in Figure 5.1. Turjite, described in Chapter 4.8 as exhibiting petrographic similarities to both the ijolites and melilite-bearing rocks, demonstrates its hybrid nature in terms of garnet composition. Most of the turjite garnet analyses fall on the right-hand side of the x-axis in Figure 5.1 and their overall range in composition almost encompasses most of the melilitolite and ijolite garnet analyses.

The carbonatite ^{vi}Al -bearing analyses are from garnet from the northern carbonatites associated with melilitolites, and some of these carbonatites also contain melilite. The three carbonatite garnet analyses that fall on the $^{iv}\text{Fe}^{3+}$ side of the x-axis in Figure 5.1 come from sample N.63.4, which is a pyroxene-calcite rock, also from the northern massif.

Thus, garnet analyses from melilite-bearing silicate rocks and melilite-bearing or related carbonatites within the northern massif contain ^{vi}Al . Garnet analyses from the ijolite series rocks contain $^{iv}\text{Fe}^{3+}$ and no ^{vi}Al , while the pyroxenite data contain little of either component. The turjite analyses spread across Figure 5.1 with a wide range of compositional variation. This study indicates that the site allocation of Al may be an important factor for distinguishing mineral analyses that come from melilite-bearing and related rocks.

5.2b Dimeric silicates

5.2bi Melilite group

Melilite is an unusual mineral due to its dimeric silicate structure ($\text{Si}_2\text{O}_7^{6-}$), Fe^{2+} and Mg in tetrahedral co-ordination, as well as its occurrence in Si-poor, Ca-rich magmas. Its main solid-solution components are $\text{Ca}_2(\text{Mg,Fe})\text{Si}_2\text{O}_7$ (akermanite_{ss}), $\text{CaNaAlSi}_2\text{O}_7$ (soda-melilite) and $\text{Ca}_2\text{Al}_2\text{SiO}_7$ (gehlenite) (Deer et al, 1986). The akermanite_{ss}-soda-melilite series occurs in the most SiO_2 -undersaturated member of the

silicate magma spectrum (melilitite/melilitolite), and also in carbonatites and ultramafic lamprophyres. Gehlenite (along with akermanite) is found primarily in Ca-rich metamorphic rocks, and in meteorites, although there is a significant gehlenite component in the melilites within the Oka Complex (Gold, 1986). Despite the fact that melilite is colourless in thin-section, it is primarily a ferromagnesian mineral, and is defined as one of the mafic group of minerals (Deer et al, 1986; Le Maitre, 1989).

Significant variations in melilite composition occur in the Turiy samples. Figure 5.2a shows variations in Na content vs Al apfu (10 cations : 14 anions) for Turiy melilites. In this figure all of the data are from turjaite except the two “N” samples which are melilite-calcite-carbonatites. “Afrikanda” is a melilite-bearing olivinite or “afrikandite” containing olivine, perovskite, magnetite and melilite from the Afrikanda Massif.

Na and Al form part of the soda-melilite molecule $\text{CaNaAlSi}_2\text{O}_7$, and if all Na and Al are accounted for by soda-melilite, then all points should fall on the 1:1 dashed line in Figure 5.2a. Analyses which fall above the line (from samples TUR I, S.162.300) contain $\text{Na} > \text{Al}$, which is in excess of that needed to form soda-melilite, and the excess Na forms $\text{CaNaFe}^{3+}\text{Si}_2\text{O}_7$. Analyses which fall below the line, from sample TUR II, contain a small gehlenite ($\text{Ca}_2\text{Al}_2\text{SiO}_7$) component. The relative location of the points along the dashed line indicates the proportion of the soda-melilite component present in the soda-melilite - akermanite - gehlenite solid solution series. For example, $\text{Na}=\text{Al}=1$ on Figure 5.2a is equivalent to 50% soda-melilite, as represented by analyses from samples TUR I, S.162.300 and C.190.105. The high soda-melilite content of the melilite correlates with the high abundances of volatile and alkali-rich minerals associated with the melilite in these samples. The two melilite-calcite-carbonatite samples [N.64.18m, N.64.19] from the northern complex contain melilite_{ss} with the lowest soda-melilite component (~30%), quite separate from the silicate melilite-bearing rocks. The presence of the gehlenite

Figure 5.2a Melilite compositions from the Turiy rocks. The Na-Al plot demonstrates both the amount of the soda-melilite component [$\text{CaNaAlSi}_2\text{O}_7$] ($\text{Na} = \text{Al} = 1 = 50\%$ soda-melilite component), the amount of the gehlenite component [$\text{Ca}_2\text{Al}_2\text{SiO}_7$] (indicated by $\text{Al} > \text{Na}$) and the amount of the $\text{CaNaFe}^{3+}\text{Si}_2\text{O}_7$ component (indicated by $\text{Na} > \text{Al}$). Both of these latter components are low in the Turiy rocks. It can also be seen that the analyses from the melilite-calcite samples from the northern complex [N.64.18m and N.64.19] contain smaller quantities of the soda-melilite component than the analyses from the silicate rocks, and that the melilite from the melilite-olivinite from Afrikanda only contains 20% soda-melilite, which is considered low for igneous rocks (Kapustin, 1980; Velde & Yoder, 1976). Y axis units = apfu.

Figure 5.2b Melilite compositions from the Turiy rocks. This plot of Na - Mg# demonstrates the relationship between the akermanite and soda-melilite components. It can be seen that samples with the highest enrichment in soda-melilite also have the lowest Mg#s, which may indicate a magmatic fractionation trend within these rocks. The analyses from the melilite-calcite samples from the northern complex [N.64.18m and N.64.19] have the highest Mg#s (81-95). Y axis units = apfu.

component in melilite from sample TUR II may be linked to the early-formed or possibly cumulus nature of this melilite-rich sample. Melilite from the melilite-perovskite-magnetite rocks in the Gardiner intrusion contains up to 35% soda-melilite component, including up to 7% of the ferri-sodamelilite component (Nielsen, 1980), while melilite from the Iron Hill intrusion is closer in composition to the Turiy melilites with 50% akermanite and 50% soda-melilite (Nash, 1972). Melilite from the Oka melilitolites is rich in gehlenite (~11%) (Gold, 1986).

Comparisons that can be drawn between melilites from different parts of Turiy are based on a very limited number of samples. However, it can be seen that there is a range of compositions between the samples from Kusnavalok (TUR I, TUR II and WKT), the type locality of turjaite, where layering is observed in some localities in the field (Figure 4.5). The texturally homogeneous melilitolite samples from the Central massif [C.5.118, C.190.105] are both volatile-rich (22 and 12% modal phlogopite respectively), while the pyroxene-bearing turjaite sample [E.133.30] from the Eastern massif shows melilite compositions (akermanite ~ 60 mol.%) close to the average of all the samples investigated from Turiy. All of the melilitolites investigated from the southern massif are extensively altered [e.g. S.78.15] but the relatively fresh ijolite-melilitolite contact sample [S.162.300] contains soda-melilite rich melilite_{ss}. The northern massif melilite-bearing samples include a melilite-calcite-carbonatite [N.64.19] and a calcite-melilitolite [N.64.18]. The melilite from both of these samples contains lower proportions of the soda-melilite component, and higher proportions of the akermanite - ferro-akermanite component, than the other melilite-bearing rocks.

The Mg#s of melilite (i.e. the Mg# of the akermanite - ferro-akermanite component) show an approximately inverse correlation between Na content and Mg#. This indicates that the Mg# and akermanite_{ss} content drop during magma fractionation, and the alkali content and hence Al content increase (Figure 5.2b). A similar trend is

seen in the ijolite pyroxenes with an increase in Na-Fe^{3+} (aegirine) as their Mg#s decrease, as will be shown in section 5.2ci. An approximate linear array shown by most of the melilite compositions in Figure 5.2b probably reflects a fractionation trend. The melilite compositions from TUR I which lie above the trend reflect the alkali-enriched nature of the whole rock by having atypically high Na contents for their Mg#s. The melilite from the melilite-bearing olivinite from Afrikanda has the highest Mg#, and the melilites from the two melilite-calcite-carbonatites again fall in a Mg-rich cluster separate from the melilites from silicate rocks.

To summarise, most of the samples show systematic variations in soda-melilite and akermanite components. However, the analyses from the melilite-rich turjaite TUR II contain excess Al (gehlenite component), those from the volatile-rich turjaite [TUR I] contain excess Na ($\text{CaNaFe}^{3+}\text{Si}_2\text{O}_7$ component), and those from the carbonatites have higher Mg#s than would be expected from the fractionation trend defined by the silicate rocks. The melilite analyses from the texturally homogeneous turjaite sample WKT from the turjaite type-locality, Kusnavalok, are consistent with the average melilite composition for this rock-type.

5.2c Chain silicates

5.2ci Pyroxene group

The clinopyroxene solid-solution series occurs extensively in Turiy rocks. Compositions range from pure diopside to 80 mol.% aegirine, and also encompass tschermakitic (aluminian sub-silicic) augites. In a way similar to the garnet series, the pyroxene compositions relate to the rock type in which they occur. Pyroxene compositions can also be used to help distinguish the separate phases of ijolite intrusion within Turiy. Bulakh & Ivanikov (1984, 1996) and Ronenson et al. (1978) suggested that there were two separate phases of intrusion of ijolite series rocks, pre- and post- intrusion

of the melilitolites. The first suite of ijolite rocks was believed to be more mafic (melteigite-ijolite) than the second (ijolite-urtite) (Bulakh & Ivanikov, 1984, 1996), although this is not borne out by the modal mineralogy of the ijolite samples examined. Systematic variations in mineral chemistry within minerals such as pyroxene and mica, may help to provide some answers to this problem.

It should be noted that in all figures used in this discussion $Mg\# = 100[Mg/(Mg+Fe^{2+})]$. In this definition, Fe^{2+} represents Fe_{total} except where Fe^{3+} has been calculated, in which case $Fe^{2+} = [Fe_{total} - Fe^{3+}]$. The pyroxene analyses have been normalized on the basis of 4 cations : 6 anions valence calculation applied to all samples. Although such normalization procedures may be subject to error, particularly with respect to Fe^{3+} values, these procedures are conventionally used as a first approximation of pyroxene composition. All references to “pyroxene” refer to “clinopyroxene”.

The pyroxenites have distinct Mg-Al-rich, Si-poor ($Mg\# > 90$) (Figure 5.3a&b) compositions with TiO_2 contents that range from 1.2 - 3.0 wt.%. Al substitutes for Si in the tetrahedral site, according to a $Si > Al > Fe^{3+}$ tetrahedral substitution preference scheme (Figure 5.3b), as well as occurring in octahedral co-ordination. Na contents are low, leading to the creation of Tschermak components $CaTiAl_2O_6$, $CaFe^{3+}AlSiO_6$, $CaAl_2SiO_6$ and $CaFe^{3+}_2SiO_6$ to charge-balance the Ti, Al and Fe^{3+} contents. In the absence of Mössbauer data for these samples (as is also the case for the garnet and mica analyses where the Fe^{3+} content has been calculated), the precise cation allocation cannot be confirmed, and the cation allocation scheme of Smyth (1980), adapted by Dunworth & Wilson (submitted), has been used for all pyroxene analyses in this study.

Although one of the pyroxenite samples [C.AG.347] has been dated as being ~200 Ma older than the majority of the rest of the complex (Chapter 6), it is still valid to look for, and examine similarities or differences between its mineral chemistry and the rest of the samples from Turiy. While its difference in age means that it cannot be directly

Figures 5.3ai&ii Fe^{3+} vs Mg# of clinopyroxene compositions from the Turiy rocks. Figure 5.3aii is an enlargement of the bottom right hand corner of Figure 5.3ai. TD 2 = a late-stage aegirine-apatite-carbonate vein on the south coast. The upper diagram shows the high aegirine component of the fenite pyroxenes, two analyses from the eastern ijolite sample [E.123.106] and the deep green groundmass clinopyroxenes from the olivine melteigite [C.AG.48A].

The lower diagram shows distinctions among the different rock groups. Most of the pyroxenite and carbonatite pyroxenes have $\text{Mg\#} > 91$, while almost all the ijolite analyses have $\text{Mg} < 91$. The olivine melteigite and phoscorite analyses span both groups, although most of the olivine melteigite analyses fall within the ijolite group. The five melilitolite pyroxene analyses do not appear to show systematic behaviour on this diagram. X and Y axis units = apfu.

Figure 5.3b Clinopyroxene compositions from the Turiy rocks. The close proximity of most of the data points to the $Al + Si = 2$ line indicates that there is just sufficient Al in these pyroxenes to fill vacancies in the Si site. Points that fall above the line contain ^{vi}Al , while points which fall below the line (e.g. ijolite WKM) contain $^{iv}Fe^{3+}$. It can also be seen that the pyroxenites contain significantly more ^{iv}Al in their pyroxene compositions than other rocks from within the massif. TD 2 = a late-stage aegirine-apatite-carbonate vein on the south coast. X and Y axis units = apfu.

Figure 5.3c Clinopyroxene compositions from the Turiy rocks. This plot emphasizes a lack of aegirine component in most of the analyses except those from the fenite, the late stage aegirine-apatite vein, and two ijolite pyroxenes. The bimodality of Tschermak enrichment vs aegirine enrichment in the Turiy rocks is clearly demonstrated in this plot. TD 2 = a late-stage aegirine-apatite-carbonate vein on the south coast. X and Y axis units = apfu.

related to any of the other rocks present (e.g. as a conjugate cumulus residue), comparison of its mineral chemistry with that of younger pyroxenites and rock types within Turiy will aid in the understanding of its magma chemistry, which is independent of age.

Total Tschermak components in the pyroxenite pyroxenes range from 11-21 mol.% and aegirine contents are < 4 mol.% (Table 5.A). Kapustin (1980) reports typical pyroxene compositions for Kola Peninsula clinopyroxenites as 70-90% diopside, < 20% hedenbergite and 12-15% aegirine-augite. Given the presence of chrome-spinel exsolution rods in the cores of the pyroxenes in sample C.AG.347 [pyroxenite], indicating high-pressure exsolution, one might expect the pyroxene host to contain significant ^{vi}Al which is often assumed to be an indicator of high-pressure conditions (Simonetti, 1994). The ^{vi}Al contents in the pyroxenes in sample C.AG.347 are low, but the Al and Cr contained in the chrome-spinel exsolution rods probably originated from the pyroxenes, making the current Al and Cr contents of the pyroxenes a minimum estimate of their original compositions. Cr₂O₃ was not included in the microprobe analysis table for the silicate phases due to a restriction on the number of elements which could be analyzed at any one time. It is believed that the Cr₂O₃ contents of the pyroxenes in pyroxenite sample C.AG.347 are minimum values given the presence of apparent Cr-spinel exsolution features within the cores of the crystals.

There are few pyroxenes present in the melilitolites, as the reaction

makes pyroxene and melilite incompatible with each other in a carbonate-rich magma unless there is not enough of the carbonate phase present to enable the reaction to go to completion. Pyroxenes are present in quantities > 1 modal % in samples TUR I and E.133.30, and are texturally associated with melilite and calcite. The composition of the pyroxenes in both of these samples is somewhat varied, but there is a tendency for the pyroxenes to contain a high percentage of Tschermak components (6.5-15.5 mol.%) and

Table 5.A   **Pyroxene compositions**

Sample No.	Rock type	Mg #	aegirine (mol.%)	Tschermak* (mol.%)
C.AG.347	pyroxenite	90-98	1.4-4.0	11.0-21.0
C.90.29	pyroxenite	86-91	4.0	17.3-21.0
N.63.4	carbonatite	92-100	1.5-3.0	3.5-15.0
TUR I	turjaite	86-98	3.5-6.0	6.5-15.0
C.18.26	phoscorite	88-94	2.7-6.2	2.5-15.0
C.AG.122	melteigite	80-90	3.7-10.7	2.5-14.0
C.AG.48A	ol. melteigite	77-90	2.5-8.5	4.9-14.0
N.187.100	carbonatite	85-100	2.0-10.1	4.5-9.6
C.ND.500	ijolite	85-90	4.4-5.4	3.5-4.7
<u>ij</u> -fen^	ijolite	85	4.5-7.7	3.5-5.0
WKM	ijolite	87-92	5.2-6.3	0-7.5
E.133.30	turjaite	83	1.0-5.0	0-6.8
C.IJ.14	ijolite	81-86	5-6.5	0-6.4
S.162.300	ijolite/(turjaite)	75-90	3.3-6.7	0-2.4
C.XC	carbonatite	97-100	5.5-8.8	3.1-4.4
N.19.64	carbonatite	92-96	1.2-1.9	2.6-4.3
C.TL.344	carbonatite	91-93	2.6-2.9	0.2-1.0
C.K.7	phoscorite	92-95	1.8-5.5	0
E.123.106	ijolite	60-85	4.5-43.0	0-4.3
C.90y	ijolite	< 75	8.0-20.0	0-4.2
TD 2	aeg-aug-ap vein	65-80	27.0-46.0	0
NC.153.82	fenite	47-76	33.0-45.0	0
C.AG.48A	ol. melteigite	42-44	24.0-35.0	0
<u>ij</u> -fen^	fenite	5-25	10.0-89.0	0

* Tschermak includes $\text{CaAl}_2\text{SiO}_6$, $\text{CaTiAl}_2\text{O}_6$, $\text{CaFe}^{3+}\text{AlSiO}_6$ and $\text{CaFe}^{3+}_2\text{SiO}_6$ components.

^ ij-fen is taken from a contact zone between an ijolite and a fenite. The part of the name which is underlined represents the side from which the pyroxene analyses were taken.

relatively low aegirine contents (<6 mol.%) (Figure 5.3c). Pyroxenes in the melilitolites from the Gardiner intrusions contain ~ 90% diopside (Nielsen, 1980).

The pyroxenes from the phoscorites and some of the carbonatites overlap the Mg#s and Fe^{3+} contents of the pyroxenite pyroxenes forming a group of analyses in Figure 5.3a_{ii} with $\text{Mg\#} > 90$ and $\text{Fe}^{3+} < 0.15$ apfu. The data from the phoscorite pyroxenes that plot among those from the silicate rocks come from a “phoscoritized pyroxenite” sample [C.18.26] which contains texturally heterogeneous pyroxenes, with variable compositions. Pyroxenes from a more typical phoscorite (C.K.7) contain no Tschermak components, and charge-balance low Na and Ti contents by components such as $\text{Na}(\text{Mg})_{0.5}(\text{Ti})_{0.5}\text{Si}_2\text{O}_6$. Like the pyroxenites and phoscorites, the Na contents are also low in the carbonatite pyroxenes. All the carbonatite pyroxene analyses, except those from sample N.63.4, contain fairly pure diopside with Tschermak components mostly < 5 mol.% (Figure 5.3c). Exceptions are sample N.187.100 where Tschermak components reach 8.5 mol.%, and sample N.63.4. Sample N.63.4 comes from part of the complex characterized by melilite-rich calcite-carbonatite, is essentially bi-mineralic (it could also be classified as a calcite pyroxenite), and its isotopic signature (Chapter 6) is the same as the nearby melilite-bearing rocks. Pyroxene compositions from sample N.63.4 contain 3.5-15 mol.% Tschermak components, low aegirine content (<3 mol.%), and are similar to pyroxenes from TUR I (melilitolite) and C.AG.347 (pyroxenite) (Table 5.A). It is suggested that sample N.63.4 is a product of reaction (iv) above.

The fenite pyroxenes contain aegirine-augite as an essential component, and have no Tschermak components present. Ti is charge-balanced by Na where necessary, similar to pyroxenes found in syenites in the Gardiner intrusion (Nielsen, 1979), and in groundmass pyroxenes in olivine melilitite from the Rhinegraben (Dunworth & Wilson, submitted). Sample NC.153.82, the least fenitized sample, contains pyroxenes with 33-45 mol.% aegirine. The two ijolite-fenite contact samples [ij-fen, C.190.x] contain up to

90% aegirine component in the pyroxenes from the fenite part of the samples, demonstrating the influence of fenitizing fluids rich in alkalis. Sample TD 2, a late-stage aegirine-augite-apatite-ankerite vein exposed on the south coast of the Turiy Peninsula contains pyroxenes with 25-50 mol.% aegirine (Figure 5.3ai).

The olivine melteigite [C.AG.48A] contains zoned, pleochroic Ti-bearing, aluminian augite with compositions lying between pyroxenes from the pyroxenites and ijolites ($\text{TiO}_2 = 0.3\text{-}2.0$ wt.%). $\text{Fe}^{3+} + \text{Al}$ contents are again higher than can be compensated for by Na, and Tschermak components (5-15 mol.%) must be invoked to maintain charge balance (Figure 5.3c). Some of these compositions match those found in the central massif ijolite sample C.IJ.14. The interstitial nepheline groundmass in the olivine melteigite encloses a few deep green, euhedral aegirine-diopside crystals which represent extreme fractionation products at the end of crystallization. They exhibit $\text{Na} > [\text{Fe}^{3+} + \text{Al}]$ in a number of cases, creating components similar to those needed to charge-balance the phoscorite compositions.

Finally, the ijolite series rocks, of which pyroxene is an essential constituent, exhibit an interesting variety of pyroxene compositions. Modal mineralogy (Appendix Biii) does not distinguish between the two generations of ijolites proposed by Bulakh & Ivanikov (1984) in terms of mafic : felsic index, and the compositional data from the pyroxenes do not exemplify such a division either. Instead, the modal ratio of garnet : pyroxene may be used with caution to distinguish between the early ijolites (pyroxene-rich) and later ijolites (garnet-rich).

Plots utilizing Fe^{3+} , Al, Si, aegirine contents, Tschermak components and Mg# are most useful for visualising the varying compositional relationships of the pyroxene analyses (Figs. 5.3di&ii). Most of the ijolite-series pyroxenes have Mg#s between 75 and 90, with C.90y and E.123.106 having the lowest Mg# values (Figure 5.3di). All of the data lie on, or close to the $\text{Al} + \text{Si} = 2$ substitution line in Figure 5.3b, indicating a lack of

^{vi}Al and fairly significant ^{iv}Al contents (< 0.17 apfu) according to ideal substitution patterns. One of the most interesting relationships is that between Fe³⁺ and Al. Figure 5.3dii indicates two trends, one involving Fe³⁺ enrichment with consistently low Al content as seen in samples C.90y, WKM and E.123.106 (i.e. a variation in the aegirine component) and the other showing a proportional relationship between Al and Fe³⁺, as demonstrated by samples C.ND.500, C.AG.122 and C.IJ.14 (i.e. a variation in Tschermak components). Most of the pyroxene compositions from sample S.162.300, the ijolite-melilitolite contact sample, have trace quantities of both Al and Fe³⁺ but are more enriched in Fe³⁺ than Al. The pyroxenes from an ijolite sample from the outer ring of the eastern complex [E.123.106] contain augitic cores (enrichment in Al with constant and low Fe³⁺) and aegirine-rich rims (enrichment in Fe³⁺ with constant and minimal Al contents) and thus appear to trend away from the (0,0) corner of the graph in two directions. Sample WKM, a fine-grained melteigite-ijolite body near Kusnavalok on the west coast, starts from equivalent Fe³⁺-Al enrichment, and then shows additional enrichment in Fe³⁺ which is related to the CaFe³⁺₂SiO₆ component. Few compositions contain Al > Fe³⁺ in Figure 5.3dii outside experimental error, apart from most of the pyroxene analyses from the olivine melteigite sample [C.AG.48A]. The effect of Ti content on the Al and Fe³⁺ values also needs to be considered. In the pyroxenites, for example, the incorporation of Ti creates components such as CaTi(Fe³⁺,Al)₂O₆ which charge-balance most of the remaining trivalent ions after the formation of the aegirine component.

Tschermak components, in general, are present in quantities of 0-7 mol.% in the pyroxenes from the ijolites, except for pyroxenes from sample C.AG.122 which contains 2.5-14 mol.%. Sample WKM contains unusually high (< 4 mol.%) CaFe³⁺₂SiO₆ contents. Sample S.162.300, the melilitolite-ijolite contact sample, contains pyroxenes with virtually no Tschermak components and a Na-Ti charge balance mechanism

$[\text{Na}(\text{Ti}_{0.5}\text{Mg}_{0.5})\text{Si}_2\text{O}_6]$ with low aegirine contents, similar to the phoscorite analyses discussed earlier, and similar to late-stage melilitite pyroxenes from Urach, Germany (Dunworth & Wilson, submitted). A study of such “Ti-aegirines” in the Gardiner province by Nielsen (1979) showed that they were part of a syenite fractionation trend, indicative of high SiO_2 activity but low Fe activity, common to peralkaline liquids (Marsh, 1975), formed under conditions of low pressure, but not necessarily low temperature. Additionally, it was suggested that the $[\text{Na}(\text{Ti}_{0.5}\text{Mg}_{0.5})\text{Si}_2\text{O}_6]$ component, believed to be present in some of the Turiy pyroxenes, formed under conditions of higher $f\text{O}_2$ than the $[\text{Na}(\text{Ti}_{0.5}\text{Fe}^{2+}_{0.5})\text{Si}_2\text{O}_6]$ component (Nielsen, 1979).

The trends exhibited by the pyroxene compositions also show some distinct variations. Fractionation trends in alkalic rocks tend to enrich the residual liquids in alkalis and decrease their Mg#s, hence creating aegirine enrichment (Mitchell & Platt, 1982). However, using the central massif ijolites as an example, Figures 5.3d(i&ii) shows that pyroxenes from melteigite sample C.AG.122 have high Mg# and Fe^{3+} content, but low aegirine contents, while pyroxenes from ijolite sample C.IJ.14 have decreasing Fe^{3+} and Al, and become more diopsidic in comparison to those from sample C.AG.122. The pyroxenes from sample C.90y are aegirine-rich, but do not appear to have fractionated from compositions such as those seen in sample C.AG.122, as seen in Figure 5.3dii.

Table 5.A compares and contrasts three major variants in pyroxene compositions (Mg#, aegirine content, and calculated Tschermak component content) from the Turiy massif. Two of these variants (aegirine vs Tschermak component contents) are plotted in Figure 5.3e. Although there is distinct overlap in the pyroxene compositions among the samples, some significant correlations and groupings can be seen. The pyroxenites [C.90.29, C.AG.347], a calcite-pyroxene rock [N.63.4] and melilitolite sample TUR I are similar in all three variables listed in Table 5.A. These similarities have implications for

Figures 5.3di&ii Clinopyroxene compositions from the ijolites at Turiy. Figure 5.3di (above) shows the relationship between Fe^{3+} and Mg#, while Figure 5.3dii (below) shows the relationship between Fe^{3+} and Al. These diagrams provide an opportunity to look at the ijolite data sample by sample, in an attempt to ascertain whether two distinct series of ijolite series rocks at Turiy can be identified on the basis of pyroxene compositions. See text for further discussion. X and Y axis units = apfu (except Mg #).

the formation of calcite-pyroxene sample N.63.4. This sample, associated geographically and isotopically with melilite-calcite carbonatites in the northern part of Turiy, contains no actual melilite, and may be a product of the reaction: $\text{melilite} + \text{CO}_2 = \text{diopside} + \text{calcite}$. The similarity of its pyroxene compositions to those of melilitolite sample TUR I appear to corroborate this petrogenetic relationship. If the reaction: $\text{melilite} + \text{CO}_2 = \text{diopside} + \text{calcite}$ has gone to completion from left to right in a melilitolite such as TUR I, then the calcite + diopside mixture seen in sample N.63.4 may well be at least part of the end product of the reaction. Given the fact that sample TUR I is also substantially enriched in volatiles and the pyroxenes in sample TUR I are enclosed within melilite ($\pm$ calcite), it is likely that they also formed from the reaction above. The similarity in the pyroxene compositions from samples N.63.4 and TUR I hence emphasizes the petrogenetic link that the former sample has with the melilitolites. Extreme enrichment in Tschermak components, high Mg #s and low aegirine contents are also typical of melilite-bearing extrusive rocks (e.g. Dunworth and Wilson, submitted) and from jacupirangites within melilitolite complexes such as Oka (Gold et al., 1986). Samples C.AG.122 [melteigite] and C.AG.48A [olivine melteigite], are enriched in Tschermak components, with intermediate aegirine contents. Sample C.18.26, a “phoscoritized pyroxenite” (listed as a phoscorite in Table 5.A) also has pyroxene compositions with high Tschermak components, high Mg#s and moderate aegirine contents, consistent with derivation from a pyroxenite or melteigite.

Similar Mg#s and aegirine contents but greatly reduced Tschermak contents are shown by ijolite samples WKM, C.ND.500 and the ijolite part of the “ijolite-fenite” contact sample. These compositions are closely followed in Table 5.A by pyroxenes from the eastern melilitolite [E.133.30], a central massif ijolite [C.IJ.14] and the southern melilitolite-ijolite contact, with low aegirine and Tschermak contents and lower Mg#s.

Most of the central massif carbonatites (carbonatite sample numbers starting with “C”) and the phoscorite sample C.K.7 have pyroxene compositions with very high Mg#s and very low aegirine and Tschermak components. The phoscorite pyroxene analyses have very low $\text{Fe}^{3+}/(\text{Fe}^{3+}+\text{Fe}^{2+})$ ratios. The $\text{Fe}^{3+}/(\text{Fe}^{3+}+\text{Fe}^{2+})$ ratio of pyroxenes from the carbonatites tends to be more varied, although the total Fe content in pyroxene compositions from both rock types is very low. The phoscorite analyses from sample C.K.7 form a distinct cluster in Figure 5.3e and are closest to carbonatite sample C.TL.344 and the melilitolite-ijolite contact sample S.162.300.

Finally, pyroxenes with high aegirine contents, and minimal or zero Tschermak content are found in the eastern ijolite [E.123.106], the urtite sample C.90y, the late-stage vein TD 2, the late-stage fractionation products of the olivine melteigite [C.AG.48A], and the fenites. Mg#s are also low in all of these samples. The pyroxenes from the ijolite-fenite contact samples [ij-fen; C.190.x] are substantially more enriched in aegirine and have much lower Mg#s than the fenite sample NC.153.82.

Variations in pyroxene composition can be attributed to two main factors: 1) variations in bulk magma composition, and 2) element partitioning between the different mineral phases present in each rock-type. Variations in bulk rock chemistry will be discussed in a later chapter. The role of element partitioning is clearly demonstrated by the pyroxene compositions. For example, ijolite sample C.IJ.14 contains 55% nepheline, 25% garnet and 5% pyroxene, while ijolite sample C.ND.500 has a similar nepheline content, but reversed modal percentages of garnet and pyroxene. Both samples come from the central complex, and their pyroxene compositions have sequential ranges of Mg# (81-86 and 85-90, respectively). Comparison of samples C.ND.500 and E.123.106, both of which contain almost identical quantities of pyroxene and garnet, have very different pyroxene compositions, in terms of both aegirine content and Mg#, as indicated in Table 5.A. Comparison of these two sample pairs just described indicates that element

Figure 5.3e Clinopyroxene compositions from the Turiy rocks. This plot uses data summarized in Table 5.A and compares the % contents of aegirine and Tschermak components in the analyses. Included in the Tschermak components are $\text{CaAl}_2\text{SiO}_6$, $\text{CaFe}^{3+}\text{AlSiO}_6$, $\text{CaFe}^{3+}_2\text{O}_6$ and $\text{CaTiAl}_2\text{O}_6$, calculated according to the method of Smyth (1980). The x-axis only displays analyses up to 12% aegirine component, although analyses with up to 80% aegirine component are found in the fenite samples.

partitioning has played very different roles within each pair.

It should also be noted that magnetite and other iron oxide minerals are rare in the ijolites, common in the melilitolites, pyroxenites and phoscorites, and have varied abundance in the carbonatites. Thus, Fe partitions into the mafic silicate phases, instead of non-silicate oxide phases in the ijolitic melt, a characteristic that may be related to the fO_2 of the magma.

The pyroxenoid wollastonite is found in trace quantities in the melilite-calcite samples N.64.19 and N.64.18m, and as coarse anhedral grains in the melilitolite-ijolite contact sample S.162.300. The cancrinite in sample TUR I contains needles of the pyroxenoid pectolite, described by Nielsen et al. (1997) as the “hydrous equivalent of combeite” which is characteristic of the peralkaline nephelinite magmas at Oldonyo L’Engai. Pectolite is also found in trace quantities in a few of the ijolite samples in Turiy.

To summarise, different groupings of pyroxenes can be distinguished, and these can be partly related to rock-type. Pyroxenes with high Mg# and Tschermak contents but low aegirine contents are predominantly found in pyroxenitic and melilitolitic samples. Samples with low Mg#, low Tschermak contents but high aegirine contents belong to a few of the ijolites, late-stage veins, and fenites. The carbonatite pyroxenes have very high Mg# and generally low aegirine and Tschermak components. Samples with intermediate quantities of all three variables generally belong to the ijolite series. The existence of two ijolite series, as described by previous field studies (Bulakh & Ivanikov, 1984, 1996), is not mirrored by the chemical composition of the pyroxenes contained within them, nor their mafic index, but may be distinguished on the basis of their modal garnet and pyroxene contents. Sample N.63.4, a pyroxene-calcite-carbonatite, is believed to be derived from a melilitolite through reaction of melilite with CO_2 .

5.2cii *Amphibole group*

Amphiboles are not common at Turiy, and occur only in the pyroxenites, phoscorites, carbonatites and rarely in the Proterozoic host rock. Amphibole is absent from the ijolite and melilitolite series.

Magnesio-hastingsite (Leake, 1978) occurs in the pyroxenites from Turiy, and its interstitial textures suggests that it may represent part of the residual melt after the crystallization of the pyroxene. Amphibole is more abundant in pyroxenite sample C.AG.347, than in sample C.90.29. The compositions of the amphibole are similar in the two pyroxenites, although they have different Mg#s. Experimental studies suggest that magnesio-hastingsite is stable at relative high oxygen fugacities (magnetite-hematite buffer) (Spear, 1981), and at $< 1100^{\circ}\text{C}$ (Stone et al., 1997 and references therein). Magnesio-hastingsite is also one of the first silicate minerals to crystallize in many carbonatites (Hogarth, 1989), which is interesting, given that calcite and amphibole are associated in sample C.AG.347.

Of special interest is an amphibole inclusion found in a pyroxene phenocryst in the olivine melteigite [C.AG.48A] which is almost identical in composition to the amphibole from pyroxenite sample C.AG.347, as seen in Figure 5.4. The similarity in amphibole compositions between the two samples cannot be disregarded, even though the two rocks appear to demonstrate different petrogenetic histories as will be seen throughout this study.

A significant compositional break occurs between the amphiboles present in the silicate rocks (low $\text{Na}/(\text{Na}+\text{Ca})$) and those in the phoscorites and carbonatites (high $\text{Na}/(\text{Na}+\text{Ca})$). Most of the amphiboles in the phoscorites and carbonatites are richterite, with rare magnesio-arfvedsonite in two samples. Richterite mainly occurs as radiating clusters of elongate crystals, which are probably either late-stage or secondary in most

Figure 5.4 Amphibole compositions from the Turiy rocks. The cluster of Na-rich richterite analyses with high Mg#s are from late calcite carbonatites and two phoscorites [C.K.7 and C.3.93] from the central complex. The cluster of data points in the centre of the plot are pargasite and magnesio-hastingsite analyses from the central complex pyroxenites [C.AG.347 and C.90.29]. Note that the analysis from an inclusion within a pyroxene grain in the olivine melteigite sample [C.AG.48A] which is identical to the analyses from pyroxenite sample C.AG.347. The isolated point on the left-hand side of the graph is from the fenite sample [NC.153.82].

samples. Modal abundances are listed in Appendix Biii and compositions in Appendix Jii. Na/(Ca+Na) contents increase with decreasing Mg# (Figure 5.4). Fe^{3+} contents of the amphiboles have not been calculated, so Mg# is calculated using Fe_{TOTAL} as Fe^{2+} . The analyses rarely overlap between samples, and the two phoscorite samples [C.K.7 and C.3.93] are consistently Mg-rich.

Rare ferroan pargasitic hornblende (Leake, 1978) was found within the fenite sample NC.153.82. Compositionally, the hornblende has low Mg# and intermediate Na:Ca content. It bears no resemblance to the amphiboles from the main intrusion.

The reason for the low abundance of amphibole in Turiy, and its complete absence from the melilitolites and ijolites, is not entirely clear. Mica is the dominant volatile-rich silicate phase within the melilitolites and ijolites. The absence of amphibole from the melilitolites corroborates the fact that amphibole, like feldspar, is generally regarded as incompatible in melilite-bearing rocks (Carmichael et al., 1970; Deer et al., 1992). Amphibole is listed by Hogarth (1989) as one of the commonest minerals in carbonatites, yet its presence in the Turiy samples is < 5 modal % in only a few samples.

5.2d Sheet silicates

5.2di Mica group

A wide variety of solid-solution variations are seen in representatives of the mica group, which appear in all rock-types in Turiy except the pyroxenites. Another common sheet-silicate, chlorite, is also found as an alteration product around the edges of some mica grains, most commonly in the carbonatites. Like the garnet group, different rock-types are characterized by specific mica compositions within a mineral group or solid-solution series.

Almost all of the micas are phlogopite whose basic formula is $\text{K}_2\text{Mg}_6\text{Al}_2\text{Si}_6\text{O}_{20}(\text{OH},\text{F},\text{Cl})_4$ with substitutions seen in all cation sites. Analyses from the

Turiy micas show that substitutions in the interlayer site are minimal in most samples, with K remaining as the dominant cation. Minor amounts of Ba (common in phlogopite from alkaline rocks, e.g. Zhang, 1993) and trace amounts of Na and Ca are the only substituting cations. Such substitutions reach a maximum in the melilitolites, and a minimum in the phoscorites.

The octahedral cation site, primarily filled by Mg in phlogopite, also contains Fe, Mn, Al, and Ti in the Turiy micas. The valency of the Fe in this site is unknown since Mössbauer spectroscopy was not carried out to quantify the $\text{Fe}^{2+} : \text{Fe}^{3+}$ ratio. In all analyses all ^{56}Fe is thus assumed to be Fe^{2+} . Mn contents are generally low (< 0.3 wt.% MnO except for carbonatite sample N.187.100 where the values rise to 0.6 wt.% MnO). Ti contents are partially correlated with Ba in micas from melilitites (Dunworth and Wilson, submitted), and while Ti is low in abundance in Turiy micas, it shows a maximum in the melilite-bearing samples and a minimum in the phoscorites, in common with the behaviour of Ba. ^{27}Al only occurs in the melilitolite micas, just as ^{27}Al only occurred in melilitolite garnets. A clear distinction can thus be drawn between the melilite-bearing and the non-melilite bearing samples on the basis of Al contents and site allocations. Vacancies are also common in the octahedral site.

The substitution variations in the tetrahedral site are dominated by Si, Al and Fe^{3+} . It is again assumed that the site allocation preferences occur in the order $\text{Si} > \text{Al} > \text{Fe}^{3+}$, whilst also recognising that there are two tetrahedral sites, one dominated by Si and the other by Al, in phlogopite. The presence of tetrahedrally co-ordinated Fe^{3+} is generally believed to be signified by orange pleochroism and reverse absorption (Rimskaya-Korsakova & Sokolova, 1964; Kapustin, 1980; Gaspar & Wyllie, 1987), and this is particularly typical of micas from the phoscorites, carbonatites, turjite, and some ijolites from Turiy.

Because of the lack of any Mössbauer spectroscopy data, it was necessary to find a way of characterising and quantifying the $^{iv}\text{Fe}^{3+}$ contents without spectroscopic studies. One approach is to normalise the analyses with 16 cations : 22 oxygen-equivalents (i.e. assuming full occupancy of all sites). Previous studies have shown that micas typically contain vacancies in both the octahedral and inter-layer sites, and assuming full occupancy of all sites would almost certainly lead to erroneous normalizations. In the normalization procedures used in this study, two assumptions were made: 1) that the tetrahedral site would maintain full occupancy (filled by Si, Al and Fe^{3+}), and 2) that the number of anions would also be a maximum. Similar assumptions were made by Hogarth (1997). The first of these assumptions is probably correct, but the second may not be so because substitution of OH^- for O^{2-} has previously been suggested for samples needing unusual charge-balance mechanisms (Zhang, 1993). Two difficulties with this substitution mechanism are the unlikelihood of the O-site actually containing OH^- ions, thus creating O-site vacancies and over-saturation of the OH^- site, and the difficulty of identifying samples in which this substitution mechanism occurs (Henderson & Foland, 1996). While anion-saturation may not be a realistic assumption, the calculations based on such saturation provide good results. A FORTRAN programme was kindly written by T.S. Ercit to iteratively calculate a normalization based on 8 tetrahedral cations and 22 oxygen-equivalents. This allows for conversion of enough $^{vi}\text{Fe}^{2+}$ to $^{iv}\text{Fe}^{3+}$ to fill any remaining vacancies in the tetrahedral site while maintaining charge-balance, and still allows for vacancies in both the octahedral and inter-layer sites. Any remaining Fe is assumed to be Fe^{2+} and in octahedral co-ordination. All analyses carried out on Turiy samples, and analyses from the literature used for comparison, have been recalculated using this programme. The results show suprisingly good correlation with the pleochroic schemes of the micas. The melilitolite micas, rich in ^{vi}Al , are invariably brown - straw,

the micas from the northern carbonatites associated with melilite-bearing rocks which also contain $^{\text{vi}}\text{Al}$ are pale green - straw (typical of micas from early carbonatites, (Kapustin, 1980)), whereas the phoscorite micas, rich in $^{\text{iv}}\text{Fe}^{3+}$, show bright orange - pale orange pleochroism and reverse absorption. The boundary between the two is delineated by micas from the turjite sample TU 120A, which show a mixture of pale green and orange pleochroism (and reverse absorption in the micas showing orange pleochroism), indicating the change from $^{\text{vi}}\text{Al}$ to $^{\text{iv}}\text{Fe}^{3+}$ -bearing compositions. The tetrahedral Fe^{3+} substitutes almost exclusively for tetrahedral Al, as is typical in such $^{\text{iv}}\text{Fe}^{3+}$ -bearing micas (Kapustin, 1980), leading to $^{\text{iv}}[\text{Al}_{0.5}\text{Fe}^{3+}_{1.5}\text{Si}_6]$ in the most extreme case [carbonatite C.49.40]. It is possible to argue, in the case of the garnet compositions, that the only reason that there is $^{\text{vi}}\text{Al}$ in garnet from the melilitolite samples is that the tetrahedral site in the melilitolite garnets is too Si-rich to allow significant $^{\text{iv}}\text{Al}$ substitution, forcing all the Al into octahedral co-ordination, and that a similar overall concentration of Al in garnets from the ijolites, for example, substitutes for Si in tetrahedral co-ordination, leaving none available for $^{\text{vi}}\text{Al}$ occupancy. However, in the melilitolite micas the tetrahedral Al site is completely filled by Al, and their $\text{Al}_{\text{total}}/\text{Si}$ ratios are the highest of all of the micas analyzed (Figure 5.5a).

The iterative normalization procedure has an important effect on the Mg# of the $^{\text{iv}}\text{Fe}^{3+}$ -rich samples, as significant proportions of $^{\text{vi}}\text{Fe}$ are “removed” from the octahedral site through conversion to $^{\text{iv}}\text{Fe}^{3+}$, hence significantly increasing the Mg# which is based on the remaining $^{\text{vi}}[\text{Mg}:\text{Fe}_{\text{TOTAL}}]$ ratio.

The compositions of the micas also provide a link between sample C.IJ.14 (ijolite) and sample TU 119 (turjite), as these are the only two samples that show a range in compositions from $^{\text{iv}}\text{Fe}^{3+}$ -bearing-biotite to annite with gradually decreasing Al content, providing another link in the puzzle of the origin of turjite.

Figure 5.5a Mica compositions from the Turiy rocks. All analyses have been normalized to 8 tetrahedral cations and 22 oxygen-equivalents. Tetrahedral Fe^{3+} has been calculated, where necessary to fill the tetrahedral site, as a result of this normalization process. It should be noted that the normalisation scheme assumes $\text{Si} > \text{Al} > \text{Fe}^{3+}$ preferential occupancy within the tetrahedral site, and that ^{vi}Al and ^{iv}Fe are thus mutually exclusive within a single analysis. On this plot, the x-axis refers to the $\text{Mg\#}$ of the octahedral site, where all remaining Fe (after conversion of $^{vi}\text{Fe}^{2+}$ to $^{iv}\text{Fe}^{3+}$) is assumed to be in the ferrous state. It should be noted that the melilitolite micas have the highest Al/Si ratios of all the samples. TD 2 = a late-stage aegirine-apatite-carbonate vein from the south coast.

Figure 5.5b which plots ${}^{\text{iv}}\text{Al}$ vs ${}^{\text{vi}}\text{Al}-{}^{\text{iv}}\text{Fe}^{3+}$ shows clearly the change from melilite-bearing samples (including two melilite-rich calcite carbonatites) to non-melilite-bearing samples. The melilitolite micas contain substantial ${}^{\text{iv}}\text{Al}$ but more importantly ${}^{\text{vi}}\text{Al}$ instead of ${}^{\text{iv}}\text{Fe}^{3+}$ (these latter two are incompatible, given the constraints of the normalization scheme). The other mica analyses that also contain ${}^{\text{vi}}\text{Al}$ are those from two carbonatites (which also contain significant ${}^{\text{vi}}\text{Al}$ in their garnet compositions and are associated with melilite-bearing rocks and carbonatites), a phoscoritized pyroxenite [C.18.26], a melteigite with significant Tschermak components in its pyroxene compositions [C.AG.122], sample C.ND.500 which is the ijolite part of an ijolite-melilitolite contact sample, and two analyses from the west-coast ijolite [WKM]. The turjite mica analyses, as already mentioned, have a broad compositional spread, similar to the wide range of compositions seen in the turjite garnet analyses in Figure 5.1. The turjite analyses which plot in the melilite-bearing area are from sample TQ, the turjite with relict quench melilite from the chilled margin of a dyke. It can also be seen from Figure 5.5b that the phoscorites and carbonatite sample C.49.40 are the most depleted in ${}^{\text{iv}}\text{Al}$, and that the melilite-bearing samples contain the highest total Al. This latter point is emphasized in Figure 5.5c, a plot of Al_{total} vs ${}^{\text{iv}}(\text{Fe}^{3+}-\text{Al})$, as a change in slope seems to occur at the boundary of the ijolite/melteigites and the melilite-bearing samples. Figure 5.5a also shows that most melilitolite micas also have the highest Al/Si ratio.

Probably the most critical plot of mica compositions is that of ${}^{\text{vi}}\text{Mg}\#$ vs ${}^{\text{vi}}\text{Al}-{}^{\text{iv}}\text{Fe}^{3+}$ (Figures 5.5di&ii) which shows very distinct groupings and different fractionation trends. One “horizontal” trend, shown by the carbonatites and phoscorites, exhibits virtually continuous, high ${}^{\text{vi}}\text{Mg}\#$ s with changing ${}^{\text{iv}}\text{Fe}^{3+}$ contents. A second trend is dominated by the majority of the silicate rocks, starting with the melilitolites with high $\text{Mg}\#$ s on the ${}^{\text{vi}}\text{Al}$ -bearing side of the x-axis, trending across the x-axis to ${}^{\text{iv}}\text{Fe}^{3+}$ -bearing compositions from the ijolites and some turjites, and exhibiting a trend of decreasing $\text{Mg}\#$ with

Figure 5.5b Mica compositions from the Turiy rocks. The melilitolite micas almost all plot within the ^{vi}Al field while the majority of the micas from other rock types plot within the $^{iv}Fe^{3+}$ field. Any analyses which contain $^{iv}Al > 2$ imply that Al is also substituting for Si within the Si tetrahedral site. TD 2 = a late-stage aegirine-apatite-carbonate vein on the south coast. X and Y axis units = apfu.

Figure 5.5c Mica compositions from the Turiy rocks. The diagram illustrates another way of demonstrating both the enrichment in ^{vi}Al (y axis > 2) and substitution of Al for Si, both of which are almost unique to the melilitolite micas. TD 2 = a late-stage aegirine-apatite-carbonate vein on the south coast. X and Y axis units = apfu.

increasing ${}^{\text{iv}}\text{Fe}^{3+}$ content. A third group of analyses appears to lie between the two trends already described, and includes a cluster of melilitolite-ijolite-carbonatite-turjite mica analyses with values of about $\text{Mg\#} = 80$. The melilite-bearing rocks with the highest phlogopite $\text{Mg\#}$ s are melilite-bearing calcite carbonatites from the northern complex (included in the “melilitolite” group), just as the ${}^{\text{iv}}\text{Fe}^{3+}$ -bearing phlogopites from the central complex carbonatites have higher $\text{Mg\#}$ s than those of the ${}^{\text{iv}}\text{Fe}^{3+}$ -bearing phlogopites from the ijolites which surround the carbonatite stock. The cluster of carbonatite mica analyses around (0.25, 93) come from sample C.54.x, a late-stage natrolite-calcite carbonatite. Phlogopite from the melilitolites follow a fractionation trend in ${}^{\text{vi}}\text{Al}$ space continued by the ijolites, turjites and carbonatites in ${}^{\text{iv}}\text{Fe}^{3+}$ space. It is interesting that a group of turjite analyses from different samples mimic the melilitolite-ijolite trend on the right-hand side of Figure 5.5d, but the turjite analyses have consistently lower $\text{Mg\#}$ s.

A number of studies on Al-activity and co-ordination state in silicate melts using Raman spectroscopy have been carried out by Bjorn Mysen and his colleagues at the Carnegie Institute of Washington, and the applicability of their results to the Turiy samples will be discussed in Chapter 8. It has been shown in this thesis that the co-ordination state of Al in the minerals, and presumably in their host melt, is an important characteristic of the samples. Whether the consistent octahedral co-ordination of Al in minerals in melilitolites is related to the co-ordination state of Al in the actual melt, and whether it has a bearing on actual melilite formation, are both factors that have never been fully investigated. Essentially, the fact that Al occurs in octahedral co-ordination in micas (and garnets) relates to the formation of “Tschermak components” in these minerals (e.g. ${}^{\text{vi}}\text{Al}^{\text{iv}}\text{Al}$ substituting for ${}^{\text{vi}}\text{Mg}^{\text{iv}}\text{Si}$) - a factor normally considered in pyroxenes and amphiboles, but rarely in micas or garnets. As we have already seen (Table 5.A), Tschermak components are richest in pyroxenes from or associated with

Figure 5.5d Mica compositions from the Turiy rocks. Figure 5.5dii is an enlargement of the top right hand corner of Figure 5.5di, and both graphs demonstrate the relationship between octahedral site Mg# and $^{vi}Al - ^{iv}Fe^{3+}$ occurrence.

There are several points to note about this graph. The first concerns the melilitolite data (including data from melilite calcite carbonatites) which plot to the right of the x-axis, signifying the presence of ^{vi}Al . Also to the right of the x-axis is a cluster of carbonatite data points from the late stage, zeolite-bearing calcite carbonatite sample C.54.x. Almost all other analyses plot to the left of the x-axis, signifying the presence of $^{iv}Fe^{3+}$ in these analyses.

Next, it should be noted that most of the carbonatite and phoscorite data form a trend with constant but high Mg# and varying amounts of $^{iv}Fe^{3+}$. This trend does not continue into ^{vi}Al space.

There is a second trend of data points involving most of the mica analyses from the silicate rocks, starting with the high Mg# melilitolites to the right of the x-axis, and trending down the graph, across the x-axis, through the olivine melteigite and ijolite data, to Mg# ~ 60. This trend includes a cluster of carbonatite analyses around Mg# 75 and the data from sample TD 2 (the late-stage, aegirine-augite apatite vein from the south coast).

A final cluster of data points between these two previous trends (around Mg# 80) includes data from the central complex carbonatites, one melilitolite sample, and a number of the turjite analyses. The rest of the turjite analyses fall to the right of melilitolite-ijolite trend, and contain ^{vi}Al . X axis units = apfu.

melilite-bearing rocks.

To summarise, micas are common to almost all rock-types in Turiy and show a range of compositions dependent on the rock-type in which they occur. The possibility of using micas as “indicator minerals” in such massifs was noted by Kapustin (1980). In Turiy, the phoscorites and carbonatite C.49.40 show enrichment in $^{iv}\text{Fe}^{3+}$ and the majority have high $^{vi}\text{Mg\#}$ s. Almost all the melilitolite mica analyses contain ^{vi}Al , unlike almost all other samples. The significant ^{vi}Al component in the melilitolite micas is taken as a sign of Tschermak component involvement in the mica compositions. A trend dominated by the ijolite micas shows a fractionation curve with decreasing $^{vi}\text{Mg\#}$ and increasing $^{iv}\text{Fe}^{3+}$, and an intermediate trend may be outlined by a few central-complex carbonatites and some of the turjite analyses. The turjite analyses, in general, are scattered across the range of compositions exhibited by the other samples, emphasizing their hybrid nature, in terms of texture, chemistry and origin.

5.2e Framework silicates

5.2ei Feldspathoid group

Nepheline is an essential phase in ijolites and most melilitolites from Turiy, and a minor constituent of the pyroxenites. It is not found in the carbonatites, phoscorites or fenites, although Bulakh & Ivanikov (1996) report trace quantities of nepheline in some late-stage carbonatites. No potassic feldspathoids such as leucite and kalsilite have been discovered anywhere at Turiy.

All the nepheline compositions fall between 70-90% NaAlSiO_4 . There is no systematic variation between the compositions of nepheline from the ijolite and melilitolite series rocks, although average compositions indicate that the melilitolite nephelines are slightly more enriched in the NaAlSiO_4 component. The nepheline from the groundmass of the olivine melteigite [C.AG.48A] is the most enriched in the

NaAlSiO₄ component (90%) and there is a compositional gap between this and the Na-rich melilitolite nephelines. The low KAlSiO₄ content of the olivine melteigite nepheline analyses is unusual for intrusive rocks, but is found in association with kalsilite exsolution features in some volcanic samples (Hogarth, pers. comm. 1997). No such exsolution features are, however, seen at Turiy, and the reason for the low K content in the nepheline from this sample is unknown.

5.2eii *Feldspar group*

Potassium and plagioclase feldspars are limited to the aegirine-augite-apatite-ankerite vein [TD 2], the fenite part of the two ijolite-fenite contact samples [ij-fen and C.190.x] and the fenite sample [N.153.82]. It should be noted that melilite and feldspar are generally incompatible due to differences in SiO₂ activity needed to stabilize each mineral (Carmichael et al., 1970; Deer et al., 1992), although Sr-melilite was found in equilibrium with potassium feldspar in a volcanic rock from Morocco (Velde & Rachdi, 1988). While ijolites may contain feldspar, according to the classification of Le Maitre (1989), no feldspar has been found in any of the Turiy ijolites. Feldspar is also a permissible constituent of carbonatites (Hogarth, 1989), but again, none has been found in any of the Turiy samples examined in this study.

Sample TD 2 is the closest approximation to a feldspar-bearing “carbonatite” seen at Turiy. It contains 20% [ankerite + calcite] and almost pure albite and potassium feldspar, in the groundmass. The two ijolite-fenite contact samples [ij-fen and C.190.x] contain medium- to coarse-grained potassium feldspar on the fenite side of each contact. The fenite sample [NC.153.82] contains abundant feldspar, most of which is either pure albite or potassium feldspar with 3-45% albite exsolution. Morogan & Martin (1985), in a description of fenitized crustal xenoliths in the natrocarbonatite lavas at Oldonyo L’Engai, suggest that the high temperature mineral assemblages developed during initial

fenitization re-equilibrate at lower temperatures and are thus highly ordered, e.g. sanidine re-equilibration to form microcline. No calcic plagioclase or nepheline are present in any of the Turiy fenite samples. Large rounded crystals of potassium feldspar in sample NC.153.82 also contain wollastonite within apparent perthitic exsolution lamellae (Figure 4.16).

Euhedral celsian $[\text{BaAl}_2\text{Si}_2\text{O}_8]$ is a significant component of the groundmass of turjite sample TU 120A. Hydrothermal mineralization commonly contains Ba and Mn minerals (Deer et al., 1992) and thus it may be significant that ilmenite with a considerable pyrophanite $[\text{MnTiO}_3]$ content is also found in sample TU 120A.

5.2eiii *Neosilicates*

In a silica-depleted intrusion such as Turiy, quartz is a rare constituent, although it is considered a common component of late-stage ankerite-bearing veins (Kapustin, 1980). In Turiy, microcrystalline quartz occurs in a vein in one of the northern carbonatites which appears to be associated with a garnet-pyroxene band which shows signs of incipient alteration. Quartz also occurs interstitially within the groundmass in the calcite-fluorite vein sample TD 1, and contains rare Nb-rich anatase/brookite(?) needles. Neither quartz nor nepheline have been found in the fenite side of any of the contact ijolite-fenite samples or in sample NC.153.82 [fenite].

Hydrous alkali-silicates are common as groundmass minerals in some samples. Analcime is the main constituent of the groundmass in turjite sample TU 119 and is also found in the fenitized country rock sample NC.153.82. Natrolite is an abundant constituent of the carbonatite sample C.54.x where it encloses veins of apatite and concentrically zoned, rhombohedral calcite. This is typical of natrolite occurrences in late carbonatite veins (Kapustin, 1980). Natrolite is also found in both ijolite-fenite contact samples. Cuspidine, which occurs in the melilitolites in trace quantities, contains

up to 14 wt.% F (Appendix Jix). Cuspidine has also been found in the Fort Portal carbonatites (Nixon & Hornung, 1973). Several unidentified hydrous silicates are listed in Appendix Ji. under “additional minerals”.

5.3 Non-silicates

5.3a Oxides

5.3ai *Fe-(Ti) oxides*

Members of both the titanian magnetite (spinel) and ilmenite series are present at Turiy, although the former is much more abundant than the latter. Ilmenite almost always occurs as micro-intergrowth blebs (Lattard, 1995) within magnetite which are commonly rich in Mg or Mn. Nb-rich anatase (rutile?) is found in the late-stage veins TD 1 and TD 2, normally as needle-like inclusions in quartz.

Magnetite and ilmenite analyses listed in Appendix Jxvi have been normalized with 3 cations : 4 oxygens for magnetite and 2 cations : 3 oxygens for ilmenite. Additional elements beyond those listed in the analyses, are below detection limits. Calculated Fe^{3+} contents, based on the normalization procedures, will thus represent maximum values, given that trace quantities of other elements present in the minerals will actually occupy space which has been theoretically allocated to Fe^{3+} . It is also assumed in these calculations that Fe is the only element present in these samples in more than one valence state although Mn and Ti can also exhibit multiple valencies.

An additional analytical problem is the characterization of ilmenite micro-intergrowths. Some samples contain magnetite with ilmenite micro-intergrowth blebs that are large enough to be individually analyzed. Some other micro-intergrowths, however, occur on such a fine scale that the microprobe beam diameter was too large to analyse the two oxide phases, hence providing an average composition for magnetite + ilmenite combined. Thus, the comparison of magnetite compositions from different

samples should be carried out with caution.

Ilmenite micro-intergrowth features within magnetite can take a number of forms (Lattard, 1995), and the typical lamellae described by Buddington & Lindsley (1964) are rare in Turiy magnetites. Instead irregular, almost interstitial ilmenite blebs are far more common, and their abundance is roughly proportional to magnetite grain size. The blebs are rarely more than 20µm diameter and are normally much smaller. Lattard (1995) showed that such intergrowths are not just the product of a redox reaction within magnetite (Buddington & Lindsley, 1964), but can also be due to vacancy relaxation in magnetite causing exsolution, according to the reaction:

This reaction can only be detected by analysing the oxygen content of the two phases involved (Lattard, 1995), since the reaction involves conservation of oxygen. Because oxygen was not measured in these samples, it is assumed that magnetite-ilmenite micro-intergrowths in the Turiy samples are solely due to redox reactions within the rocks.

Compositional trends of magnetite are shown in Figure 5.6. A number of variations and trends can again be seen, with composition dependent on rock type. $\text{Fe}^{3+} \#$ (defined as calculated $100\text{Fe}^{3+}/[\text{Fe}^{3+} + \text{Fe}^{2+}]$) and $\text{Fe}^{2+} \#$ (defined as $100\text{Fe}^{2+}/[\text{Fe}^{2+} + \text{Mg}]$), were used to evaluate fractionation trends, as well as Al and Ti abundances.

Suprisingly, the Al content is one of the most important variable in distinguishing magnetite analyses from different rock types (Figure 5.6a). The olivine melteigite contains high proportions of $(\text{Fe}^{2+}, \text{Mg})(\text{Cr}, \text{Al})_2\text{O}_4$ and most of the analyses have higher Al than is seen on the Al scale shown in Figure 5.6a. The melilitolites and pyroxenites have the highest Al contents which are not associated with Cr substitution (i.e. pure spinel substitution MgAl_2O_4). This is similar to using the distinctive Al co-ordination characteristics in the melilitolite and pyroxenite garnet, pyroxene and mica data analyses. It should be remembered that, in addition, pyroxenite sample C.AG.347 also contains

Figure 5.6a Magnetite compositions from the Turiy rocks. Note that the melilitolite, pyroxenite and turjite magnetite analyses have higher Al contents than the rest of the samples. Most of the olivine melteigite data is off scale, due to the presence of a significant spinel component in the analyses. $\text{Fe}^{2+}\# = 100(\text{Fe}^{2+}/(\text{Fe}^{2+}+\text{Mg}))$. X axis units = apfu.

Figure 5.6b Magnetite compositions from the Turiy rocks. The effect that extensive chrome spinel substitution has on the olivine melteigite magnetite analyses can be seen in this graph. Note also the relative purity of the magnetite from many of the carbonatites and phoscorites in the top left hand corner of the diagram. X axis units = apfu.

Figure 5.6c Magnetite compositions from the Turiy rocks. The magnetite - ulvöspinel solid-solution trend is shown by the negative slope demonstrated by most of the data. The carbonatites and phoscorites (which contain significant microintergrowth blebs of ilmenite within the magnetite) are most enriched in the magnetite component, while the melilitolites (which show no microintergrowths of ilmenite) are more enriched in the ulvöspinel component. Data from the olivine melteigite do not fall on the trend due to the relatively low concentrations of the magnetite and ulvöspinel components in these analyses. X axis units = apfu.

Figure 5.6d Ilmenite compositions from the Turiy rocks. Ilmenite is only found as exsolution features in magnetite in the phoscorites, carbonatites and occasionally in the ijolites. In the turjite sample TQ, ilmenite is found as isolated grains, in trace amounts. The carbonatite and phoscorite analyses are significantly enriched in Mg (geikelite component) and less enriched in Mn (pyrophanite component). The analyses from the ijolite and turjite contain no Mg, but variable quantities of Mn. X axis units = apfu.

chrome spinel exsolution lamellae within the pyroxene cores.

The olivine melteigite magnetite analyses form distinct patterns in Figures 5.6(b&c). The plot of $\text{Fe}^{2+}\#$ vs Ti shows that, with increasing Ti content, there is an decrease in $\text{Fe}^{2+}\#$ and then a reversal of this trend due to an inverse relationship between Ti and Cr contents. The reversal in the $\text{Fe}^{2+}\#$ trend occurs as Cr drops below 0.5 apfu (not shown). Cr-free analyses from the olivine melteigite have high Ti contents and high $\text{Fe}^{2+}\#$ s, with no ilmenite exsolution.

The melilitolite and pyroxenite magnetite composition fields overlap in $\text{Fe}^{2+}\#$ and Al space, again suggesting a possible petrogenetic relationship between these two rock types. General variation also occurs between trends of magnetite analyses from *individual* pyroxenite and melilitolite samples (not shown), in that the melilitolite trends are variable in Al and Ti space, but relatively constant with respect to $\text{Fe}^{2+}\#$, whereas magnetite from each of the individual pyroxenite samples varies with respect to all three variables used in Figures 5.6(a&b). The calcite melilitolites [N.64.18 and N.64.18c] have a bimodal distribution of magnetite compositions with $\text{Fe}^{2+}\#$ inversely proportional to Al content at constant Ti values (not shown diagrammatically). The magnetite analyses from the melilite calcite carbonatite [N.64.19] (not shown) have similar Ti contents to N.64.18(c) but lie between the two groups of analyses from N.64.18(c) in $\text{Fe}^{2+}\#$ and Al space. The ijolites have higher $\text{Fe}^{2+}\#$ values and lower Al contents than the pyroxenites and melilites, while the $\text{Fe}^{3+}\#$ s and Ti contents of the pyroxenites, melilitolites and ijolites are all fairly similar.

The range of magnetite compositions from the three Kusnavalok turjaite is interesting. In TUR II, the melilite-rich sample, magnetite occurs as coarse-grained, subhedral crystals with perovskite overgrowths, as well as fine-grained rounded crystals lacking perovskite overgrowths. This suggests that there are at least two generations of magnetite in this sample. The perovskite rims also have an effect on the compositions of

the enclosed magnetite grains. The perovskite rims are believed to have formed by reaction of the magnetite with the Ca-rich host melt, in which Ti is derived from the rims of the magnetite. The resultant chemical zonation within the magnetite grain from core to rim is thus characterized by decreasing Ti content. Perovskite overgrowths and/or reaction products are common to most of the melilitolites, but the variations in magnetite composition are greatest in TUR II. The magnetite zonation (Fe increase from core to rim) as a result of Ti-extraction into perovskite rims should, however, be treated with caution as Dawson et al. (1996) also reported an enrichment in Fe in magnetite rims in carbonatites, unrelated to perovskite rims. There is a five-fold variation in Ti content in the magnetite analyses from sample TUR II, but their Fe^{2+} #s remain low (i.e. Mg-rich), and Fe^{2+} #, Ti and Al contents are similar to those found in the melilite-calcite-carbonatite N.64.18 and C.AG.347 [pyroxenite].

Magnetite is rare in the ijolites, and only five analyses were obtained from all of the samples. The analyses tend to have low Al but variable Ti contents, and their Fe^{2+} # values are generally constant and high. The magnetite from C.ND.500 [ijolite from an ijolite-melilitolite contact] has high Fe^{2+} # and low Ti and Al values that are similar to values from WKM [western ijolite] in Fe^{2+} # - Al space, while analyses from sample WKM have higher Ti contents. It should be noted that the WKM magnetite grains also have perovskite rims, similar to those found in the melilitolites.

The phoscorites and carbonatites have similar magnetite compositions, with relatively high Fe^{2+} #s (> 90) and low Ti and low Al contents. Only analyses from one sample N.63.4, a pyroxene-calcite-carbonatite, do not follow this trend. This sample is associated with melilite-calcite-carbonatites in the northern stock and has pyroxene compositions which overlap with those of the melilitolites. Similarly, the magnetite compositions from sample N.63.4. also overlap those from the northern melilite-bearing samples in Fe^{2+} #, Al and Ti space (Figures 5.6a&b).

Figure 5.6c, a plot of Fe^{3+} # vs Ti, reflects the purity of the magnetite component in the analyses (Fe^{3+} # = 67%, Ti = 0 for pure magnetite). The phoscorites and carbonatites dominate the magnetite-rich end of the trend, while the chilled-margin turjite sample [TQ] defines the Ti-rich end. Most analyses from the other rock-types fall in between, with the pyroxenite and WKM analyses filling a compositional gap in the data from the melilitolite magnetite analyses. The Ti-rich group of melilitolite magnetite analyses is marked by data from melilitolite samples TUR II, E.133.30 and S.78.15.

The turjite magnetite analyses are commonly compositionally associated with those from the eastern melilitolite sample [E.133.30]. This is interesting given that sample E.133.30 contains relatively alkalic pyroxene and abundant nepheline, with low melilite content. It was suggested in the previous chapter that turjite represents a reaction product between an ijolite and a carbonatite. However, the link between the magnetite analyses from the turjite and melilitolite sample E.133.30 raises the question as to whether turjite could also represent a reaction product between a melilitolite and a carbonatite. Similarities have already been seen between turjite and melilitolite garnet and mica analyses, and laths of relict melilite are also seen in sample [TQ], the chilled margin of a turjite dyke. This similarity now reported in the magnetite analyses adds another piece of evidence to the possible melilitolite - turjite link.

The fact that most of the magnetite analyses from Turiy form an inverse correlation between Fe^{3+} # and Ti is an indication that the primary substitution mechanism is contained within the magnetite-ulvospinel series. The one exception to this negative correlation is demonstrated by the olivine melteigite analyses which show a positive correlation, but which are also dominated by chrome-spinel substitutions.

Ilmenite analyses from micro-intergrowth features (Lattard, 1995) within magnetites from ijolites, phoscorites, carbonatites and turjite, show very Mg and Mn-rich compositions, commonly found in micro-intergrowth features from magnetite in igneous

and metamorphic rocks (Deer et al., 1992). FeTiO_3 remains the dominant solid-solution component, but pyrophanite (MnTiO_3) and geikeilite (MgTiO_3) components, together, account for up to 55%. An important fact is that ilmenite from the phoscorites and carbonatites is rich in Mg, whereas the ilmenite from one ijolite sample contains no Mg, but significant quantities of Mn (Figure 5.6d). The turjite sample TU 120A contains ilmenite with undetectable Mg and low Mn contents. A comparison with ilmenite from other SiO_2 -poor rocks, shows that ilmenite from kimberlite is commonly rich in MgTiO_3 while $(\text{Mn} > \text{Mg})\text{TiO}_3$ -rich ilmenites are commonly found in carbonatites (Deer et al., 1992). The carbonatite compositions reported by Deer et al. (1992) are the reverse of what is seen in ilmenite from the Turiy carbonatites. The absence of ilmenite micro-intergrowth features in magnetite from the melilitolites or pyroxenites, is also characteristic of the Gardiner intrusion (Nielsen, 1980). The Mg-Mn ilmenite relationship, again, appears to provide a way of distinguishing between rock-types, using mineral composition, although the data-base for the ijolite sample is somewhat small, due to the lack of oxide phases in the majority of the ijolite samples analyzed.

In summary, the presence of Al (although not co-ordination state) can be used to identify magnetite analyses from the melilitolites and pyroxenites, compared to analyses from the ijolites series rocks and the carbonatites and phoscorites. Compositional variations in magnetite were previously used by McMahon & Haggerty (1979) for similar purposes. The olivine melteigite [C.AG.48A] is the only sample to contain a significant chrome spinel solid-solution component, along with the chrome spinel exsolution lamellae in pyroxenite sample C.AG.347. Magnetite is rare in the ijolite series rocks, but common in all other rock types in Turiy. Identifiable ilmenite micro-intergrowth features in magnetite are only found in carbonatites and phoscorites, and rarely in the ijolites and turjite rocks.

5.3a_{iii} *Perovskite*

Two types of perovskite are found at Turiy. The first, normally a deep brown colour, occurs as discrete, fine- to medium-grained crystals with cores which may contain a multitude of solitary inclusions. The second type occurs as thin (5-100 μ m) colourless rims around the edges of magnetite crystals. The perovskite rims around magnetite may be subsequently overgrown or replaced by minerals such as garnet or phlogopite. Such perovskite rims are especially prevalent in the melilitolites, and similar rims have been reported in the melilitolites from the Gardiner complex (Nielsen et al., 1997) where they were attributed to Ti exsolution from magnetite during “re-equilibration”. Perovskite in the melteigites from Gardiner were reported to be rimmed by magnetite and titanite (i.e. an apparent reversal of the rims seen in the melilitolites) by Nielsen (1980). Perovskite rims have also been reported around Mg-ilmenite in kimberlites (Boctor & Boyd, 1979).

There is a compositional difference between the two types of perovskite. The solitary dark brown crystals are almost pure CaTiO_3 , with some enrichment in Nb, while the rims contain < 10% of the loparite component ($\text{NaREETi}_2\text{O}_6$, Mitchell, 1996) as well as latrappite ($\text{Ca}_2\text{Fe}^{3+}\text{NbO}_6$) to account for the Nb substitution. This is a reversal of the enrichment seen in the Gardiner perovskites (Nielsen, 1980), where early perovskites are rich in Nb and REE and the rims are almost pure CaTiO_3 . Figure 5.7 is a plot of Na vs [La + Ce] for all Turiy perovskite analyses. It can be seen that the Na concentrations are proportional to the REE for most samples, indicating varying quantities of loparite. The melilitolites generally have atomic $\text{SLa+Ce} > \text{Na}$, and the ijolites have atomic $\text{SLa+Ce} < \text{Na}$. A group of perovskite analyses from the northern complex melilite-calcite rocks are more highly enriched in REE than any of the other samples, and contain ~8% latrappite. Most of the melilitolite and ijolite samples contain ~4.5% loparite and 1-2% latrappite, while the pyroxenite analyses are almost pure perovskite. One exception to this is an

Figure 5.7 Perovskite compositions from the Turiy rocks. A plot of Na vs [La + Ce] shows the degree of loparite ($\text{NaREETi}_2\text{O}_6$) component in the perovskite analyses. The perovskites from the northern melilite-carbonatite samples are additionally enriched in REE. In general, the melilitolite perovskite analyses contain $[\text{La} + \text{Ce}] > \text{Na}$, while the ijolite perovskite analyses contain $\text{Na} < [\text{La} + \text{Ce}]$. X and Y axis units = apfu.

analysis from the ijolite sample C.IJ.14, which contains 30% loparite, 8% lueshite (NaNbTiO_3) and only 48% perovskite (CaTiO_3).

That much of the magnetite in the melilitolites is surrounded by rims of perovskite (and/or garnet) has profound implications for the oxygen fugacity of the melt. The oxygen fugacity of a melt is usually buffered by the iron-oxide minerals that are either present, forming, or being dissolved, but the “removal” of significant quantities of the magnetite from the magma will drastically alter the oxygen fugacity and hence pathways for the remainder of the crystallization process. Dunworth & Wilson (submitted) have shown that the saturation of a melt in Fe^{3+} is critical to the timing of melilite formation in extrusive melilitites, as previously suggested by Mysen et al. (1976). Mysen et al. (op cit.) showed that the presence of Fe^{3+} within a melt enhanced the solubility of CO_2 and hence artificially enhanced the polymerization of the melt. Saturation in Fe and precipitation of iron oxide phases caused the release of CO_2 , depolymerization of the melt, and precipitation of melilite, in the experimental system studied. Given the abundance of magnetite in the melilitolites in Turiy (especially compared to the ijolites), Fe^{3+} saturation may also be a key factor in melilite crystallization in intrusive rocks. It is also possible that the growth of the perovskite rims may take place under sub-solidus conditions, in which case the effective “removal” of magnetite from the system due to its enclosure by perovskite, will have minimal effect on the oxygen fugacity of the rest of the rock.

5.3b Apatite

Apatite is another ubiquitous mineral at Turiy. This calcium phosphate mineral represents the major source and/or repository for the last of the major network-forming cations (P) to be discussed. Like many of the other minerals already discussed, apatite compositions at Turiy show systematic variations depending on the rock type in which

they occur. Le Bas & Handley (1979) described similar findings from East African carbonatites. Unfortunately, Ca and P dominate the cation sites, and substitutions, such as they are, occur only on a minor scale. Typical substitutions include Sr for Ca, and coupled substitutions such as $\text{Ca} + \text{P} \leftrightarrow \text{REE} + \text{Si}$. MnO and total REE patterns, commonly used to characterise carbonatite apatite chemistry (Hogarth, 1989), were not obtained for the Turiy apatites, the former because MnO was not included in the “phosphate” element list in the microprobe analyses, and the latter because it involves mineral separation which was not carried out for trace element analyses. Due to the inability to analyse accurately for total REE content using the electron microprobe, a plot of Sr vs Si (Figure 5.8) was used to demonstrate variations in apatite composition among the different rock types.

Although Sr-Si variations might be considered a somewhat unorthodox method to “categorise” the apatite compositions, the dependence of the apatite compositions on their host rock-type is clearly seen in Figure 5.8. The low concentrations of these two elements in apatite imply that the results are prone to error. However, the analyses plotted in Figure 5.8 were carried out over a period of two years, and the results remain consistent.

The pyroxenite apatites have low Sr and significant Si contents, as shown on Figure 5.8. The apatite analyses from the melilitolites are clustered in a sub-horizontal grouping on the Si-rich part of the diagram, lying between 0.06-0.15 apfu Si and < 0.05 apfu Sr. The majority of the apatite analyses from the phoscorites also have $\text{Si} > \text{Sr}$.

The olivine melteigite apatites contain roughly equal Si and Sr contents (~ 0.07 apfu of each), while most of the ijolite analyses have $\text{Sr} > \text{Si}$, particularly those from the more fractionated samples such as C.IJ.14 and E.123.106. The ijolite apatite analyses with $\text{Si} > \text{Sr}$ were obtained from sample C.ND.500 (ijolite-melilitolite contact) and sample WKM (west coast micro-ijolite). The magnetite analyses from both rocks are also

Figure 5.8 Apatite analyses from the Turi rocks. The plot provides a means of dividing the analyses into groups according to the rock type from which they are derived. See main text for further explanation. Sample TD 1 = late-stage fluorite-apatite-carbonate vein, and TD 2 = late-stage aegirine-augite - apatite - ankerite vein, both from the south coast. X and Y axis units = apfu.

very similar, and sample WKM contains magnetite-perovskite-garnet overgrowths typical of those found in the melilitolites. Mention should also be made of the strontian fluorapatites from the ijolite-fenite contact sample ij-fen, which contain virtually no Si, but up to 0.42 apfu Sr (6.6 - 8.1 wt.% SrO).

The large cluster of carbonatite and phoscorite analyses between 0 - 0.5 apfu Si and 0 - 0.5 apfu Sr come from samples from the central complex and show little overlap with analyses from the silicate rocks. The carbonatite analyses with high Si contents (0.12 - 0.22 apfu) and low Sr contents, are derived from the northern complex carbonatites which either contain melilite or are associated with melilite-bearing rocks. The apatite analyses from the late-stage carbonatite sample C.54.x plot around [0, 0.12]. At Turiy the Sr contents in apatite are consistently lower than the Sr contents of co-existing calcite, a feature that was also shown by many African carbonatites (Dawson et al., 1996).

The turjite data points are derived from two of the turjite samples. The data from the TU 120A apatites fall close to those from ijolite samples C.IJ.14 and E.123.106 (two possible candidates for a turjite precursor) while sample TQ, which contains quenched relict laths of melilite, has apatite compositions that plot close to those from the melilitolites. The late-stage vein samples [TD 1 and TD 2] both contain apatite with $Sr > Si$. The apatite from sample TD 2 is richer in Sr than apatite from sample TD 1, and the data points from both samples fall just above the main cluster of central complex carbonatite analyses in Figure 5.8.

On the basis of the systematic variations in apatite composition based on rock type, rough divisions are shown in Figure 5.8 to create three different fields: the first, defining a field of low Sr and low Si contents, contains the central complex carbonatite-phoscorite analyses. The remaining area is divided using a line of slope $Si = Sr + 0.04$ apfu. This line meets the carbonatite-phoscorite field at approximately [0.06, 0.03]. The

large Sr-rich field lying to the left of the $\text{Si} = \text{Sr} + 0.04$ apfu line, contains apatite data from the olivine melteigite, ijolites, turjite sample TU 120A, and the late-stage vein samples TD 1 and TD 2. The Si-rich field to the right of the $\text{Si} = \text{Sr} + 0.04$ apfu line contains data from the melilitolites, pyroxenites, northern complex carbonatites, and turjite sample TQ. The use of these fields is effective in helping to divide the apatite data according to rock type, although there are a few data points which do not fall in the field associated with most of the data from their particular rock type. For example, there are a few ijolite analyses in the Si-rich field with the melilitolites, and two carbonatite data points in the ijolite field.

Sample N.64.19 also contains trace quantities of a phosphate mineral which may be Ce-britholite $(\text{Ce,Ca})_5(\text{SiO}_4, \text{PO}_4)_3(\text{OH,F})$. The low wt.% oxide total in sample N.64.19 (Appendix Ji) is probably due to the large abundance of REE present which were not included in the analysis table. Additionally, any error in the REE calibration will affect the totals since the analyses contain >50 wt.% REE oxides. The analysis in sample N.64.19 is also rich in Sr. An unidentified Ca-Si-P mineral containing F and Cl was found in sample N.63.4.

In summary, these results show that a plot of Sr vs Si is effective in separating the apatite analyses from the different rock types at Turiy into three distinct fields. Those from the ijolites, olivine melteigite, late-stage vein samples TD 1 and TD 2, and the fenites have $\text{Sr} > \text{Si}$, those from the melilitolites, phoscorites and pyroxenites have $\text{Si} > \text{Sr}$. The carbonatite analyses can be divided into two main groups as the central complex apatite analyses have distinctly lower Si contents than the apatite analyses from the northern complex carbonatites. Both groups of carbonatites have low Sr contents.

5.3c Carbonates

A significant number of carbonate minerals are found in Turiy. The dominant carbonate mineral is calcite which is present in minor or trace quantities in many of the silicate rocks, and is a major constituent of both the phoscorites and carbonatites. Calcite is also the dominant carbonate mineral in the Kovdor massif (Zaitsev & Bell, 1996) which also contains melilitolites and lies further to the north-west along the Kandalakha Graben.

Solid solution substitutions in the carbonates from Turiy are rare, although Mg and Sr are the two most common elements which substitute for Ca, along with Mn and Fe. MnO contents are < 0.25 wt.% in all calcite analyses from Turiy, < 0.4 wt.% in all dolomite analyses and up to 2.2 wt.% in ankerite from the ankerite-fluorite vein sample TD 1. Zaitsev (1996), in a study of carbonatites from the Khibina massif, showed that Sr enrichment in calcite (~1wt.% SrO) is characteristic of primary calcite, and low Sr contents characterized secondary calcite. No systematic variation has been found between calcite composition and rock type at Turiy.

Calcite is found as an accessory mineral in a significant number of the silicate rocks (pyroxenites, melilitolites, ijolites), particularly in those samples which contain an abundance of other volatile-rich minerals (e.g. turjaite sample TUR I). Calcite is abundant in the groundmass of turjite sample TU 120 A. Small grains of strontianite occur along grain boundaries in many of the samples, but most of these were too small to analyse. Dolomite has only been found in one silicate rock, turjaite sample WKT.

In the carbonatites, calcite is the dominant carbonate mineral and occurs in a variety of textures discussed in Chapter 4. Dolomite occurs as small rhombs in the phoscorites and forms veins in some of the magnetite-apatite-^{iv}Fe³⁺-mica-rich carbonatites from the central complex. It is extremely rare in melilite-bearing carbonatites from the northern complex. The late carbonatites and carbonate veins are

more enriched in dolomite (sample C.DC contains approximately equal amounts of dolomite and calcite), while ankerite has only been found in the late-stage veins on the south coast [samples TD 1 and TD 2]. Strontianite forms small crystals along grain boundaries in the carbonatites and phoscorites. Most of the calcite from the carbonatites and phoscorites contains consistent Sr concentrations of between 4000 - 6200 ppm, and Nd concentrations between 20 - 132 ppm, as determined by isotope dilution analysis

Trace quantities of Ba and REE carbonates have been found in some of the carbonatites from Turiy, but in insufficient quantities for the purposes of accurate analysis and identification.

5.3d Sulphides

A detailed analysis of sulphide minerals has not formed part of this study. Sulphide phases are found in almost all rock types, consisting primarily of pyrrhotite and chalcopyrite, and occurring with greatest abundance in the youngest carbonatites and the phoscorites. One notable occurrence is found in one of the phoscorites where what appears to be an immiscible “droplet” of pyrite occurs between two large magnetite grains, enclosed within calcite and associated with euhedral pyroxene. Kogarko et al. (1995) also reported immiscible sulphide-carbonate textures in a mantle xenolith from the Canary Islands.

An unidentified sulphide phase, $[\text{Fe}_3\text{CuS}_4]$, was found in melteigite sample C.AG.122, and djerfisherite $[\text{K}_6\text{Na}(\text{Fe,Cu})_{24}\text{S}_{26}\text{Cl}]$ was found in melilitolite sample TUR II. Both of these samples also contain pyrrhotite. Kapustin (1980) reports pyrrhotite as the most common sulphide in alkaline carbonatite complexes. Galena appears to be absent from the carbonatites at Turiy.

5.4 Other minerals

5.4a Cancrinite

Cancrinite is an important constituent (16%) in melilitolite sample TUR I, and also occurs in trace quantities in the olivine melteigite, several ijolites from the central massif and most melilitolites. There is little variation in composition, as seen in Appendix Jvi, except for the Na contents. Na contents cannot be measured using a standard, focused electron beam on hydrous phases such as cancrinite, and the Na results, other than those from sample TUR I which were obtained using a low beam current and a rastered electron beam, should be ignored. The cancrinite in sample TUR I was also analysed for SO₃ content (5.3 - 7.5 wt.%) and Cl content (0.1 - 2.0 wt.%).

5.4b Banalsite & Stronalsite

Banalsite [Na₂BaAl₄Si₄O₁₆] occurs along grain boundaries in melteigite sample C.AG.122 and ijolite sample C.IJ.14, whilst stronalsite [Na₂SrAl₄Si₄O₁₆], identified for the first time in Japan in 1987, was found in trace quantities along grain boundaries in the eastern melilitolite sample E.133.30.

5.4c Titanite

Titanite [CaTiSiO₅] is limited in its occurrence at Turiy. It is only found in pyroxenite, olivine melteigite, turjite and ijolite, including the ijolite-fenite contact zones. Of particular note, is its absence from the melilitolites. It is fine-grained, only found in trace quantities, and shows little compositional variation or substitution, although ZrO₂ can be present in minor quantities. It appears that the SiO₂ activity in the melilitolites is seemingly too low (or Ti activity not high enough) to allow the formation of titanite, and instead, abundant Fe-Ti oxides and perovskite are formed. In the case of the the ijolites most of the Ti is preferentially partitioned into garnet (and pyroxene), preventing

significant titanite formation.

5.4d *Zircon, zirsinalite, calzirtite and baddelyite*

The four zirconium minerals found in the Turiy rocks were zircon, zirsinalite, calzirtite and baddelyite. Less than half a dozen zircon grains have been found in all the samples analyzed (mostly from the carbonatites), and all are less than 20µm diameter. One carbonatite contained a 5µm zircon inclusion in apatite and two of the youngest central massif calcite carbonatites contain trace quantities of the minerals zirsinalite $[\text{Na}_6(\text{Ca}, \text{Mn}, \text{Fe})\text{ZrSi}_6\text{O}_{18}]$ and calzirtite $[\text{Ca}_2\text{Zr}_5\text{Ti}_2\text{O}_{16}]$ along grain boundaries. Baddelyite and zircon analyses also contained Hf in measurable quantities.

Baddelyite $[\text{ZrO}_2]$ was found in the texturally heterogeneous carbonatite C.6.15, where it occurs as extremely thin rims around apatite grains in a band of magnetite and apatite, corroborating the intimate association between apatite and baddelyite reported by Kapustin (1980). Another similar, slightly coarser-grained apatite band within the same thin-section of sample C.6.15, contained no traces of baddelyite.

The alternating bands of coarse and fine-grained magnetite-apatite-calcite sequences in sample C.6.15 (Figure 4.10b), and the presence of baddelyite in only one band suggests that either the baddelyite crystallized around the apatite prior to its incorporation into the sample, or that a Zr-rich fluid percolated through the rock sometime after formation, and precipitated baddelyite around one vein of apatite crystals. This sample is a good example of the veining and mixing of small carbonatite melts and the heterogeneity of this massif.

5.4e *Sulphates*

Trace quantities of the sulphate minerals barite and celestine have been found along grain boundaries in a few silicate and carbonate samples, and may be of secondary

origin. No analyses were obtained from these minerals, although semiquantitative EDS scans suggest that there may be some solid solution variations between the two end-members.

5.4f *Fluorite*

Fluorite has been found in two samples from Turiy. It is an essential constituent of TD 1 - a carbonate-fluorite-apatite-quartz vein on the south coast, and it is also found in the fenite part of the ijolite-fenite contact sample. Its composition appears to be pure CaF_2 with no detectable cation substitutions. In carbonatites, fluorite is a well-known constituent of pneumato-hydrothermal veins and typically occurs in association with ankerite $\pm$ quartz (Kapustin, 1980).

CHAPTER 6

ISOTOPE GEOCHEMISTRY

6.1 Introduction

Three long-lived radioactive decay schemes (Rb-Sr, Sm-Nd and Ar-Ar) and two stable isotope systems (C and O) have been used to investigate the isotope geochemistry of selected mineral separates and whole rock samples from Turiy. The aim of this part of the investigation was to provide answers to the following questions:

1. are the individual mineral phases within each sample in isotopic equilibrium?
2. how old are the rocks within the massif?
3. were all the magmatic rocks originally derived from the mantle?
4. did all the igneous rocks have the same source?
5. did Turiy exhibit closed-system fractionation, or open-system behaviour?
6. if open-system conditions prevailed, then how can the isotope data help constrain the effect of the open-system behaviour on the evolution of the massif?

It was important to establish whether all the magmatic rocks in the massif are the same age, or whether Turiy was affected by several phases of intrusive activity. A study by Kramm et al. (1993) showed that a considerable number of Palaeozoic alkaline intrusions within the Kola Peninsula, including Turiy, Kovdor, Afrikanda, Lovozero and Khibina, all contained alkaline and ultramafic igneous rocks emplaced between 360 - 380 Ma. Little work has been done on intra-massif age relationships in Kola, apart from the recent study of the Kovdor phoscorites and carbonatites by Zaitsev & Bell (1996). This present study provides a more detailed look at the initial isotopic signatures of both silicate and non-silicate Turiy rocks, and provides some interesting new information regarding the magma sources and age relationships of the different rock types.

Samples used for this isotope study were mostly mineral separates. Calcite, apatite and phlogopite were analyzed from various carbonatites and phoscorites, and

apatite, amphibole, pyroxene, garnet, nepheline and perovskite were separated from silicate samples from Turiy, for both Sr and Nd isotopic analysis. Al mineral separates were subjected to acid-leaching using dilute HCl, prior to analysis. Unfortunately, no galena was found for common Pb isotope work. Phlogopite separates were used for Ar-Ar laser ablation isotope determinations at the University of Toronto. A few carefully chosen, texturally homogeneous whole-rock samples were also analyzed for bulk-rock Sr-Nd isotopic signatures.

Analytical methods for sample selection, rock crushing, mineral separation, chemical preparation and mass spectrometry can be found in Appendix E.

6.2 Radiogenic isotope systems

6.2a Rb-Sr

The decay of ^{87}Rb to ^{87}Sr by β -decay, with a half-life of 48.8×10^9 y, is a radioactive decay scheme which has been utilized for many years for geological studies (e.g. Faure, 1986). Initial $^{87}\text{Sr}/^{86}\text{Sr}$ ratios have proven very useful as natural tracers in monitoring both mantle and crustal histories (e.g. Faure, 1986), and have been applied extensively to carbonatites and related rocks (e.g. Bell & Blenkinsop, 1987a; Nelson et al., 1988). Rb is a highly incompatible element in most minerals and is a member of the LILE group of elements (Chapter 7.3b). It most commonly substitutes for K in minerals such as feldspar, nepheline, amphibole and phlogopite. Sr, a member of the alkaline earth group of elements, is considered to be a more compatible element than Rb, and substitutes readily for Ca and Ba in a large variety of minerals. Within the mantle, Sr is found most commonly as a trace element in clinopyroxene, and less commonly in amphibole and carbonate phases. It is also found in fluid inclusions in olivine and pyroxene and within grain boundary phosphate melts (Rosenbaum, 1996; Rosenbaum et al., 1997).

Given that Ca-bearing minerals are some of the most essential minerals in a carbonatite intrusion, it is hardly surprising that Sr has extremely high abundances in these igneous rocks. The Sr content of the Turiy carbonatites range from 3,300 - 22,400 ppm and the Rb contents range from 2 - 32 ppm. Many of the mineral separates utilized in this study (e.g. apatite, calcite, perovskite, garnet) contain no Rb, and thus provide initial $^{87}\text{Sr}/^{86}\text{Sr}$ ratios which represent the ratio at the time of magma formation. Phlogopite, on the other hand, contains varied Rb and Sr concentrations, hence providing a wide range of $^{87}\text{Rb}/^{86}\text{Sr}$ ratios. Few of the minerals listed in section 6.1 contain “intermediate” Rb/Sr concentration ratios. Nepheline, the most suitable candidate for intermediate Rb/Sr values in these rocks, has not been found in pyroxenites, carbonatites or phoscorites at Turiy, and can be difficult to separate from the melilitolites and ijolites due to its similarity in appearance to apatite and calcite in some samples and its association with alteration products or late-stage, volatile-rich silicates (e.g. cancrinite, pectolite).

6.2b Sm-Nd

The radioactive decay of ^{147}Sm to ^{143}Nd by alpha-particle emission, with a half-life of 1.06×10^{10} y, has been utilized during the last 10-15 years for geological age determinations and isotope tracer studies. It is generally believed that the Sm-Nd system is less subject to post-magmatic alteration processes, given that both daughter and parent are rare earth elements (REE) and have similar chemical properties. However, Andersen (1984) pointed out that Sm-Nd isochrons have yet to be obtained from carbonatites, a feature which is also true for the Turiy massif.

There are no minerals which are known to concentrate Nd to the preference of Sm, which means that there is no method for directly measuring the $^{143}\text{Nd}/^{144}\text{Nd}$ initial ratio (unlike the Rb-free apatite and calcite whose $^{87}\text{Sr}/^{86}\text{Sr}$ signatures represent initial values). Therefore all Sm-Nd data must be back-corrected to obtain the initial

$^{143}\text{Nd}/^{144}\text{Nd}$ ratio.

Tables 6.A, 6.B and 6.C list the data obtained in this study by isotope dilution. The initial ratio data for pyroxenite sample C.AG.347 have been back-corrected to 588 Ma and the data for all other samples have been back-corrected to 373 Ma for reasons which will be explained in a later section. Calculation of the estimated present-day $^{143}\text{Nd}/^{144}\text{Nd}$ ratios from isotope dilution analyses are similar to the $^{143}\text{Nd}/^{144}\text{Nd}$ ratios measured on unspiked samples, although, on average, they are not as similar as those obtained from similar measurements in the Rb-Sr system. The isotopic values measured from the unspiked samples are quoted in Tables 6.A, 6.B and 6.C. Sr isotopic compositions were normalized to $^{88}\text{Sr}/^{86}\text{Sr} = 8.375$ relative to simultaneous measurements of international standard NBS 987. Average NBS 987 value = 0.710267 (n = 25). Nd isotopic compositions were normalized to $^{146}\text{Nd}/^{144}\text{Nd} = 0.7219$ relative to simultaneous measurements of the La Jolla standard. Average La Jolla value = 0.511849 (n = 21).

6.2c Ar-Ar

A number of phlogopite separates from a variety of melilitolite, phoscorite and carbonatite samples, with grains > 1mm diameter, were subjected to neutron-radiation for one month at McMaster University, and are subsequently being analyzed for ^{40}Ar - ^{39}Ar isotope ratios using laser ablation at the University of Toronto. The data are not available for inclusion into this thesis.

6.3 Isochron relationships

The ISOPLOT program (Ludwig, 1988) was used to calculate the ages and initial ratios for the Rb-Sr (and Sm-Nd) isochrons. Figure 6.1 shows a Rb-Sr isochron plot for Turiy, using all mineral separate and whole rock data, and data for mineral separates and

Table 6.A Rb-Sr decay scheme isotope and concentration data: Turly Massif

Sample	Rock type	Mineral	Rb ppm	Sr ppm	Rb/Sr	$^{87}\text{Rb}/^{86}\text{Sr}$	$^{87}\text{Sr}/^{86}\text{Sr}$ measured	$^{87}\text{Sr}/^{86}\text{Sr}$ initial	epsilon Sr (T)
C.AG.347	pyroxenite	amphibole	9.64	738.7	1.3E-02	0.0377	0.70375	0.70343	-5.87
C.AG.347	pyroxenite	perovskite	0.12	1655	7.2E-05	0.0002	0.70343	0.70343	-5.87
C.AG.347	pyroxenite	pyroxene	1.60	336.9	4.7E-03	0.0137	0.70355	0.70343	-5.84
C.AG.347	pyroxenite	calcite	0.42	3997	1.1E-04	0.0003	0.70346	0.70346	-5.46
C.90.29	pyroxenite	pyroxene	0.81	316.8	2.6E-03	0.0074	0.70358	0.70365	-5.64
C.90.29	pyroxenite	whole rock	2.91	538	5.4E-03	0.0157	0.70376	0.70367	-5.84
C.5.118	melilitolite	phlogopite	354.2	371.3	9.5E-01	2.763	0.71926	0.70455	6.57
C.190.105	melilitolite	whole rock	38.96	2058	1.9E-02	0.0548	0.70391	0.70362	-6.56
N.64.18	melilitolite	calcite	0.15	5519	2.7E-05	0.0001	0.70436	0.70436	3.97
E.133.30	melilitolite	apatite	15.95	6990	2.3E-03	0.0066	0.70378	0.70374	-4.83
WKT	melilitolite	apatite	66.70	1574	4.2E-02	0.1226	0.70454	0.70386	-3.24
TUR I	melilitolite	phlogopite	363.3	676.0	5.4E-01	1.556	0.71207	0.70379	-4.23
TUR I (2)	melilitolite	apatite	0.13	6163	2.1E-05	0.0001	0.70357	0.70357	-7.34
TUR II	melilitolite	apatite	6.20	3123	2.0E-03	0.0057	0.70399	0.70407	-0.22
TUR II	melilitolite	phlogopite	359.7	584.6	6.2E-01	1.781	0.71337	0.70389	-2.81
TUR II	melilitolite	perovskite	0.38	2687	1.4E-04	0.0004	0.70392	0.70391	-2.43
Sample	Rock type	Mineral	Rb ppm	Sr ppm	Rb/Sr	$^{87}\text{Rb}/^{86}\text{Sr}$	$^{87}\text{Sr}/^{86}\text{Sr}$ measured	$^{87}\text{Sr}/^{86}\text{Sr}$ initial	epsilon Sr (T)
C.AG.48A	olivine melteigite	apatite	8.77	8354	1.0E-03	0.0030	0.70418	0.70416	1.10
C.AG.48A	olivine melteigite	calcite	17.30	7204	2.4E-03	0.0069	0.70415	0.70411	0.43
C.AG.122	melteigite	whole rock	38.76	1118	3.5E-02	0.1003	0.70428	0.70374	-4.86
C.ND.500	ijolite	phlogopite	242.7	274.8	8.8E-01	2.558	0.71716	0.70356	-7.46
C.ND.500	ijolite	pyroxene	0.53	441.1	1.2E-03	0.0035	0.70346	0.70317	-13.0
C.ND.500	ijolite	garnet	0.82	221.8	3.7E-03	0.0107	0.70358	0.70344	-9.10
C.II.14	ijolite	whole rock	71.85	1213	5.9E-02	0.1714	0.70479	0.70388	-2.88
C.II.14	ijolite	apatite	22.61	16374	1.4E-03	0.0040	0.70388	0.70385	-3.27
C.90.y	ijolite	apatite	0.79	4965	1.6E-04	0.0005	0.70382	0.70382	-3.78
C.90.y	ijolite	calcite	0.26	5599	4.6E-05	0.0001	0.70335	0.70335	-10.4
WKM	ijolite	nepheline	140.7	226.2	6.2E-01	1.801	0.71400	0.70441	4.61
TU 119	turjite	apatite	7.78	9398	8.3E-04	0.0024	0.70410	0.70448	5.66
TU 119	turjite	whole rock	57.84	2832	2.0E-02	0.0591	0.70466	0.70435	3.76

Table 6.A Rb-Sr decay scheme isotope and concentration data: Turiy Massif (cont.)

Sample	Rock type	Mineral	Rb ppm	Sr ppm	Rb/Sr	$^{87}\text{Rb}/^{86}\text{Sr}$	$^{87}\text{Sr}/^{86}\text{Sr}$ measured	$^{87}\text{Sr}/^{86}\text{Sr}$ initial	epsilon Sr (T)
NC.153.82	fenite	whole rock	90.52	873.2	1.0E-01	0.3000	0.70859	0.70699	52.9
TD 2	aeg-ap vein	pyroxene	8.61	248.0	3.5E-02	0.1004	0.70371	0.70398	-1.47
TD 2	aeg-ap vein	apatite	6.97	10112	6.9E-04	0.0020	0.70400	0.70399	-1.32
C.K.7	phoscorite	phlogopite	165.2	8.66	1.9E+01	56.76	0.99628	0.69403	-143
C.K.7	phoscorite	phlogopite	115.2	6.10	1.9E+01	56.17	n.a.	n.a.	n.a.
C.K.7	phoscorite	pyroxene	0.38	119.9	3.2E-03	0.0092	0.70322	0.70318	-13.0
C.K.7	phoscorite	calcite	2.27	4538	5.0E-04	0.0014	0.70317	0.70316	-13.3
C.3.93	phoscorite	phlogopite	213.0	6.22	3.4E+01	104.4	1.25299	0.69693	-102
C.3.93	phoscorite	phlogopite	220.2	6.33	3.5E+01	106.0	n.a.	n.a.	n.a.
C.3.93	phoscorite	calcite	0.03	6083	5.1E-06	0.0000	0.70306	0.70306	-14.8
C.18.30	phoscorite	phlogopite	148.0	9.22	1.6E+01	47.52	0.94450	0.69148	-179
C.18.30	phoscorite	pyroxene	0.46	2015	2.3E-04	0.0007	0.70317	0.70316	-13.3
C.18.30	phoscorite	apatite	0.04	2427	1.6E-05	0.0000	0.70313	0.70313	-13.8
C.19.163	phoscorite	phlogopite	280.8	27.40	1.0E+01	30.10	0.86506	0.70476	9.43
C.19.163	phoscorite	apatite	0.48	3249	1.5E-04	0.0004	0.70320	0.70319	-12.9
C.19.163	phoscorite	calcite	0.11	5678	2.0E-05	0.0001	0.70312	0.70312	-13.9
C.49.40	carbonatite	calcite	0.20	5871	3.5E-05	0.0001	0.70340	0.70340	-9.69
C.49.40	carbonatite	apatite	0.01	4642	2.4E-06	0.0000	0.70349	0.70349	-8.41
C.54.x	carbonatite	phlogopite	209.4	657.0	3.2E-01	0.9224	0.70889	0.70397	-1.56
C.54.x	carbonatite	calcite	0.81	3301	2.5E-04	0.0007	0.70341	0.70340	-9.68
C.54.x	carbonatite	calcite	0.58	4066	1.4E-04	0.0004	0.70341	0.70340	-9.66
C.XC.5	carbonatite	apatite	0.52	2715	1.9E-04	0.0006	0.70407	0.70363	-6.42
C.XC.5	carbonatite	calcite	0.14	3987	3.5E-05	0.0001	0.70348	0.70347	-8.66
C.DC	dolomite	dolomite	0.07	4906	1.4E-05	0.0000	0.70334	0.70334	-10.6
N.63.4	carbonatite	calcite	1.35	6194	2.2E-04	0.0006	0.70436	0.70435	3.83
N.64.18c	carbonatite	calcite	0.52	5806	8.9E-05	0.0003	0.70429	0.70428	2.85
N.64.19	carbonatite	calcite	0.32	5755	5.5E-05	0.0002	0.70430	0.70430	3.00
S.W.58	carb. dyke	whole rock	1.33	18774	7.1E-05	0.0002	0.70382	0.70381	-3.84

Isotope results normalized to $^{86}\text{Sr}/^{88}\text{Sr} = 0.1194$. NBS 987 Sr standard = 0.71026 ± 1 (uncertainty quoted at the 2 sigma level, average of 24 analyses). Rb/Sr ratios accurate to 1%.

Data from sample C.AG.347 corrected using an age of 588 Ma. All other data corrected using an age of 373 Ma.

Table 6.B Sm-Nd decay scheme isotope and concentration data: Turiy Massif

Sample	Rock type	Mineral	Sm ppm	Nd ppm	$^{147}\text{Sm}/^{144}\text{Nd}$	$^{143}\text{Nd}/^{144}\text{Nd}$ measured	$^{143}\text{Nd}/^{144}\text{Nd}$ initial	epsilon Nd (T)
C.AG.347	pyroxenite	amphibole	23.71	180.84	0.131	0.51264	0.51234	7.85
C.AG.347	pyroxenite	perovskite	101.4	529.84	0.191	0.51267	0.51223	5.70
C.AG.347	pyroxenite	pyroxene	6.22	32.666	0.190	0.51277	0.51232	7.56
C.AG.347	pyroxenite	calcite	0.76	3.064	0.248	0.51266	0.51208	2.89
C.90.29	pyroxenite	pyroxene	4.40	23.88	0.184	0.51257	0.51236	3.36
C.90.29	pyroxenite	whole rock	7.07	40.824	0.173	0.51260	0.51234	2.92
C.5.118	melilitolite	phlogopite	0.47	3.7698	0.126	0.51242	0.51224	0.91
C.190.105	melilitolite	whole rock	20.46	133.41	0.153	0.51265	0.51242	4.47
N.64.18	melilitolite	calcite	3.48	26.674	0.131	0.51255	0.51236	3.31
E.133.30	melilitolite	apatite	226.8	1761.1	0.129	0.51227	0.51208	-2.22
WKT	melilitolite	apatite	0.99	6.0599	0.163	0.51268	0.51244	4.84
TUR I	melilitolite	phlogopite	2.68	21.792	0.123	0.51259	0.51241	4.31
TUR I (2)	melilitolite	apatite	211.4	1599	0.130	0.51260	0.51240	4.12
TUR II	melilitolite	apatite	13.23	68.823	0.192	0.51260	0.51232	2.51
TUR II	melilitolite	phlogopite	0.47	3.609	0.129	0.51260	0.51240	4.14
TUR II	melilitolite	perovskite	657.7	4976.2	0.132	0.51244	0.51232	2.50
Sample	Rock type	Mineral	Sm ppm	Nd ppm	$^{147}\text{Sm}/^{144}\text{Nd}$	$^{143}\text{Nd}/^{144}\text{Nd}$ measured	$^{143}\text{Nd}/^{144}\text{Nd}$ initial	epsilon Nd (T)
C.AG.48A	olivine melteigite	apatite	103.4	789.9	0.131	0.51241	0.51221	0.43
C.AG.48A	olivine melteigite	calcite	10.13	74.511	0.136	0.51242	0.51221	0.34
C.AG.122	melteigite	whole rock	28.68	141.55	0.203	0.51260	0.51229	2.03
C.ND.500	ijolite	phlogopite	0.72	3.12	0.231	0.51304	0.51270	9.98
C.ND.500	ijolite	pyroxene	0.45	1.4396	0.313	0.51255	0.51209	-2.07
C.ND.500	ijolite	garnet	31.50	118.0	0.267	0.51279	0.51240	4.04
C.II.14	ijolite	whole rock	17.20	96.325	0.179	0.51247	0.51220	0.21
C.II.14	ijolite	apatite	52.04	520.43	0.100	0.51241	0.51226	1.29
C.90.y	ijolite	apatite	57.26	362.3	0.158	0.51268	0.51244	4.92
C.90.y	ijolite	calcite	6.55	39.734	0.165	0.51263	0.51239	3.83
WKM	ijolite	nepheline	1.48	18.063	0.082	0.51241	0.51229	1.97
TU 119	turjite	apatite	84.21	588.38	0.143	0.51256	0.51235	3.05
TU 119	turjite	whole rock	24.19	103.08	0.235	0.51263	0.51228	1.71

Table 6.B Sm-Nd decay scheme isotope and concentration data: Turiy Massif (cont.)

Sample	Rock type	Mineral	Sm ppm	Nd ppm	Sm/Nd	$^{147}\text{Sm}/^{144}\text{Nd}$	$^{143}\text{Nd}/^{144}\text{Nd}$ measured	$^{143}\text{Nd}/^{144}\text{Nd}$ initial	epsilon Nd (T)
NC.153.82	fenite	whole rock	4.98	28.40	0.175	0.1060	0.51190	0.51164	6.19
TD 2	aeg-ap vein	pyroxene	0.50	1.73	0.289	0.1746	0.51268	0.51225	1.12
TD 2	aeg-ap vein	apatite	45.76	337.3	0.136	0.0820	0.51249	0.51229	1.95
C.K.7	phoscorite	phlogopite	0.03	0.15	0.206	0.1244	0.51644	0.51613	76.94
C.K.7	phoscorite	pyroxene	0.90	3.79	0.238	0.1439	0.51233	0.51197	-4.56
C.K.7	phoscorite	calcite	5.49	36.81	0.149	0.0901	0.51267	0.51245	4.79
C.3.93	phoscorite	phlogopite	0.03	0.21	0.140	0.0848	0.51218	0.51188	-6.11
C.3.93	phoscorite	calcite	10.41	72.18	0.144	0.0872	0.51272	0.51251	5.85
C.18.30	phoscorite	phlogopite	0.04	0.15	0.274	0.1658	0.51424	0.51446	43.9
C.18.30	phoscorite	pyroxene	22.90	113.28	0.202	0.1222	0.51273	0.51243	4.29
C.18.30	phoscorite	apatite	35.83	207.31	0.173	0.1045	0.51290	0.51264	8.42
C.19.163	phoscorite	phlogopite	0.07	0.37	0.198	0.1200	0.51185	0.51156	-12.34
C.19.163	phoscorite	apatite	54.01	341.0	0.158	0.0958	0.51268	0.51245	4.71
C.19.163	phoscorite	calcite	9.30	61.28	0.152	0.0918	0.51273	0.51250	5.78
C.49.40	carbonatite	calcite	6.24	37.35	0.167	0.1009	0.51259	0.51234	3.00
C.49.40	carbonatite	apatite	22.51	132.86	0.169	0.1024	0.51257	0.51232	2.42
C.54.x	carbonatite	phlogopite	2.34	32.414	0.072	0.0437	0.51229	0.51218	-0.16
C.54.x	carbonatite	calcite	1.41	25.14	0.056	0.0338	0.51245	0.51237	3.44
C.XC.5	carbonatite	apatite	13.87	74.917	0.185	0.1119	0.51251	0.51224	0.88
C.XC.5	carbonatite	calcite	3.59	19.833	0.181	0.1094	0.51256	0.51229	1.99
C.DC	dolomite	dolomite	3.99	25.50	0.156	0.0946	0.51247	0.51224	0.94
N.63.4	carbonatite	calcite	12.61	102.2	0.123	0.0746	0.51269	0.51250	6.08
N.64.18c	carbonatite	calcite	16.50	132.5	0.125	0.0753	0.51260	0.51241	4.33
N.64.19	carbonatite	calcite	10.71	84.95	0.126	0.0762	0.51269	0.51251	6.18
S.W58	carb. dyke	whole rock	61.45	523.0	0.117	0.0710	0.51263	0.51245	5.10

Isotope results normalized to $^{146}\text{Nd}/^{144}\text{Nd} = 0.7219$. La Jolla Nd standard = 0.51185 ± 2 (uncertainty quoted at the 2 sigma level, average of 24 analyses). Sm/Nd ratios accurate to 1%.

Data from sample C.AG.347 corrected using an age of 588 Ma. All other data corrected using an age of 373 Ma.

Table 6.C Terskii Coast isotope data

Rb-Sr decay scheme isotope and concentration data

Sample	Rock type	Mineral	Rb ppm	Sr ppm	Rb/Sr	$^{87}\text{Rb}/^{86}\text{Sr}$	$^{87}\text{Sr}/^{86}\text{Sr}$ measured	$^{87}\text{Sr}/^{86}\text{Sr}$ initial	epsilon Sr (T)
K77A	kimberlite	whole rock	96.95	1580	6.1E-02	0.1775	0.70493	0.70399	-0.98
K77A	kimberlite	apatite	0.22	17162	1.3E-05	0.0000	0.70391	0.70391	-2.10
K77A	kimberlite	calcite	1.71	5764	3.0E-04	0.0009	0.70397	0.70397	-1.24
K77B	kimberlite	whole rock	107.0	1710	6.3E-02	0.1810	0.70484	0.70388	-2.50
K77B	kimberlite	apatite	0.16	17657	9.3E-06	0.0000	0.70385	0.70385	-2.96
K77C	kimberlite	whole rock	187.9	3297	5.7E-02	0.1648	0.70490	0.70402	-0.54
K77C	kimberlite	apatite	0.19	24932	7.5E-06	0.0000	0.70389	0.70389	-2.40
OM1	olivine melilitite	W.R.	22.67	441.0	5.1E-02	0.1487	0.70629	0.70550	20.5
OM29	olivine melilitite	W.R.	34.62	1310	2.6E-02	0.0764	0.70635	0.70594	26.7
OM29	olivine melilitite	"calcite"	20.42	11128	1.8E-03	0.0053	0.70522	0.70519	16.1
ON165	olivine nephelinite	whole rock	47.26	469.6	1.0E-01	0.0053	0.70972	0.70969	80.0
ON165	olivine nephelinite	apatite	9.58	292.1	3.3E-02	0.0949	0.70633	0.70582	25.1
ON165	olivine nephelinite	"calcite"	40.46	161.6	2.5E-01	0.7261	0.72703	0.72317	271

Sm-Nd decay scheme isotope and concentration data

Sample	Rock type	Mineral	Sm ppm	Nd ppm	Sm/Nd	$^{147}\text{Sm}/^{144}\text{Nd}$	$^{143}\text{Nd}/^{144}\text{Nd}$ measured	$^{143}\text{Nd}/^{144}\text{Nd}$ initial	epsilon Nd (T)
K77A	kimberlite	whole rock	9.73	66.46	0.146	0.0885	0.51231	0.51210	-1.23
K77A	kimberlite	apatite	23.49	90.74	0.259	0.1565	0.51246	0.51207	-1.67
K77A	kimberlite	calcite	26.26	85.97	0.305	0.1847	0.51257	0.51212	-0.77
K77B	kimberlite	whole rock	16.08	140.4	0.115	0.0692	0.51241	0.51224	1.49
K77B	kimberlite	apatite	33.73	217.5	0.155	0.0938	0.51235	0.51212	-0.81
K77C	kimberlite	whole rock	5.11	23.24	0.220	0.1329	0.51251	0.51218	0.44
K77C	kimberlite	apatite	25.44	185.8	0.137	0.0828	0.51236	0.51215	-0.11
OM1	olivine melilitite	whole rock	9.03	63.56	0.142	0.0858	0.51261	0.51240	4.62
OM29	olivine melilitite	whole rock	8.36	54.63	0.153	0.0926	0.51263	0.51240	4.75
OM29	olivine melilitite	"calcite"	4.97	28.00	0.177	0.1073	0.51239	0.51213	-0.62
ON165	olivine nephelinite	whole rock	8.49	53.42	0.159	0.0961	0.51244	0.51221	0.91
ON165	olivine nephelinite	apatite	65.66	161.1	0.408	0.2465	0.51306	0.51246	5.81
ON165	olivine nephelinite	"calcite"	28.58	166.1	0.172	0.1040	0.51243	0.51217	0.22

Isotope results normalized to $^{86}\text{Sr} = 0.1194$ and $^{146}\text{Nd}/^{144}\text{Nd} = 0.7219$.

Rb/Sr and Sm/Nd ratios accurate to 1%.

NBS 987 Sr standard = 0.71026 ± 1 ; La Jolla Nd standard = 0.51185 ± 2 (uncertainty quoted at the 2 sigma level. Average of 24 analyses)

Data from sample C.A.G.347 corrected using an age of 588 Ma. All other data corrected using an age of 373 Ma.

whole rock samples from the Terskii Coast pipes, kindly supplied by A. Arzamastsev. Figure 6.1a illustrates, in particular, the high $^{87}\text{Rb}/^{86}\text{Sr}$ ratio of the phlogopites from the phoscorite samples relative to the whole Figure 6.1b shows the very low $^{87}\text{Rb}/^{86}\text{Sr}$ ratios for the remainder of the samples. Figure 6.4 shows the Sm-Nd data plotted by both rock type (Figure 6.2a) and mineral type (including whole rock data) (Figure 6.2b).

Table 6.D lists the isochron ages obtained using the Rb-Sr decay scheme, along with their corresponding MSWD values and calculated initial $^{87}\text{Sr}/^{86}\text{Sr}$ ratios with 2σ error, for a variety of individual samples. However, it can be seen that of the calculated ages listed in Table 6.D the isochron age from the pyroxenite sample C.AG.347 (588 ± 72 Ma) is quite different from the remaining ages which are closer to the age obtained by Kramm et al. (1993) from a central complex ijolite from Turiy (373.1 ± 5.6 Ma). The isochron age obtained from phoscorite sample C.K.7 (363 ± 3.5 Ma) is similar to the date obtained by Kramm et al. (1993).

Thus, the first phase of magmatism at Turiy is reflected in the data from the pyroxenite sample C.AG.347. This suggests that at least some of the central complex rocks may be Precambrian in age. The initial values for the Turiy pyroxenite sample C.AG.347 are $^{87}\text{Sr}/^{86}\text{Sr} = 0.70343$ and $^{143}\text{Nd}/^{144}\text{Nd} = 0.51223 - 0.51234$, based on amphibole, perovskite and pyroxene mineral separates. Other carbonatite massifs of similar age include the Alnö and Fen complexes which also lie within the Baltic Shield and were intruded at 584 and 580 Ma respectively (Andersen, pers. comm., 1997), while the Quigussaq and Sarartok carbonatites in north-east Greenland were intruded ~600 Ma ago (Woolley, 1989). The Iron Hill melilitolite-carbonatite complex in Colorado was intruded at ~550 Ma (Armbrustmacher & Futa, 1985), Hall et al. (1995) report the presence of a carbonatite associated with nepheline syenite in Antarctica, dated at 531 Ma, and the Wet Mountains carbonatite in Colorado and a number of carbonatites in eastern Russia are also of similar age (Woolley, 1989). These data indicate that

Figure 6.1a Rb-Sr isochron plot for the Turiy samples listed in Table 6.A showing the wide range in $^{87}\text{Rb}/^{86}\text{Sr}$ ratios. The wide variation is due to the data from the phlogopite separated from various phoscorite samples. TD 2 = aegirine-augite -apatite vein from the south coast.

The graph also includes a point (12.5, 0.76790) from a phlogopite separate from “kandagubite”, the type-locality rock from the nearby Kandaguba Massif, located just south of Kandalaksha. On the basis of field relationships, this massif is believed to be similar in age to the Turiy Massif.

Figure 6.1b An expansion of the low- $^{87}\text{Rb}/^{86}\text{Sr}$ end of the Rb-Sr isochron plotted in Figure 6.3a. A couple of the melilitolite phlogopite points, just visible at $\sim ^{87}\text{Rb}/^{86}\text{Sr} = 3$ in Figure 6.3a are off the x-axis scale used in Figure 6.1b. This figure also shows the cluster of very depleted data from the phoscorites around $^{87}\text{Sr}/^{86}\text{Sr} = 0.70315$. TD 2 = aegirine-augite - apatite vein from the south coast.

carbonatite magmatism was not uncommon during this period in Earth history, and, furthermore, that such activity was not uncommon within the Baltic Shield at this time.

Most of the Turiy samples reflect Devonian magmatic activity. Samples from which Devonian age Rb-Sr isochrons were obtained (Table 6.D) include phoscorite sample C.K.7 (363 ± 3.5 Ma, MSWD = 0.26), and melilitolite sample TUR II (372 ± 53 Ma, (MSWD = 4.4). Most of the ages listed in Table 6.D are within error of the age obtained by Kramm et al. (1993) (373.1 ± 5.6 Ma, MSWD = 1.6) from a 5-point Rb-Sr isochron from a central complex ijolite.

The initial $^{87}\text{Sr}/^{86}\text{Sr}$ - $^{143}\text{Nd}/^{144}\text{Nd}$ ratios of the phoscorites are the lowest of any seen at Turiy (0.703150, 0.51290). Initial $^{87}\text{Sr}/^{86}\text{Sr}$ ratios from carbonatites from Kovdor, Sebyavr and Vuoriyarvi were measured at 0.70310-0.70357 by Demaiffe et al. (1997), and fall within the range shown by the Turiy phoscorites and carbonatites. The $\epsilon_{\text{Nd}}(\text{T})$ values of the carbonatite data of Demaiffe et al. (1997) and this study, are also similar.

Alteration of the original isotopic signature of the magma can result from contamination during ascent, magma mixing, subsequent alteration of the resulting rocks by metasomatic fluids, or post-magmatic hydrothermal alteration, and weathering. The isotope data from the whole-rock samples are similar to the results from acid-leached mineral separates from the same samples, and indicate that sample contamination by post-magmatic alteration is not significant in Turiy. Post-magmatic alteration commonly produces high $^{87}\text{Sr}/^{86}\text{Sr}$ ratios in the contaminated rocks. The central complex pyroxenites, which have some of the lowest Sr concentrations, and are therefore most likely to show the influence of crustal fluids, also have some of the lowest initial $^{87}\text{Sr}/^{86}\text{Sr}$ ratios from the silicate rocks, suggesting that the silicate magmas within the central complex have not been subjected to post-magmatic alteration. Similar arguments against high-level crustal contamination were put forward by Huang et al. (1995) to explain the relationship between the pyroxenite and carbonatite isotope data at Jacupiranga. High

Table 6.D   Rb-Sr isochron ages, initial isotopic ratios and MSWD errors

Sample	Rock type	Minerals	Age (Ma)	$^{87}\text{Sr}/^{86}\text{Sr}_i$	MSWD
C.AG.347	pyroxenite	amph, prv, pyx	588 ± 72	0.70343 ± 2	0.02
C.ND.500	ijolite	gt, phlog, pyx	376 ± 100	0.70348 ± 52	16.7
TUR II	melilitolite	ap, phlog, pvk	372 ± 53	0.70394 ± 27	4.4
C.K.7	phoscorite	cal, phlog, pyx	363 ± 3.5	0.70317 ± 2	0.26
C.19.163	phoscorite	ap, cal, phlog	378 ± 94	0.70316 ± 50	15.4
C.18.30	phoscorite	cal, phlog, pyx	372 ± 44	0.70315 ± 23	3.42
K77A	kimberlite	ap, cal, W.R.	393 ± 250	0.70394 ± 35	7.75

Errors quoted as 2σ values.

$^{87}\text{Sr}/^{86}\text{Sr}$ ratios (0.70428-0.70436) are, however, seen within a number of samples from the satellite complexes at Turiy. The minerals with the highest $^{87}\text{Sr}/^{86}\text{Sr}$ ratios from these samples include Sr-rich apatite (Table 6.A), and thus these isotopic signatures are unlikely to have been affected by post-magmatic, low-Sr crustal fluids. This suggests that any possible crustal contamination of the magmas which formed the satellite complexes by crustal material must have occurred during emplacement, rather than as a result of post-magmatic alteration.

There are several experimental constraints which affect a study of this type. Because of the fine grain size, inclusions, and intergrown minerals in some samples, it was sometimes difficult to obtain pure mineral separates. Emphasis was placed on obtaining initial ratio values from a large number of samples, rather than obtaining isochrons from a few samples. The calculated initial ratios of some of the samples (Tables 6.A and 6.B), indicate that there has been a certain amount of open-system behaviour of the Rb-Sr and Sm-Nd systems since, or during crystallization., although the samples show few signs of alteration.

Despite the difficulties outlined above, one of the most interesting findings from this study are the variations shown by initial $^{87}\text{Sr}/^{86}\text{Sr}$ from the Rb-poor, Sr-rich mineral separates, apatite and calcite (Table 6.A). These data indicate that the Turiy rocks were not derived from one magma source which evolved in a magmatically closed system. This feature of the data set will be examined in further detail in the next section.

The Terskii Coast kimberlite, melilitite and nephelinite samples, kindly provided by Andrei Arzamastsev from the Kola Science Centre in Apatity, were difficult to date. The “calcite” separates from the Terskii melilitites, as listed in Table 6.C, appeared to be secondary in origin. However, one of the kimberlite samples [K77A] provided a Rb-Sr age of 393 ± 230 Ma (MSWD 7.75), but the errors are much too large to evaluate the temporal relationship between the kimberlite and the Turiy massif. K-Ar dates obtained

by Kalinkin et al. (1993) range from 337 to 384 Ma. Kalinkin et al. (1993) also suggested that the other pipes and dykes of olivine melilitite and nephelinite along the Terskii Coast may be older than the kimberlites based on K-Ar ages obtained from Terskii Coast olivine melilitites (465 ± 35 Ma from amphibole and 620 ± 50 Ma from phlogopite). Two-point isochrons (not shown) from some of the melilitite and nephelinite samples listed in Table 6.C provide some support for the findings of Kalinkin et al. (1993).

No Sm-Nd isochrons were obtained from the Turiy Massif or Terskii Coast samples analyzed in this study. Andersen (1984) pointed out that carbonatites rarely provide Sm-Nd isochrons, and this has also held true in the Turiy intrusion. Plotting mineral separate and whole rock data for individual samples produces lines with negative slopes in $^{147}\text{Sm}/^{144}\text{Nd}$ vs $^{143}\text{Nd}/^{144}\text{Nd}$ space, commonly with low MSWD values. It is interesting to note that samples from which Rb-Sr isochrons have been obtained (e.g. pyroxenite sample C.AG.347) do not produce Sm-Nd isochrons, indicating closed-system behaviour of the Rb-Sr system, but open-system behaviour of the Sm-Nd system during crystallization. Although difficult to interpret, one possible explanation involves REE fractionation by halogen-rich fluids during crystallization, which assumes that the halogens combine preferentially with the REE rather than Rb and Sr. In a volatile-rich environment, F complexes with the HREE, while Cl complexes with the LREE (Haas et al., 1995), but this behaviour is also pH-dependent. Little is known about the pH conditions of fluids co-existing with natural magmas, including carbonatites (Pyle, pers. comm., 1997), and further discussion of this model would be premature.

Beard et al. (1996), in a ^{40}K - ^{39}Ar and ^{40}Ar - ^{39}Ar study of the Kandalaksha dyke series, located near the town of Kandalaksha, obtained (K-Ar) ages of 360 ± 16 Ma and 368 ± 15 Ma from lamprophyre dykes. Calciocarbonatite and hornblende-cumulate xenoliths contained within them yielded (Ar-Ar) ages of 386 ± 1 Ma and 395.6 ± 4.4 Ma.

Figure 6.2a Sm-Nd isochron plot for the Turiy samples listed in Table 6.B plotted according to rock type. No consistent isochrons are shown by this plot, and data from mineral separates from individual samples commonly exhibit negative correlations (not shown). TD 2 = late-stage aegirine-apatite-ankerite vein from the south coast.

Figure 6.2b Sm-Nd isochron plot for the Turiy samples listed in Table 6.B plotted according to mineral type. Again, no consistent variations are seen. TD 2 = late-stage aegirine-apatite-ankerite vein from the south coast.

The Kandalaksha lamprophyre dykes thus appear to be synchronous with the main phase of Devonian magmatism at Turiy. The xenoliths were probably derived from a deep-seated intrusion beneath Kandalaksha which was sampled by the erupting lamprophyres (Beard et al., 1996). The difference in ages between the xenoliths and the host lamprophyres near Kandalaksha suggests there have been repetitive phases of magmatism within the Kandalaksha Graben during the Devonian, corroborated by the three separate phases of dyke activity (pre-, syn-, post-massif emplacement) represented on the south coast of the Turiy Peninsula (Bulakh & Ivanikov, 1984).

6.4 Isotopic components and relationships in Sr-Nd space

It is important to define, at this stage, what is meant by open-system and closed-system behaviour, as used in this study. During melt production at source, closed-system behaviour is taken to represent the derivation of magma with a uniform isotopic signature, which has subsequently been intruded into the crust and cooled. Such behaviour would produce identical isotopic ratios in all samples. Open-system behaviour at source includes mixing of melts from different mantle sources. Closed-system behaviour during intrusion refers to the isolated emplacement of each batch of magma, without interaction and contamination by or mixing with surrounding material, be it crustal or magmatic in origin. Closed-system behaviour during crystallization refers to the achievement of isotopic equilibrium on a hand-specimen scale, so that isochrons can be obtained from mineral separates within one sample. Open-system behaviour during crystallization can be reflected by scattering of initial isotope ratios of mineral separates taken from within a single hand-specimen, outside the limits of analytical uncertainty.

Tables 6.A, 6.B and 6.C show that the initial $^{87}\text{Sr}/^{86}\text{Sr}$ and $^{143}\text{Nd}/^{144}\text{Nd}$ ratios of the Turiy samples vary widely, and that in some of the samples the Rb-Sr and Sm-Nd systems do not behave as closed chemical systems during crystallization. The data in

these tables were back-corrected using an age of 373 Ma, except for pyroxenite sample C.AG.347, the data from which was back-corrected using an age of 588 Ma. Figure 6.3 shows a plot of initial $^{87}\text{Sr}/^{86}\text{Sr}$ vs $^{143}\text{Nd}/^{144}\text{Nd}$ for the Turiy samples, back-corrected to 373 Ma, the age obtained by Kramm et al. (1993) from a Turiy ijolite. Although it is likely that the Devonian magmatism at Turiy represents more than one phase of intrusive activity, the data have been plotted on one Sr-Nd diagram (i.e. back-corrected to a single age) for simplicity. Due to the older, Precambrian age obtained from sample C.AG.347, the data from this sample are not included in this diagram.

Most of the points fall in the top left-hand quadrant of a $^{87}\text{Sr}/^{86}\text{Sr}$ - $^{143}\text{Nd}/^{144}\text{Nd}$ anti-correlation plot (Figure 6.3a). All data that fall in this quadrant reflect derivation from a “depleted mantle”, i.e. mantle which has a time-integrated high Sm/Nd - low Rb/Sr ratio (Faure, 1986). The most interesting finding to emerge from Figure 6.3a is that samples from each rock type fall, on the whole, within a distinct part of the Sr-Nd anti-correlation plot, indicating some degree of closed-system behaviour during intrusion. The cluster of the phoscorite data, which have the most depleted $^{87}\text{Sr}/^{86}\text{Sr}$ and $^{143}\text{Nd}/^{144}\text{Nd}$ signatures of all the samples, is one such example.

There is little evidence for intra-sample isotopic disequilibrium in the Rb-Sr system (Tables 7.A, 7.B and 7.C), although there is evidence for Rb-Sr isotopic disequilibrium between different samples of the same rock type. The Sm-Nd data do show intra-sample isotopic disequilibrium, as reflected by the lack of Sm-Nd isochrons. Similar isotopic disequilibrium in the Sr and Nd systems has been previously reported from the Jacupiranga complex (Huang et al. 1995), the Gardiner Complex (Nielsen & Buchardt, 1985) and in extrusive nephelinites (Simonetti & Bell, 1993).

The most depleted samples, in terms of Sr and Nd isotope compositions at Turiy are the phoscorites, which have a very distinct $^{87}\text{Sr}/^{86}\text{Sr}$ - $^{143}\text{Nd}/^{144}\text{Nd}$ signature which are unlike that obtained from any other samples analyzed so far. Slightly less depleted are

Figure 6.3a Anti-correlation plot for $^{143}\text{Nd}/^{144}\text{Nd}$ - $^{87}\text{Sr}/^{86}\text{Sr}$. All data back-corrected to 373 Ma. The pyroxenite sample C.AG.347 has been omitted due to the 588 Ma age obtained for this sample. Sample plotted according to rock type. TD 2 = late-stage aegirine-apatite-ankerite vein from the south coast. Terskii = Terskii Coast kimberlite samples. BE-CHUR = bulk Earth - CHUR intersection at 373 Ma.

Figure 6.3b Identical plot to Figure 6.3a, but plotted according to mineral type. Comparison of these two graphs enables the minerals from each rock type to be identified.

the central massif carbonatites which have lower $^{143}\text{Nd}/^{144}\text{Nd}$ ratios and higher $^{87}\text{Sr}/^{86}\text{Sr}$ ratios. Most of the data from the melilitolites and the northern massif carbonatites have more enriched $^{87}\text{Sr}/^{86}\text{Sr}$ ratios than the phoscorites but have initial $^{143}\text{Nd}/^{144}\text{Nd}$ ratios that are similar. The majority of the melilitolite and carbonatite data plot to the right of the cluster of phoscorite data. Some of the data points from this latter group fall in the upper right-hand quadrant of the anti-correlation diagram.

The most depleted melilitolite data point (closest to the phoscorites) comes from the central complex whole rock sample C.190.105. Data from the melteigite sample C.AG.122, ijolite sample C.IJ.14 and olivine melteigite [C.AG.48A], in particular, fall closest to the bulk-earth - CHUR intersection, near to the data from the Terskii Coast kimberlite and related pipes. Sr-Nd isotopic data from the Terskii kimberlite pipes plotted in Beard et al. (1996) are much more enriched in $^{87}\text{Sr}/^{86}\text{Sr}$. The spread of the Turiy data across the depleted quadrant suggests that more than one mantle ($\pm$ crustal) source has been involved in the genesis of the Turiy magmas. None of the data from Turiy fall in the lower right-hand quadrant, which represents "enriched mantle" (i.e. mantle with a time-integrated low Sm/Nd, high Rb/Sr ratio).

Most mantle-derived igneous rocks lie along the "Mantle Array" (Faure, 1986), a broad band trending from the upper left-hand quadrant into the upper left-hand corner of the enriched quadrant, although with an increase in new data, this general trend is not quite as clear as was previously thought. In modern-day environments, the most depleted signatures are found in mid-ocean-ridge-basalts (MORBs) while ocean-island-basalts (OIBs) and subduction-related magmas have ratios that trend from the depleted quadrant into the lower right-hand quadrant of the Sr-Nd plot (e.g. Wilson, 1989). Average upper continental crust signatures lie within the enriched quadrant.

In the early 1980s it was noted that data from a variety of young carbonatites (< 30 Ma) and related silicate rocks from the East Africa Rift defined a mixing line just

below and to the left of the mantle array, which was named as the “East Africa Carbonatite Line” (EACL) by Bell et al. (1982). This linear array is very similar to the Lo-Nd array defined by Hart et al. (1986). Following this approach, Kramm (1993) defined a Devonian “Kola Carbonatite Line” (KCL) using data from a variety of Palaeozoic massifs within the Kola Peninsula, including Khibina, Kovdor, Lovozero, Sokli, Oznaya Varaka and Turiy. This line has a much steeper slope than the EACL. Of note on the original KCL epsilon plot from Kramm (1993) is the presence of two carbonatites from Khibina plotting at (-2, 4) which is similar to data from the northern complex carbonatites, and western complex melilitolites seen at Turiy. The carbonatite data point from Turiy, used by Kramm (1993) to help define the KCL, overlaps with the phoscorite field defined in this study (Figure 6.4).

The Turiy data from this study are compared to the KCL in Figure 6.4. The main point that emerges from this diagram is that many of the central complex silicate rocks lie along or close to the KCL, while the central complex carbonatites and phoscorites fall into distinct groups along or just below the KCL and most of the data from samples taken from the satellite complexes and south coast dykes lie above the KCL. Data from the two pyroxenite samples are also included in Figure 6.4, although the data from pyroxenite sample C.AG.347 have been corrected to an age of 588 Ma, based on the Rb-Sr isochron obtained from this sample. Data from sample C.90.29 have been corrected using an age of 373 Ma. The resulting $\epsilon_{Sr}(T)$ values from both pyroxenite samples are almost identical and range from -5.46 to -5.87 (Table 6.A) although they have quite different $\epsilon_{Nd}(T)$ values (2.89 - 7.85) (Table 6.B). The intra-sample Sm-Nd disequilibrium seen between the pyroxenite mineral separates allows little interpretation to be placed on the REE characteristics of their source.

One of the interpretations of the EACL proposed by Bell & Blenkinsop (1987b) and Bell (1997) is that it reflects a mixing line between two mantle-derived components

Figure 6.4 Epsilon diagram (373 Ma) to compare the Turiy data with the Kola Carbonatite Line (KCL) of Kramm (1993). Samples plotted according to rock type. TD 2 = aegirine-augite - apatite vein from the south coast. Terskii = Terskii Coast kimberlite data. The points which have epsilon Sr > -5 and which fall to the right of the KCL are from the satellite complexes. All other data lying either on, or below the KCL, come from central complex samples. There are three data points from the melilitolites which do not follow this rule. The point at (6.57, 0.91) is a phlogopite separate from central complex sample C.5.118. The points at (-4.83, -2.22) and (-7.34, 4.12) are from the eastern and western complexes respectively.

defined by Zindler & Hart (1986) based on data from (modern) oceanic basalts. One is believed to be relatively depleted, and corresponds to the HIMU mantle component, normally associated with mantle plume isotopic signatures (Bell & Simonetti, 1996). The other is an enriched component which corresponds to EM I, one of two known enriched mantle components (the other being EM II). Although there is some controversy about where such components are housed in the mantle, HIMU is normally believed to be asthenospheric in origin, while EM I can be either asthenospheric or lithospheric in origin (Zindler & Hart, 1986), although Hart (1988) now favours a model in which HIMU and EM I are located at the core-mantle boundary.

Kramm (1993) argued that one end-member of the KCL is similar to EM I, and lies in the enriched quadrant. The other end-member lies in the depleted quadrant, and Kramm (1993) argued for a similarity between it and the source of the Grenville and Superior Province carbonatites in Canada (Bell & Blenkinsop, 1987a). The critical feature of the EACL and KCL on the Sr-Nd isotope plot is that both mixing relationships are represented by straight lines, which means that the end-member sources must have had the same Sr/Nd concentration ratios. This has implications for the mineralogy of the mantle sources. While it is difficult to measure the trace element contents of some of the minerals within the mantle source rocks (e.g. phosphate and carbonate phases) that may control the Rb, Sr and REE contents of mantle-derived melts, approximate estimates may be gleaned from similar mineral phases found in the Turiy samples. Figure 6.5 shows the Sr/Nd concentration ratios of minerals and whole-rock samples from Turiy, analyzed in this study, plotted against the Sr concentration. Of the minerals with very high Sr contents, the Sr/Nd ratios decrease systematically in the order calcite (~ 100) > apatite >> perovskite (~ 1). The minerals with lower Sr contents generally have a wide range of Sr/Nd ratios, including pyroxene with ratios ranging from $\sim 500 - 10$. The whole rock data points plot close to or within the apatite and calcite analyses, suggesting that the Sr-

Figure 6.5 Sr/Nd vs Sr (ppm) for mineral separates and whole-rock data from the Turiy samples. The mineral separate data was obtained by isotope dilution, and the whole rock data were obtained by ICP-MS. Whole rock samples are represented by: "pyx" = pyroxenite [C.90.29], melteigite [C.AG.122] and melilitite dyke [TU 107]; mel = melilitolite [C.5.118, C.190.105, WKT, TUR II]; carb = carbonatite [C.49.40, C.TL.344, C.DC, S.W58]; OMIT = olivine melteigite [C.AG.347], ijolite [C.IJ.14], turjite [TU 119] and fenite [NC.153.82]. A systematic variation in Sr content is seen both between the mineral phases and between the rock types. Note the wide variation in Sr content of the phlogopite samples and the tight cluster of calcite data, taken from a wide variety of silicate and carbonate rock types.

Nd -rich phases, which occur in trace quantities in many of the samples, can control the Sr and Nd concentrations in the magmas. It is therefore possible that the accessory phases within the mantle such as carbonate (Kogarko et al., 1995) and phosphates (Rosenbaum et al., 1997) play a dominant role in controlling the Sr/Nd ratio of the sources.

Correlating the Turiy data to the two hypothetical end-members proposed by Kramm (1993) for the KCL, suggests that the phoscorites, and possibly the carbonatites and melilitolites were derived primarily from a depleted mantle source, while the olivine melteigite and ijolites (and possibly the Terskii kimberlites) were more strongly influenced by an enriched mantle component.

6.5 Enriched Mantle or crustal contamination?

One of the most important questions about the genesis of the Turiy rocks, in terms of their Sr and Nd isotope ratios, is whether the spread of data shown in Figure 6.4 reflects mantle source variations, or whether it reflects open-system behaviour. The data from the satellite complexes (including the northern complex carbonatites) and the south coast dykes, plot to the right of the KCL, and even within the upper right-hand quadrant. On the basis of the geochemical behaviour of Rb, Sr, Nd and Sm, data derived from different sources should not fall in either the upper right-hand quadrant or the lower left-hand quadrant. Data points which do fall within the upper right-hand quadrant are normally considered to reflect the effects of crustal contamination, or post-magmatic groundwater interaction and alteration. In general, the Nd isotope ratios remain unchanged but the $^{87}\text{Sr}/^{86}\text{Sr}$ ratios are taken to higher values during post-magmatic contamination processes. Given the very high concentration of Sr and Nd within the Turiy rocks and mineral separates (Tables 6.A and 6.B), it is unlikely that sub-solidus alteration (either by post-magmatic crustal fluids or seawater) could have affected the

isotopic signature of these rocks. If the enriched $^{87}\text{Sr}/^{86}\text{Sr}$ values are primary features of these magmas, then they must be attributed to an, as yet unidentified, mantle source component. The characteristics of the enriched mantle component EM II, defined on the basis of data from modern oceanic basalts (Zinder & Hart, 1986), are not sufficiently well known for us to constrain the behaviour of this component in the Palaeozoic mantle. The presence of such a component in the magmas which formed the satellite complexes at Turiy cannot be ruled out. However, this implies that the melt production and plumbing systems of the central and satellite complexes must have been completely separate, which seems somewhat unlikely. If the enriched Sr values are due to crustal contamination, then the contamination must have occurred during emplacement - either due to wall-rock assimilation in situ, or mixing with lower crustal material during magma ascent.

Most of the samples which fall to the right of the KCL, and in the upper right-hand quadrant are from the satellite complexes, including mineral separates and whole rock data from four carbonatite samples, and it is suggested that they have been contaminated by material with more enriched isotopic signatures during magma emplacement. In order to understand the isotopic constraints on this process, and given that many of the data points which lie to the right of the KCL are derived from melilitolites and melilite-bearing carbonatites from the satellite complexes, isotopic mixing calculations were carried out between a whole-rock central complex melilitolite sample [C.190.105] which has the most depleted isotopic ratios of all the melilitolite analyses and appears to be uncontaminated, and a variety of crustal end-members, to see if the isotopic signatures of the contaminated melilitolites and related carbonatites from the northern complex could be produced. The results of these calculations, involving mixing with the Turiy fenite sample NC.153.82 (FEN), charnockite (CHARN) (de Mulder et al., 1986), average upper crust (AUC) (McCulloch & Chappell, 1982) and average Precambrian lower crustal granulite (ALC) (DePaolo et al., 1982), are shown in

Figure 6.6a. The mixing lines show that the slope of the melilitolite-fenite mixing line is virtually identical to that involving the melilitolite-charnockite end-member. This might be expected if the crustal end member was charnockitic in composition. Figure 6.6b shows the mixing lines superimposed on the Turiy data, and demonstrates that mixing between the central complex melilitolite [C.190.105] and charnockite (or fenite) does indeed produce a mixing curve that passes through many of the points to the right of the KCL and in the top right-hand quadrant. The percentage of charnockite that is needed to contaminate the satellite complexes to produce the highest $^{87}\text{Sr}/^{86}\text{Sr}$ isotopic signatures from the silicate rocks from Turiy is 35%, which, in terms of bulk mixing, is unrealistic. The melilitolite - AUC line suggests that smaller quantities of “average upper crust” (10%) would also be sufficient to produce the most contaminated isotopic signatures by bulk mixing, but the crustal rocks through which the magmas have intruded do not include material typical of “average upper crust”, according to the cross-section shown in Figure 2.2. The melilitolite-ALC mixing line in Figure 6.6b passes through the ijolite data and the data from the Terskii Coast kimberlites. It is highly unlikely that the kimberlites were formed by 35% assimilation of “average lower crust” into the melilitolite magmas and thus the KCL trend is almost certainly mantle-derived.

An alternative to simple binary mixing to produce the points in the upper right-hand quadrant is the AFC model (assimilation and fractional crystallization) of DePaolo (1981). The parameters needed to quantify this model include the composition of the initial magma, the composition of the host rock, the composition of the fractionating phases, and the partition coefficients of the relevant elements (Sr and Nd) between the fractionating phases and the residual melt. Parameters such as the rate of assimilation and crystallization also have to be estimated. At Turiy, the composition of the original host rock is not accurately known, the composition of the original magma is not well constrained, neither are the compositions of the fractionating phases (hidden beneath the

Figure 6.6a Mixing calculations between a central complex melilitolite whole rock isotopic analysis [C.190.105], and each of four crustal components. ALC = average lower crust (Weaver & Tarney, 1984); AUC = average upper crust (McCulloch & Chappell, 1982); charnockite = Zairean charnockite (de Mulder et al., 1986); fenite = fenite sample NC.153.82. The symbols on each mixing curve represent incremental changes of 5% in the weight ratio of melilitolite : crustal rock involved in the calculation. Large circles = 100% crustal end-member. Note the overlap between the melilitolite-fenite and melilitolite-charnockite mixing lines.

Figure 6.6b Superposition of the Turiy isotope data with some of the mixing lines in Figure 6.6a above. The Turiy data is plotted in grey using identical symbols as in Figure 6.4, while the points on the “melilitolite-crustal component” lines use identical symbols to the diagram above (TD 2 = late-stage aegirine-augite - apatite vein on the south coast; Terskii = Terskii Coast kimberlite data; CHARN = charnockite, FEN = fenite; references as above). The overlap between the melilitolite-charnockite and melilitolite-fenite lines is again clearly seen, but note that the percentage values marked by the points closest to epsilon Sr = 5 are different for the two lines. The Turiy data from the satellite complexes which lies to the right of the KCL (not shown) and within the upper right-hand quadrant may be explained by contamination of the magmas by charnockite or upper crustal material. The fact that the melilitolite-ALC line also passes through most of the ijolite, olivine melteigite and Terskii kimberlite data, is not considered significant, given that these data points also lie close to the Mantle Array and the KCL (not shown).

current level of the intrusion). As a result, estimates of the Sr and Nd partition coefficients between two unknown starting materials (primary melt and host rock) and an unknown fractionating phase (or phases), cannot be accurately estimated. Additionally, the method of proposed assimilation is hard to constrain, as the textures seen in the fenite samples suggest that fluid migration and preferential dissolution of particular minerals may have been the primary means of contamination. Due to the lack of constraints on the necessary data, any AFC calculation would be subject to gross oversimplification for these rocks, and thus numerical AFC modelling was not carried out.

The fact that the samples from the satellite complexes show isotopic evidence of crustal contamination by charnockitic material, and the charnockites surrounding the intrusion show mineralogical and isotopic evidence of fenitization by magmatic fluids, implies that the magma-crust interaction has been a two-way process at Turiy. The apparent contamination of the satellite complex rocks with the charnockite host rocks might initially involve minor amounts of bulk-rock assimilation, coupled with more significant quantities of trace element-rich fluids that circulated between the Proterozoic charnockite and the ascending magma. Such a process would be driven by the heat from the intrusion, during the ascent and emplacement of the magmas which formed the satellite complexes. It seems unlikely that pure bulk assimilation occurred. Most of the highly “contaminated” data points in Figure 6.4 and 6.6.b represent mineral separates with high trace element concentrations. The migration of fluids rich in trace elements during equilibration between the massif and the host rock during magma ascent through the crust and emplacement at current levels, is a far more realistic model than bulk mixing to explain the isotopic signatures than contamination of trace-element rich minerals by secondary fluids during or after magma emplacement and cooling. After emplacement, and during the cooling and crystallization of the magmas, magmatic fluids subsequently migrated out of the satellite complexes to fenitize the surrounding country

rocks. The crustal contamination model examined above has been constrained using a melilitolite as the magmatic end-member, although it is also possible that similar mixing effects occurred involving other rocks from the massif, such as the ijolites.

Woolley & Jones (1987), in a study of the Chilwa alkaline province, came across similar source-identification and contamination problems, and invoked metasomatism of the lower crust to create the geochemical signatures of the alkaline melts. Simonetti & Bell (1994) re-examined the model of Woolley and Jones using Sr-Nd-Pb isotopic data, and agreed that contamination by lower crustal material was indeed a possibility. An additional problem at Turiy is that the charnockites suggested as a possible end-member in the isotopic mixing process, are not believed to be representative of the thick Archean granulite-facies basement underlying almost the whole of the Kola Peninsula (Figure 2.2). The mixing relationships in Figure 6.6a suggest that mixing of the depleted melilitolite with average Archean, lower crustal granulites, does not produce the required enrichment in the $^{87}\text{Sr}/^{86}\text{Sr}$ signatures.

Almost all of the data from the central complex samples (including melilitolite sample C.190.105) falls on, or below the KCL, suggesting that contamination has not affected the central complex magmas, or if it has done so, it has occurred to a lesser extent than in the satellite complexes. Assuming that magmas from the satellite complexes unequivocally demonstrate contamination effects, the lack of apparent contamination in the central complex magmas requires explanation. If the derivation of the central complex involved an earlier, Precambrian phase of magmatism, as indicated by pyroxenite sample C.AG.347, then it is possible that a pre-existing magma conduit within the central complex was re-activated during the Devonian. The earlier magmas could have insulated the wall-rocks of the conduit, and thus prevented any later magmas from reacting with the crustal material. Although the data are limited, it does appear that the satellite complexes only contain Devonian magmas, and hence these melts must have

created their own plumbing system, and reacted with or assimilated wall-rock material in the process.

In summary, Turiy has maintained open-system conditions during its magmatic history, although there is also evidence of closed-system behaviour, reflected in the Rb-Sr isochrons, from the time of formation. It is likely that most of the open-system behaviour, such as heterogeneous magma mixing, occurred at source or in a magma chamber, and crystallization took place (in some cases) under closed-system conditions. Crustal material was incorporated into the melilitolites and other rock-types from the satellite complexes, probably as a result of fluid migration, prior to their intrusion to the current level of emplacement. The remainder of the samples, including the central complex melilitolites, fall within the depleted quadrant on a Sr-Nd initial ratio plot, overlapping the Kola Carbonatite Line (Kramm, 1993). The data from the silicate rocks at Turiy indicates that the KCL may represent mixing between melts from two mantle sources, whose products are represented at Turiy by the pyroxenites and melilitolites (depleted mantle source), and the olivine melteigite, ijolites and kimberlites (enriched source). The phoscorites may also have been derived from the more depleted of the two mantle sources, indicating that the pyroxenite-melilitolite group is not the most depleted derivative of this source seen at Turiy. Alternatively, the phoscorites (and central complex carbonatites) may have been derived from a separate source altogether. The isolation of the majority of the phoscorite and carbonatite data points in Figure 6.4, and the fact that they do not overlap with any of the data from the silicate rocks, suggests that they were not derived by liquid immiscibility or closed-system differentiation from any silicate magmas analyzed so far.

6.6 Stable isotopes: $\delta^{13}\text{C}_{\text{PDB}}$ and $\delta^{18}\text{O}_{\text{SMOW}}$

Carbon and oxygen stable isotope measurements were carried out on a suite of calcite separates from a variety of carbonatites and phoscorites, one calcite-melilitolite (N.64.19) and two whole-rock samples - one dolomite-calcite carbonatite and one fine-grained carbonatite dyke from the south coast (S.W58). The samples show a range of values which are listed in Table 6.E. $\delta^{18}\text{O}_{\text{SMOW}}$ values range from +7.5 to +16.7‰ with a tight clustering between +7.5 to +9.0‰. The $\delta^{13}\text{C}_{\text{PDB}}$ values range from -0.8 to -3.1‰ and are tightly clustered from -2.1 to -3.1‰. These results are displayed in Figures 6.7 and 6.8.

Figure 6.7 shows the Turiy results alone. There are two distinct clusters of points, one representing the central complex carbonatites, and the other the central complex phoscorites. The phoscorites ($\delta^{18}\text{O}_{\text{SMOW}}$ +7.9 to +8.2‰ and $\delta^{13}\text{C}_{\text{PDB}}$ -2.1 to -2.3‰) have similar $\delta^{18}\text{O}_{\text{SMOW}}$ and slightly higher $\delta^{13}\text{C}_{\text{PDB}}$ values than those of the central complex carbonatites ($\delta^{18}\text{O}_{\text{SMOW}}$ +7.5 to +9.0‰ and $\delta^{13}\text{C}_{\text{PDB}}$ -2.5 to -3.1‰). It can also be seen that there are four “outlier” points. These include sample N.63.4 which is a pyroxene-calcite sample associated with the melilite-calcite carbonatites of the northern complex and N.64.18m which is a calcite-melilitolite from a nearby location. The latter sample has the same $\delta^{18}\text{O}_{\text{SMOW}}$ value as that of the northern calcite-pyroxene sample N.63.4, but has a lower $\delta^{13}\text{C}_{\text{PDB}}$ value intermediate between those of the phoscorites and central carbonatites. The central complex dolomite-calcite-carbonatite sample C.DC shares a similar $\delta^{18}\text{O}_{\text{SMOW}}$ value to that of the phoscorites and central carbonatites, but has a more enriched $\delta^{13}\text{C}_{\text{PDB}}$ signature due to carbon isotopic fractionation between calcite and dolomite. S.W58, the carbonatite dyke on the southern coast, is the remaining outlier on this plot with the heaviest $\delta^{18}\text{O}_{\text{SMOW}}$ value of +16.7‰.

The stable isotope data were also compared with the $^{87}\text{Sr}/^{86}\text{Sr}$ and $^{143}\text{Nd}/^{144}\text{Nd}$ ratios, along with Sr concentrations, but no clear trends between the two suites of data

Table 6.E   **Stable Isotope Results**

Sample	rock type	mineral	delta ¹⁸ O _{SMOW}	delta ¹³ C _{PDB}
C.K.7	phoscorite	calcite	8.09	-2.15
C.3.93	phoscorite	calcite	8.17	-2.27
C.19.163	phoscorite	calcite	7.87	-2.14
C.XC.5	calcite carbonatite	calcite	7.60	-2.70
C.49.40	calcite carbonatite	calcite	7.62	-2.52
C.54.x	calcite carbonatite	calcite	8.96	-2.86
C.100.y	calcite carbonatite	calcite	7.56	-2.53
N.63.4	calcite carbonatite	calcite	11.22	-0.79
N.64.18c	calcite carbonatite	calcite	8.12	-2.87
N.64.19	calcite carbonatite	calcite	8.39	-3.07
N.187.100	calcite carbonatite	calcite	8.61	-2.58
S.W58	calcite carbonatite dyke	whole rock	16.58	-1.78
C.DC	dolomite-calcite carbonatite	whole rock	8.33	-1.01
N.64.18	calcite-rich melilitolite	calcite	11.28	-2.31

Figure 6.7 Stable isotope data from the Turiy carbonatites and phoscorites, as listed in Table 6.F. Data points from carbonatites lying outside the main carbonatite cluster are labelled individually.

Figure 6.8 Stable isotope $\delta^{13}\text{C}$ - $\delta^{18}\text{O}$ plot for calcite separates from Turiy carbonatites and phoscorites. Data from Kovdor (Zaitsev & Bell, 1995), Khibina (Zaitsev, 1996) and the Kandalaksha graben dykes (Beard et al., 1996). Fields of data are also shown for Fen (Andersen, 1984) and Alno (Deines & Gold, 1973) for comparison. Solid black line box = field of "primary igneous carbonatite" (Hoefs 1987). Dashed black line box = field of mantle C-O values, defined by oceanic basalts (Deines, 1989). Shaded circle = field of data from fresh natrocarbonatite lavas from Oldonyo L'Engai (Keller & Hoefs, 1995). Line (i) = effect of re-equilibration with meteoric water according to the model of Andersen (1984). Line (ii) = average fractionation trend for co-existing C-O isotope pairs (Deines, 1989). Line (iii) = effect of crustal contamination (Andersen, 1984).

were seen (not shown).

Figure 6.8 compares the Turiy stable carbon and oxygen isotope results with data from “primary igneous carbonatites” (Hoefs, 1987), the Khibina massif (Zaitsev, 1996), the Kovdor massif (Zaitsev & Bell, 1995), and carbonatite dykes from Kandalaksha (Beard et al., 1996). Fields of data from Alnö (Deines & Gold, 1973) and the Fen carbonatite complex (Andersen, 1984), both of which are similar in age to the Turiy pyroxenite sample C.AG.347, are also shown, along with a field representing mantle CO₂ as found in oceanic basalts (Deines, 1989). It can be seen that the Turiy samples have, in general, heavier $\delta^{13}\text{C}_{\text{PDB}}$ ratios than the “primary mantle-derived carbonatites” of Hoefs (1987) but similar $\delta^{18}\text{O}_{\text{SMOW}}$ signatures. Fifty percent of all mantle-derived carbonatites have average values of $\delta^{18}\text{O}_{\text{SMOW}}$ 6-9‰ and 91% of all carbonatites have $\delta^{13}\text{C}_{\text{PDB}}$ -2 to -8‰ (Deines, 1989). However, it should be remembered that those samples with heavy $\delta^{13}\text{C}_{\text{PDB}}$ values normally have similarly heavy $\delta^{18}\text{O}_{\text{SMOW}}$ values (line [ii] in Figure 6.9) - a feature which is not seen here.

The Turiy results are broadly similar to those from Kovdor, Kandalaksha, Fen and Alnö, in terms of having unusually enriched $\delta^{13}\text{C}_{\text{PDB}}$ values, as seen in Figure 6.8. Zaitsev & Bell (1995) attributed the wide variation in Kovdor $\delta^{13}\text{C}_{\text{PDB}}$ values compared to their constant $\delta^{18}\text{O}_{\text{SMOW}}$ values to mixing between two unspecified mantle components with different $\delta^{13}\text{C}_{\text{PDB}}$ values but similar $\delta^{18}\text{O}_{\text{SMOW}}$ values. Values for calcite carbonates from the Khibina massif (Zaitsev, 1996) have $\delta^{18}\text{O}_{\text{SMOW}}$ +5.9 to +10.2‰ and $\delta^{13}\text{C}_{\text{PDB}}$ -5.5 to -6.8‰ which fall within the field of “primary igneous carbonatites” of Hoefs (1987) and which represents similar $\delta^{18}\text{O}_{\text{SMOW}}$ values but with much lower $\delta^{13}\text{C}_{\text{PDB}}$ values than those of Turiy. The Kandalaksha carbonatites have $\delta^{13}\text{C}_{\text{PDB}}$ values of -3.4 and -4.1‰, and $\delta^{18}\text{O}_{\text{SMOW}}$ values of +9.3 and +8.9‰ respectively (Beard et al., 1996). The Kandalaksha $\delta^{13}\text{C}_{\text{PDB}}$ values are lower, while the $\delta^{18}\text{O}_{\text{SMOW}}$ values are higher than most of the Turiy values.

Stable isotope data measured from carbonatites associated with other melilitolite-bearing intrusions include the Oka carbonatites, which fall in the field of “primary igneous carbonatites” (Conway & Taylor, 1969; Huang et al., 1995), and the Gardiner Complex carbonatites which have values indicative of “fractionation from primary igneous carbonatites” (Nielsen & Buchardt, 1985).

Co-existing dolomite-calcite pairs in the Oka carbonatites have similar $\delta^{18}\text{O}_{\text{SMOW}}$ values. Dolomite-calcite fractionation factors are $\Delta^{18}\text{O}(\text{dolomite} - \text{calcite}) = +1$ to -1‰ (for $\delta^{18}\text{O}_{\text{SMOW}}$ values between $+7$ to $+9\text{‰}$), and $\Delta^{13}\text{C}(\text{dolomite} - \text{calcite}) = 0$ to $+1\text{‰}$ (Deines, 1970). These fractionation effects help to explain the relatively heavy $\delta^{13}\text{C}_{\text{PDB}}$ value of the dolomite-calcite-carbonatite sample [C.DC], but do not help to explain why the Turiy calcite carbonatites have significantly heavier $\delta^{13}\text{C}_{\text{PDB}}$ values with respect to the “primary igneous carbonatite” field of Hoefs (1987).

Deines (1989) made a comparison of $\delta^{13}\text{C}_{\text{PDB}} - \delta^{18}\text{O}_{\text{SMOW}}$ regression lines from 25 carbonatite complexes world-wide. 21 of the 25 complexes examined showed a relatively constant gradient of $+0.4$. The reason for this trend is currently unknown. This trend, however, is not shown by the Turiy data. Instead, there appear to be negative correlations within the main bulk of the samples (Figure 6.7).

The Turiy $\delta^{18}\text{O}_{\text{SMOW}}$ values are typical of many mantle-derived carbonatites, but Turiy’s carbon values, along with those from Kovdor, are generally more enriched than would be expected from co-existing $\delta^{18}\text{O}_{\text{SMOW}}$ values. Since typical fractionation processes cannot be used to explain this discrepancy, it suggests a possible decoupling between oxygen and carbon in the mantle source(s) of the carbonatites. The Khibina Massif lies, geographically, between Turiy and Kovdor, but contains a different suite of alkaline rocks (mostly syenites) and exhibits stable isotope ratios which lie within the field of “primary igneous carbonatites” (Hoefs, 1987). Given the similarity in ages between these three massifs, the enrichment in $\delta^{13}\text{C}_{\text{PDB}}$ values shared by the two

melilitolite-ijolite-carbonatite massifs (Turiy and Kovdor), (along with similar data from the Alnö and Fen carbonatites which have similar ages to the older Turiy pyroxenites), may relate to geographical variations in the $\delta^{13}\text{C}_{\text{PDB}}$ composition of the mantle source.

Unlike radioactive isotope systems, fractionation effects are important in defining the isotopic signatures of the light isotopes. Such fractionation effects are well known in crustal environments, but poorly defined in mantle rocks due to the difficulties with experimental work simulating mantle conditions. $\delta^{18}\text{O}_{\text{SMOW}}$ fractionation effects have been documented between mantle silicate phases (olivine, diopside, enstatite) using silicate-carbonate systems at high pressures and temperatures (Chiba et al., 1989; Rosenbaum et al., 1994), while $\delta^{13}\text{C}_{\text{PDB}} - \delta^{18}\text{O}_{\text{SMOW}}$ fractionation factors have also been measured for CO_2 -calcite at high temperatures (Rosenbaum, 1994), but more work is needed in order to understand the effects of isotopic fractionation on carbon and oxygen in carbonate-silicate mantle systems. Given the large quantities of O in the silicate phases in the mantle, it is likely that these phases will buffer the oxygen isotopes of any primary carbonate or CO_2 . However, since mantle silicate phases do not contain carbon, the source of the high $\delta^{13}\text{C}_{\text{PDB}}$ values must be sought elsewhere.

6.7 Origin of heavy $\delta^{13}\text{C}_{\text{PDB}}$ values

The origin of the heavy $\delta^{13}\text{C}_{\text{PDB}}$ values (relative to their conjugate $\delta^{18}\text{O}_{\text{SMOW}}$ values) from calcite separates from the Turiy carbonatites and phoscorites warrants discussion, especially when such enrichment is common to other carbonatites of similar age to Turiy (Alnö, Fen and Kandalaksha), all located within the Baltic Shield area. Heavy $\delta^{13}\text{C}_{\text{PDB}}$ values are also more common in intrusive than extrusive carbonatites (Deines, 1989; Zaitsev & Bell, 1995). Several factors have been proposed to produce $\delta^{13}\text{C}_{\text{PDB}}$ enrichment (with constant or lower conjugate $\delta^{18}\text{O}_{\text{SMOW}}$ values in CO_2 or carbonate phases), including re-equilibration with meteoric water (Andersen, 1984) and

oxidation of mantle diamond (Deines, 1973). More recent experimental data and discussion regarding the behaviour of carbon phases, oxidation states, and stable isotope fractionation in the mantle (e.g. Exley et al., 1986; Falloon & Green, 1990; Luth, 1993; LaTourette & Holloway (1994); Rosenbaum, 1994; Rosenbaum et al., 1994; Barker, 1996; Dalton & Presnall, 1997) provide new constraints on the interpretation of stable isotope signatures of carbon-bearing phases in igneous rocks.

One cause of carbon isotope enrichment within the Turiy carbonatites, may relate to post-magmatic alteration of the original stable isotope signatures due to interaction with seawater. However, evidence of hydrous alteration in the Turiy rocks is, on the whole, not seen. In fact, Turiy is notable for the unaltered nature of its mineralogy, particularly within the central complex where the majority of the carbonatites and phoscorites are located. Devonian seawater has values of $\delta^{13}\text{C}_{\text{PDB}} +1.5\text{‰}$ and $\delta^{18}\text{O}_{\text{SMOW}} +4\text{‰}$ (Lohmann & Walker, 1989). Mixing of isotopic compositions similar to those within the field of “primary carbonatite” (Figure 6.8) (e.g. $\delta^{13}\text{C}_{\text{PDB}} -4\text{‰}$ and $\delta^{18}\text{O}_{\text{SMOW}} +10\text{‰}$) with Devonian, or even modern-day seawater ($\delta^{13}\text{C}_{\text{PDB}} 0\text{‰}$ and $\delta^{18}\text{O}_{\text{SMOW}} 0\text{‰}$), could produce most of the variation seen in the Turiy stable isotope data. However, there is little mineralogical, geochemical or radiogenic isotope evidence for extensive involvement of hydrous fluids in the petrogenesis of the central complex rocks, and thus the involvement of sea-water in the derivation of the stable isotopic signatures of the central complex carbonatites and phoscorites is likely to be insignificant.

Andersen (1984) invoked re-equilibration of the early Fen carbonatites with meteoric water to produce some of the most $\delta^{13}\text{C}_{\text{PDB}}$ enriched and $\delta^{18}\text{O}_{\text{SMOW}}$ depleted signatures in the later Fe-rich carbonatites. This re-equilibration trend is shown as line [i] in Figure 6.8, and could explain the heavier $\delta^{13}\text{C}_{\text{PDB}}$ data for Fen, Turiy and possibly Kandalaksha and Alnö, but the Kovdor data is harder to explain by this means. If the Kovdor data originally lay in a SW-NE trending line on Figure 6.8, typical of a Rayleigh

fractionation trend (line [ii] in Figure 6.8), and the most fractionated samples were subsequently subjected to re-equilibration with meteoric water, the observed vertical array could be produced. It should be noted that the Turiy data lie on just such a re-equilibration path (trending SW-NE) in Figure 6.8, parallel to the meteoric water re-equilibration line, in between the Kandalaksha data point with heavier $\delta^{18}\text{O}_{\text{SMOW}}$ and lighter $\delta^{13}\text{C}_{\text{PDB}}$ values, and the Kovdor data with the heaviest $\delta^{13}\text{C}_{\text{PDB}}$ values. However, Taylor & Forester (1979) suggested that there is insignificant meteoric water circulation in non-porous metamorphic country rocks, such as those seen around the Turiy peninsula. This implies that if meteoric water exchange has taken place, it has done so through fracture systems, possibly relating to the formation of the grabens in which these massifs sit.

The third line [iii] shown in Figure 6.8, indicates an increase in $\delta^{18}\text{O}_{\text{SMOW}}$ values, and represents a trend corresponding to crustal contamination. Three of the outlier points in Figure 6.7 [N.63.4, N.64.18m and S.W58] lie to the right of the main cluster of Turiy data, and all three samples also show enriched $^{87}\text{Sr}/^{86}\text{Sr}$ isotopic signatures. This suggests that crustal contamination may be at least partly responsible for the enriched $\delta^{18}\text{O}_{\text{SMOW}}$ values of these three samples.

A major carbon reservoir within the Earth's carbon cycle is represented by organic carbon. The only method of incorporating organic carbon into an igneous carbonatite would be through subduction and mantle recycling of sedimentary carbon. Barker (1996) strongly advocates carbon recycling in the mantle, but does not support his arguments with any geochemical data. Plant life typically contains very light ($\delta^{13}\text{C}_{\text{PDB}} - 25\%$) carbon, and the fractionation necessary to create the enriched isotopic values seen in Turiy seems unrealistic. Subduction and recycling of marine carbonate with $\delta^{13}\text{C}_{\text{PDB}}$ values ~ 0 is an alternative, more realistic solution (assuming buffering of the $\delta^{18}\text{O}_{\text{SMOW}}$ values by silicate minerals). However, this should also be borne out by the radiogenic

isotope systems, and, even if elevated $^{87}\text{Sr}/^{86}\text{Sr}$ signatures could be considered typical of subduction zone magmatism, it seems highly improbable that such effects would only be seen in the satellite complexes, and there are certainly no sign of radiogenic subduction zone signatures from the carbonatites and phoscorites with high $\delta^{13}\text{C}_{\text{PDB}}$ values.

Zaitsev & Bell (1995) invoked mixing of unspecified mantle components to produce the vertical trend seen in the Kovdor data in Figure 6.8, on the assumption that heavy carbon is contained in one of the mantle phases. Mantle carbon occurs in a number of forms including elemental diamond and graphite. CO_2 occurs as inclusions in silicate phases (Rosenbaum, 1996), and becomes a free phase above $\sim 50\text{km}$ depth within the lithosphere, created by the breakdown of carbonate phases such as magnesite and dolomite (Falloon & Green, 1990). CO_2 is also the primary form of dissolved carbonate in silicate melts derived from the mantle (Mysen & Virgo, 1980). The isotopic signature of mantle CO_2 was measured by Exley et al. (1986) in the mantle plume-derived Hawaiian (Loihi) basalts. Using step-wise combustion to $> 600^\circ\text{C}$, the Loihi samples produced $\delta^{13}\text{C}_{\text{PDB}}$ values ranging from $\delta^{13}\text{C}_{\text{PDB}} -2$ to -9‰ , higher than Hawaiian total-carbon results, or those from MORB glasses, and the only oceanic samples to display $\delta^{13}\text{C}_{\text{PDB}}$ values higher than -5‰ (Exley et al., 1986).

The relative geographical locations, compositions and relative sizes of the Kovdor, Khibina and Turiy massifs, along with their carbon isotopic compositions, might suggest the presence of a Palaeozoic mantle plume, centred close to the Khibina/Lovozero massifs, as proposed by Zaitsev & Bell (pers. comm., 1997). If isotopic fractionation of CO_2 occurred across a plume-head, it could produce a progressive variation in stable isotope values. Isotopic fractionation of CO_2 - calcite at high temperatures (Rosenbaum, 1994) shows that the CO_2 becomes isotopically enriched in $\delta^{13}\text{C}_{\text{PDB}}$ as fractionation progresses, although the isotopic fractionation effects associated with CO_2 dissolved in a silicate melt will be more moderate than the values measured in

the pure CO₂-calcite system (Rosenbaum, pers. comm., 1997). If the earliest, heavy CO₂ produced during plume ascent migrates to the edges of the plume (Turiy, Kovdor), the residual, lighter $\delta^{13}\text{C}_{\text{PDB}}$ values will be seen at the centre (Khibina). Alternatively, the enriched $\delta^{13}\text{C}_{\text{PDB}}$ values from Kandalaksha, Kovdor and Turiy, may relate to a source specific to the Kandalaksha Graben system along which controlled the emplacement of these carbonatite massifs. The Khibina Massif is located on the Kontozero Graben trending south-west - east-north-east across the Kola Peninsula and is not affected by the Kandalaksha Graben system.

Variation in carbon isotope values may also be related to changing oxygen fugacity of the carbonatite source(s) within the mantle. Can diamond be oxidized to form part of the CO₂ or carbonate phases present in carbonatites? Schrauder & Navon (1993) reported the presence of CO₂ inclusions in natural diamond, proving the co-existence of at least two carbon-bearing phases in some parts of the mantle. Luth (1993) and LaTourette & Holloway (1994) suggested that oxidation-reduction reactions involving carbon can indeed take place under mantle conditions. Luth (1993) provided reactions demonstrating the relationships between diamond-eclogite, diamond-harzburgite and diamond-lherzolite:

dolomite   coesite   clinopyroxene   diamond

orthopyroxene   magnesite   olivine   diamond

orthopyroxene   magnesite   olivine   clinopyroxene   diamond

Differences in the $f\text{O}_2$, pressure and temperature between the stability fields of these

assemblages defines an area in P-T-fO₂ space (Luth, 1993) where diamond is stable in eclogites and carbonate is stable in harzburgites and peridotites. Any changes in mantle oxidation state, on a localized or regional scale, could be caused by any number of factors, including changes in abundance and/or oxidation state of S, Fe, and H (Canil et al., 1990), and could thus affect the stability of different carbon-bearing phases.

Deines (1973) predicted that oxidation of mantle diamond could produce magmatic carbonate phases which would be ~ 2‰ more enriched than the original diamond $\delta^{13}\text{C}_{\text{PDB}}$ values. New isotope fractionation data (Rosenbaum, 1994) now allow a quantitative examination of Deines's model. The average $\delta^{13}\text{C}_{\text{PDB}}$ value for mantle-derived diamonds brought to the surface by Russian kimberlites is -7.1‰ (Galimov, 1984) and 90% of the world's diamonds have $\delta^{13}\text{C}_{\text{PDB}}$ values of -5 to -7‰ (Deines, 1989). It should be possible to calculate the stable isotopic composition of carbonate (e.g. calcite) produced from mantle diamond, using C-CO₂ and CO₂-calcite fractionation factors. The CO₂-calcite fractionation factors also refer to fractionation between gaseous CO₂ and solid calcite. Thus, these fractionation values are maxima in environments where CO₂ is dissolved in any silicate melt as a carbonate anion (i.e. behaving as a "pseudo-carbonate"), and thus the results of such fractionation calculations should be treated as estimated values (Rosenbaum, pers. comm., 1997). However, calcite is not a stable phase within the diamond stability field in the mantle. Fractionation factors for dolomite-CO₂ have not been measured at high temperatures and pressures, but low-temperature and pressure values suggest that the formation of dolomite will produce more enriched $\delta^{13}\text{C}_{\text{PDB}}$ values than those produced by calcite. The fractionation curves for mantle carbon have been estimated by Bottinga (1969) as $\Delta^{13}\text{C}(\text{CO}_2 - \text{diamond}) = +5.5$ to $+6.0$ ‰ at 900°C. Calculated values of diamond-CO₂ and CO₂-calcite fractionation at 900°C are shown in Table 6.F below. While realising that this temperature is unrealistic, it provides a clue as to the relationship between the possible values of primary mantle

CO₂ and oxidized diamond, although in terms of mass balance, it would also be unrealistic to suggest that all the carbonate present in $\delta^{13}\text{C}_{\text{PDB}}$ -enriched carbonatites is produced from mantle diamond.

Table 6.F $\delta^{13}\text{C}_{\text{PDB}}(\text{‰})$ calcite values (calculated)

@ = 900°C	$\Delta^{13}\text{C} (\text{CO}_2 - \text{diamond}) = +5.5$		$\Delta^{13}\text{C} (\text{CO}_2 - \text{calcite}) = +2.7$	
Carbon source / phase	diamond -->	CO ₂ -->	calcite	
Russian diamond $\delta^{13}\text{C}_{\text{PDB}}(\text{‰})$	-7.1	-1.6	-4.3	
Mantle CO ₂ (Hawaii) $\delta^{13}\text{C}_{\text{PDB}}(\text{‰})$	n.a.	-5.0	-7.7	

therefore: calcite produced in equilibrium with mantle CO₂ = -7.7‰

calcite produced in equilibrium with mantle diamond = -4.3‰

[data sources: Bottinga (1969); Galimov (1984); Exley et al. (1986); Rosenbaum (1994)]

Thus it can be seen that the CO₂ and calcite produced from diamond oxidation Table 6.F, line 1) are more enriched in $\delta^{13}\text{C}_{\text{PDB}}$ than typical mantle CO₂ and the calcite produced from it (Table 6.F, line 2). This fits the prediction made by Deines (1973).

Rayleigh fractionation is an important influence on the stable isotope compositions of slowly-cooled, intrusive carbonatite bodies. The Rayleigh fractionation curve for carbon isotope values from the Oka complex (Deines, 1970) has early (unfractionated) calcite $\delta^{13}\text{C}_{\text{PDB}}$ values of -7.0‰, which matches those which would be produced by fractionation from primary mantle CO₂ according to Table 6.F. Deines (1970) subsequently plotted the remaining $\delta^{13}\text{C}_{\text{PDB}}$ results from Oka using a closed-system Rayleigh fractionation model, which produced an extremely good fit to the data, thus suggesting that mantle CO₂ plays a dominant role in the production of carbonatites. The calcite produced by equilibrium oxidation and fractionation of mantle diamond

(Table 6.F) is still not sufficiently enriched in ^{13}C to produce the $\delta^{13}\text{C}_{\text{PDB}}$ values of -3.1 to -2.1‰ seen in Turiy, but Rayleigh fractionation of the initial carbonate produced by oxidation of mantle diamond (-4.3‰) would produce values similar to those found in the Turiy carbonatites.

In evaluating the different models for producing heavy $\delta^{13}\text{C}_{\text{PDB}}$ values, it should be remembered that: a) mantle CO_2 is not necessarily isotopically homogeneous over time or space; b) the fractionation factors used in the calculations in Table 6.F for temperatures of 900°C may not be directly applicable to mantle conditions, and c) no attempt was made to constrain the effects of CO_2 inclusions in the Turiy samples.

Changes in the oxidation state of the mantle may also have profound implications for the timing of carbonatite formation during Earth history, and possibly the relationship of carbonatites to kimberlites, remembering that some of the latter are diamond-bearing. Kimberlites and carbonatites are often spatially related in many areas of the world, including Turiy and the Terskii Coast. Haggerty (1994), in a global comparison of the ages of kimberlites and carbonatites, showed that the relationship between the two is episodic, with kimberlite emplacement preceeding carbonatites by an average of 20 Ma., which matches the kimberlite - carbonatite age relationship at Terskii and Turiy, as described earlier.

One of the implications of the 20 Ma kimberlite - carbonatite age-gap proposed by Haggerty (1984) could be that kimberlite eruptions, incorporating carbon in the form of diamond (and often carbonate phases as well) are followed by oxidation events where mantle diamond and/or graphite is oxidized to CO_2 . If this were to occur, the excess CO_2 could alter the slope of a volatile-poor mantle solidus to a volatile-rich mantle solidus, and trigger subsequent carbonatitic magmatism.

The sources of mantle CO_2 are still uncertain. Although the core is probably still outgassing primitive CO_2 , the issue of recycling of crustal or atmospheric carbon and the

effects of mantle convection on the physical state of carbon and its most stable chemical form on the fractionation of carbon isotopes are still unknown. The sheer quantity of alkaline magma intruded into the Kola Peninsula between 360-380 Ma (Kramm et al., 1993) indicates rather unusual mantle conditions at that time, although the relationship between this unusually large quantity of alkaline magmatism, and the stable isotopic signature of the carbonatites contained within the resultant massifs cannot be constrained at this time.

The proposal that the heavy carbon values seen in carbonatites from Turiy, Kovdor, Kandalaksha, Alno, Fen, all lying within the Baltic-Kola province, may be derived from a source of carbon common to this terrane block, is seen as the most significant finding from this discussion.

6.8 Isotopes - current answers and remaining questions

The isotope data presented within this chapter help to provide answers to three questions regarding Turiy:

1. What is the age of the rocks being investigated?
2. What are the sources which contributed to the Turiy magmas?
3. How do the different rock types relate to each other in terms of their petrogenesis?

The answers to these questions also cover issues such as the extent of open-system disequilibrium during Turiy's magmatic history.

The most recent geochronological work done on the Turiy material by Kramm et al. (1993) suggested that Turiy is but one of a suite of alkaline intrusions across the Kola Peninsula, intruded within Devonian times. This current study has now shown that Turiy magmatism was not restricted to a single event, but contains both Precambrian and Devonian intrusions. The latter may also be represented by more than one intrusive period. The fact that some of the pyroxenites are significantly older than the remainder of

the Turiy must be accommodated by any model put forward to explain their petrogenesis.

The $^{87}\text{Sr}/^{86}\text{Sr}$ - $^{143}\text{Nd}/^{144}\text{Nd}$ isotope data presented in Figure 6.4 clearly indicate that the Devonian rocks at Turiy are not the products of closed-system fractionation of a single batch of magma. Open-system conditions have prevailed, and most of the rock types are characterized by their own unique suite of isotopic signatures. The isotope ratios at Turiy fall on or close to the Kola Carbonatite Line of Kramm (1993), consistent with mixing of melts from “enriched” and “depleted” sources. The former may reside within the lithosphere, and the latter within the asthenosphere, perhaps associated with a mantle plume. Although mixing of these sources seems to be reflected by most of the Turiy rocks, certain suites of rocks are preferentially dominated by one of the components. The rocks dominated by the depleted, asthenospheric component appear to be the pyroxenites (Precambrian and possibly Devonian), the uncontaminated melilitolites and phoscorites and some carbonatites, particularly those from the northern complex. Those dominated by the enriched, possibly lithospheric, component include the olivine melteigite, both suites of ijolitic rocks, and possibly the turjite. The source of the central massif carbonatites is more ambiguous, but appears to be closer to the asthenospheric component. If the isotopic data are linked with the intrusion sequence of the rock types at Turiy, as defined by Bulakh & Ivanikov (1984), the following pattern emerges:

1. pyroxenites	depleted	4. ijolites (2nd series)	enriched
2. ijolites (1st series)	enriched	5. phoscorites	depleted
3. melilitolites	depleted	6. carbonatites	depleted(?)

There is an interesting repetitiveness to this pattern, although there is currently not enough isotope data to fully constrain such periodicity.

The Sr-Nd isotopes also indicate that there may have been crustal contamination of some of the rocks from the satellite complexes surrounding the central complex at

Turiy, including several carbonatites. Owing to some sample bias, this feature appears to be dominant within the melilitolite samples, northern carbonatites, and south coast dykes, as seen in Figure 6.4 and Tables 7.A. and 7.B. The source of the contaminants is not known, but given the high concentrations of Sr and Nd in the melilitolites in particular, it is unlikely that they have been affected by post-magmatic contamination. Thus, their isotopic signatures are believed to have been derived prior to and possibly during their emplacement at their current level in the crust. Isotopic mixing calculations suggest that the contaminant is probably Proterozoic country rock. It is suggested that the development and emplacement of the satellite complexes involved some crustal assimilation or partial melting or dissolution of volatile-rich phases within the country rock. The Devonian magmas within the central complex would have re-utilized the plumbing system originally established by the Precambrian rocks, and restricted the effects of further crustal involvement. The reason why the central complex phoscorites and carbonatites fall mostly below the KCL will be discussed further in a later chapter.

An overall model for Turiy suggests the involvement of an asthenospheric mantle plume, rich in carbonate components, producing magmatic activity within a classic rift system (Kandalaksha Graben), accompanied by periodic melting of an enriched source, possibly located within the overlying lithospheric mantle. Crustal contamination may have occurred during the formation of satellite complexes around the main intrusion centre.

The stable isotope results also point to a mantle source for the carbonatites, and the enriched $\delta^{18}\text{O}_{\text{smow}}$ values from the northern complex carbonatites support the crustal contamination model for the satellite complex rocks suggested by the enriched $^{87}\text{Sr}/^{86}\text{Sr}$ data from these samples. There is a possible source-link between the central complex phoscorites and carbonatites with other carbonatites within the Baltic Shield area, and particularly from the Kandalaksha Graben, characterized by enriched $\delta^{13}\text{C}_{\text{PDB}}$ values. The

cause of this enrichment is uncertain, but may relate to a variety of mechanisms that range from meteoric or sea-water re-equilibration, to an isotopically zoned plume head, or to diamond oxidation within the mantle.

CHAPTER 7

ELEMENT CHEMISTRY

7.1 Introduction

Major and trace element abundances for a suite of sixteen samples were determined using XRF and ICP-MS techniques. The samples, chosen on the basis of their textural homogeneity, represented a wide range of rock types from the Turiy complex. Sample preparation techniques, analytical uncertainties and detection limits are outlined in Appendix H. The samples were initially broken in a steel jaw-crusher, and subsequently ground in an agate mill in order to minimize contamination errors. The precision of the data is dependent on the concentration of each element in each sample. A mathematical formula, based on the identity of an element and its concentration in an analysis, has been derived to calculate the precision of the XRF results (Kretz et al., unpublished; Appendix H), but similar formulae are not available for the ICP-MS data. The precision of the major element concentrations, quoted as oxides from the XRF analyses, is $< 0.1\text{ wt.}\%$, and ranges from 0.3 - 2% of the element concentration value. The size of the error is inversely proportional to the element concentration. The precision of the trace element analyses from the ICP-MS is between 5-10% for most elements. The accuracy of the ICP-MS data is estimated to be 1-20% of the quoted values, depending on the element being analyzed. The highest errors are associated with the Pb and HREE data.

The element chemistry of samples from Turiy help to evaluate relationships among samples from within the complex, and allow comparisons to be made with rocks from other melilitolite-bearing intrusive massifs, and various extrusive samples from other parts of the Earth. It should be stressed that the majority of samples examined from Turiy are texturally heterogeneous. Samples picked for trace element analyses were chosen on the basis of their textural homogeneity, and lack of visible alteration.

7.2 Major element geochemistry

7.2.a Silicate rocks

Of the intrusive silicate rocks, the melilitolites have lower SiO_2 contents (35.5-37.5 wt.%) than either the ijolite series rocks or the pyroxenite (39.2-40.9 wt.%) (Table 7.A). The Al_2O_3 contents of the silicate samples show a bimodal distribution which is unrelated to rock type. Samples C.90.29 [pyroxenite] and TUR II [melilitolite] have lower Al_2O_3 contents (6.7 & 8.2 wt.% respectively) than the ijolites and other melilitolites which have values between 12.5-15.5 wt% Al_2O_3 . The Fe values (total Fe quoted as Fe_2O_3) range from 10.0-12.3 wt.% for all samples except the pyroxenite which is more enriched than any of the other silicate rocks (14.4 wt.% Fe_2O_3). All of the silicate rocks except C.90.29 [pyroxenite] and TUR II [melilitolite] have $\text{Al}_2\text{O}_3 / \text{Fe}_2\text{O}_3 > 1$. Both of these samples may be partial cumulates, given their high pyroxene and melilite abundances respectively.

TiO_2 values range from 2.0-3.1 wt.%. Sample TUR II [melilitolite] has the lowest content and sample C.AG.122 [melteigite] the highest. P_2O_5 contents are low in all of the samples (< 0.3 wt.%) except for C.IJ.14 [ijolite] and WKT [turjaite] which have higher P_2O_5 contents (1.44 & 0.67 wt.% respectively). These P_2O_5 values are still lower than those found in their supposed volcanic equivalents nephelinite and melilitite (e.g. Hoernle & Schimke, 1993).

MgO concentrations are low in the silicate magmas (5.1-7.1 wt.% MgO) except for the pyroxenite which contains 11.3 wt.%. Magnesium numbers calculated as $\text{Mg\#} = 100[\text{Mg}/(\text{Mg} + \text{Fe}^{2+})]$, on the basis of total Fe as Fe^{2+} , range from 47.0 to 57.5, while the pyroxenite has an Mg # of 61. Typical extrusive melilitites from Germany, Algeria and the Canary Islands (including the type-locality olivine melilitite from the Urach volcanic province in southern Germany), considered to be representative of “primary magmas”

Table 7.A

Major and trace element data

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wt. %	C.90.29 pyrox.	C.AG.122 melteigite	TU 107 mel dyke	C.5.118 mel.	C.190.105 mel.	WKT mel.	TUR II mel.
SiO ₂	40.90	39.19	36.51	36.81	37.45	35.56	35.51
TiO ₂	2.92	3.13	3.00	2.51	2.49	2.45	2.03
Al ₂ O ₃	6.68	13.00	8.72	14.91	13.76	15.47	8.19
Fe ₂ O ₃	14.34	11.53	12.42	9.97	10.38	11.06	12.36
MnO	0.13	0.17	0.19	0.14	0.13	0.18	0.18
MgO	11.28	5.16	7.23	6.80	6.03	5.19	7.09
CaO	20.73	16.86	21.93	16.47	19.76	17.98	27.10
Na ₂ O	0.86	5.90	2.48	6.29	6.02	7.03	3.74
K ₂ O	0.17	2.05	1.72	3.47	2.19	2.72	0.61
P ₂ O ₅	0.05	0.02	0.30	0.18	0.33	0.65	0.05
L.O.I.	0.30	1.10	5.40	1.60	0.50	1.30	1.70
TOTAL	98.37	98.11	99.90	99.15	99.05	99.60	98.57
Mg #	60.9	47.0	53.6	57.5	53.5	48.2	53.2
ppm							
V	165	175	216	102	105	103	113
Cr	77						
Co	53.5	38.2	57.7	37.4	41.3	72.7	32.3
Ni	35					66	
Cu	612	438	595	360	266	192	65
Zn	64	118	167	95	121	98	52
Ga	15	23	18	20	19	18	15
Ge	1.7	1.6	1.2	3.0	1.4	1.0	0.9
As			8	7		5	
Rb	4	45	62	95	44	67	16
Sr	548	1147	1270	2831	2243	2519	3484
Y	12	48	18	14	21	23	15
Zr	339	326	294	53	91	70	95
Nb	19	205	111	226	227	285	146
Mo	0.17	1.27	0.53		1.03	8.31	
Ag	0.7						
In	0.1						
Sn	2.8	5.8	1.4	0.6	0.7	0.7	0.6
Sb	0.10	0.16	0.38	0.32	0.12	0.68	0.42
Cs	0.06	0.25	1.45	3.36	0.26	0.38	0.19
Ba	59	203	518	1552	328	477	256
La	30	74	107	241	203	259	194
Ce	79	235	219	458	401	451	391
Pr	9.1	29.6	21.2	41.5	36.1	37.4	35.8
Nd	44.1	151.9	85.6	155.3	143.8	140.0	136.4
Sm	7.7	29.7	12.3	20.0	21.2	21.2	18.1
Eu	2.21	8.36	3.16	4.77	5.64	5.61	4.21
Gd	5.3	21.6	8.4	12.2	13.5	15.1	11.0
Tb	0.65	2.56	0.90	1.27	1.50	1.69	1.10
Dy	2.92	11.78	4.02	4.69	6.80	7.22	4.41
Ho	0.44	1.72	0.64	0.69	0.92	0.97	0.65
Er	1.03	4.15	1.55	1.32	1.87	1.91	1.28
Tm	0.10	0.36	0.13	0.06	0.10	0.10	0.06
Yb	0.79	2.11	0.76		0.43	0.46	0.12
Lu	0.15	0.31	0.13	0.04	0.05	0.05	0.02
Hf	10.83	5.98	6.73	1.03	2.21	1.90	1.79
Ta	1.97	19.91	7.26	21.98	15.83	14.90	14.97
W	0.60	0.98	1.20	0.37	0.82	0.72	0.21
Tl	0.04	0.09	0.22	0.07	0.15	0.07	0.19
Pb			9	6	5	5	
Bi	0.30	0.11	0.19	0.23	0.12	0.14	0.14
Th	1.64	19.44	8.60	30.26	16.15	16.77	22.16
U	0.20	4.17	1.86	2.71	3.06	4.23	2.16

Table 7.A

Major and trace element data (cont.)

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wt. %	C.49.40 carb.	C.TL.344 carb.	C.DC dol.carb.	S.W58 carb.	C.AG.48A ol.melt.	C.IJ.14 ijolite	TU.119 turjite	NC.153.82 fenite
SiO ₂	4.98	4.81	0.00	0.71	44.02	40.86	31.65	56.18
TiO ₂	0.20	0.75	0.01	0.17	1.17	2.57	3.23	0.42
Al ₂ O ₃	1.21	0.26	0.01	0.15	5.08	12.57	8.84	15.02
Fe ₂ O ₃	7.91	12.57	1.87	5.07	11.37	10.12	15.18	3.91
MnO	0.16	0.14	0.31	0.64	0.19	0.20	0.24	0.08
MgO	6.52	2.39	9.34	4.80	20.00	5.51	2.56	1.62
CaO	39.45	42.58	42.84	41.63	13.94	15.91	23.28	7.85
Na ₂ O	0.03	0.21	0.00	0.02	2.01	6.49	3.26	7.04
K ₂ O	1.26	0.14	0.01	0.12	0.83	3.03	1.77	4.30
P ₂ O ₅	6.97	4.75	2.20	3.37	0.12	0.89	1.44	0.35
L.O.I.	28.50	28.60	43.10	38.80	0.90	1.00	6.00	3.00
TOTAL	97.19	97.20	99.69	95.48	99.63	99.14	97.45	99.78
Mg #	62.0	27.4	90.8	65.2	77.7	51.9	25.0	45.1
ppm								
V	92	123	6	110	116	233	506	86
Cr					1066			29
Co	49.8	30.6	8.8	6.5	81.8	24.6	25.6	8.5
Ni					589			
Cu	20	94	6	19	139	23	245	95
Zn	36	59	12	242	77	52	48	31
Ga	6	6	3	15	10	18	16	18
Ge	0.4	0.5	0.2	1.7	1.3	1.9	3.2	0.7
As				11				12
Rb	32	3	3	2	21	81	62	93
Sr	4580	3284	7123	22397	982	1186	3350	873
Y	16	20	14	110	12	39	124	17
Zr	13	181	6	164	182	565	1207	138
Nb	8	11	3	391	31	175	505	28
Mo			0.47	1.34	0.64		1.63	0.45
Ag								
In			0.1	0.1		0.1	0.2	
Sn	0.3	0.6		0.2	0.8	2.1	5.0	1.0
Sb	0.20	0.19	0.29	0.74	0.34	0.52	0.20	0.27
Cs	0.46	0.19	0.17	0.08	0.21	0.93	1.70	0.58
Ba	347	244	458	1576	358	510	885	1424
La	54	74	89	1119	42	123	122	38
Ce	111	144	159	1944	77	231	234	75
Pr	11.1	14.8	14.4	149.6	7.6	22.9	24.3	7.3
Nd	47.9	63.1	55.4	511.2	33.6	100.5	111.6	31.0
Sm	8.2	10.0	8.3	61.6	5.7	18.1	26.0	5.1
Eu	2.35	2.96	2.32	15.52	1.51	4.95	8.72	1.59
Gd	6.3	8.2	6.4	45.3	4.1	13.8	27.4	4.3
Tb	0.81	0.95	0.70	5.10	0.55	1.63	4.03	0.59
Dy	3.91	4.74	3.21	23.42	2.57	8.28	23.63	3.14
Ho	0.61	0.78	0.54	3.96	0.44	1.40	4.33	0.60
Er	1.44	1.85	1.37	9.48	1.19	3.52	11.97	1.85
Tm	0.13	0.19	0.13	0.84	0.14	0.38	1.47	0.22
Yb	0.80	0.76	0.69	3.57	0.99	2.34	9.39	1.34
Lu	0.13	0.16	0.12	0.51	0.13	0.37	1.24	0.22
Hf	0.15	2.76		0.99	4.67	11.31	24.43	3.58
Ta	1.46	1.02	0.15	3.03	1.13	11.54	5.39	0.24
W				0.50	0.95	0.52	5.28	0.14
Tl				0.07	0.03	0.11	0.22	0.17
Pb				163		6	7	
Bi	0.11	0.21	0.13	1.57	0.09	0.22	0.08	
Th	1.35	6.53	1.70	102.02	2.87	11.23	7.76	1.19
U	0.14	0.21	0.32	14.91	1.05	3.83	6.66	0.66

and commonly containing a significant amount of xenocrystic material, have Mg#s of 68-78, assuming $\text{Fe}_{\text{TOTAL}} = \text{Fe}^{2+}$ (Wilson et al., 1995). The presence of Fe^{3+} in such magmas can be determined by wet chemical methods (or estimated from mineral analyses and modal mineralogy). Fe^{3+} will not substitute directly for Mg in either the melt or the minerals that crystallize from it, due to charge imbalance, and hence increasing proportions of Fe^{3+} in the melt means that Mg#s (calculated on the basis of total Fe as Fe^{2+}) should be interpreted with caution. A “ $\text{Fe}^{3+} = 15\%$ total Fe” calculation is commonly used to obtain a first approximation of the Fe^{3+} content in silicate magmas (e.g. Simonetti & Bell, 1994). However, wet chemical analyses suggest that this estimate becomes increasingly inaccurate in SiO_2 -poor samples (e.g. Hoernle & Schmincke, 1993). The presence of dolomite in carbonatite massifs shows that Mg - Ca phases are important in such massifs, as well as the Mg - Fe^{2+} phases which are the basis for the calculation and interpretation of the Mg#. The Mg#s of the Mg-Fe-bearing minerals in the Turiy rocks (Chapter 5) are commonly extremely high, and quite dissimilar to the low Mg# values of the bulk rock samples. This discrepancy between bulk rock and mineral Mg#s is controlled by mineral/melt partitioning, and there are few measured partition coefficients available which are applicable to the Turiy melt (or mineral) compositions. The mineral compositions listed in Appendix J suggest that Mg is partitioned preferentially into the silicate mineral phases (with respect to Fe). Based on typical partition coefficients, the extreme Mg#s (~95) in the phlogopites from the melilitolites, which are much higher than the bulk rock Mg#s of ~55, indicate that substantial Fe^{3+} is present in the bulk rock composition. Thus, due to the uncertainty in the true $\text{Fe}^{3+} : \text{Fe}^{2+}$ ratios in these samples, the bulk rock Mg#s quoted in Table 7.A have been calculated on the basis of $\text{Fe}_{\text{TOTAL}} = \text{Fe}^{2+}$, giving values which are minimum estimates of the real Mg : Fe^{2+} ratios in these samples.

CaO contents vary significantly within the silicate rocks and range from 15.9-27.1 wt.%. The CaO contents are generally, but not always, higher in the melilitolites than the ijolites, although the abundance in sample C.AG.122 [melteigite] overlaps with sample C.5.118 [turjaite]. The pyroxenite [C.90.29] has the second highest measured CaO value of 20.7 wt.% which may reflect the presence of cumulus clinopyroxene.

Alkali contents and Na:K ratios vary significantly within the silicate rocks. The lowest alkali contents (0.9 wt.% Na₂O and 0.2 wt.% K₂O) are found in the pyroxenite [C.90.29], and in sample TUR II which is a melilitolite. Most of the silicate rocks have Na₂O contents that range from 5.9-7.3 wt.% and 2.0-3.5 wt.% K₂O. Na/K ratios are highest in TUR II (9.0) and C.90.29 (7.7), and range from 2.2-4.4 for almost all other samples, except for the carbonatites (Na/K = 0-0.25). Thus, it can be seen that none of the silicate rocks are potassic and even the turjite sample has an Na/K ratio of 2.8, which differs from the "average turjite" analysis published by Kranck (1928).

A classification scheme proposed by Le Bas (1989) for melilitites, nephelinites and basanites, was based on a plot of SiO₂ + Al₂O₃ versus CaO + Na₂O + K₂O. The combined SiO₂ + Al₂O₃ totals for both melilitolites, pyroxenite and ijolites suggests that the totals vary little between rock type - the higher Al and lower Si of the melilitolites balancing the lower Al and higher Si of the ijolites. A new classification scheme based on that of Le Bas (1989) (Dunworth, in prep.) suggests that Si (atom %) is a more effective and yet simple discriminant plot for classification purposes. Ca (atom %) was also found to be a more useful rock-type discriminant when plotted against Si than the CaO + Na₂O + K₂O group used by Le Bas, 1989, particularly in defining the nephelinite - melilitite boundary.

A study by Lloyd (1985) showed that the addition of any of CaO, Na₂O or K₂O to a natural melilitite sample which had lost alkalis during post-magmatic alteration, caused subsequent melilite re-precipitation. The addition of such oxides to a silicate melt has the

effect of lowering the relative SiO_2 content of the bulk rock composition which will again enhance the probability of melilite precipitation. Lloyd's results support the fact that melilite can be found in alkali-rich, CaO-poor silica-undersaturated magmas (Figure 7.1), although the major-element factors controlling the formation of melilite in the majority of melilite-bearing magmas still appear to be dominated by low SiO_2 and high CaO contents.

World-wide, most melilitites (and a few melilitolites) fall into a V-shaped area in Ca-Si space (atom %) (Figure 7.1a). The boundary between the melilitites and the nephelinites has a positive gradient, while the most Si-depleted melilite-bearing rocks in Figure 7.1 appear to have an inverse relationship between Si and Ca. The overall trend of the graph, excluding the scattered kimberlite data, reflects a negative relationship between the Si and Ca contents of the rocks. The bulk rock data from the Turiy melilitolites and pyroxenite overlap with the melilitite data in Figure 7.1b, while the melteigite [C.AG.122] (37.1, 17.1), ijolite [C.IJ.14] (37.9, 15.8) and olivine melteigite [C.AG.48A] (40.2, 13.7) from Turiy overlap the nephelinite data. Data from the pyroxenite [C.90.29] (39.8, 21.6) and the melteigite [C.AG.122] fall very close to the melilitite-nephelinite "boundary" which is poorly-defined at such high Ca contents. This is consistent with the idea that the pyroxenites are in some way related to the melilitolite rocks, as previously shown by the mineralogical and isotopic data, and may explain why melteigite sample C.AG.122 contains Tschermak-rich pyroxenes, typically found in melilite-bearing rocks, as described in Chapter 5.

Another point of interest is the combined CaO + MgO contents of the silicate magmas. Dolomite and magnesite are both stable within the mantle (e.g. Falloon & Green, 1990), and their melting will elevate the Ca/Si and Mg/Si ratios of a mantle melt above those created solely by the melting of silicate minerals in the mantle. The effect of such signatures can be seen in a plot of Ca/Si vs Mg/Si for silicate rocks, as shown in

Figure 7.1a Atomic % Ca vs Si for the Turiy silicate samples and a variety of mostly extrusive silicate rocks from around the world. List of data references available from the author.

The data in this diagram and in Figure 7.1b, and those data in Figures 7.2a and 7.2b have all been calculated volatile-free, and the values are quoted in atom %. The calculation of the data in this way is easily achieved by use of a mineral normalization program, normalizing the wt.% oxide data (not including volatiles) to "100 cations". A variety of different plots of the same data set by the author have shown that this method is the most effective for providing good discriminants between the different rock types shown on the diagram. Additionally, the use of atom % aids in the understanding of melt structure behaviour, in terms of fractionation and immiscibility, as will be discussed further in the next chapter.

Figure 7.1b Atomic % Ca vs Si for the Turiy silicate samples which have been divided into groups, and a variety of mostly extrusive silicate rocks from around the world. List of data references available from the author.

Turiy Group 1 = pyroxenite [C.90.29], melilitite dyke [TU 107] and melteigite [C.AG.122]

Turiy Group 2 = melilitolites [C.5.118, C.190.105, WKT, TUR II]

Turiy Group 4 = olivine melteigite [C.AG.48A], ijolite [C.IJ.14], turjite [TU 119] and fenite [NC.153.82].

The reasons behind the division of the samples into the groups listed above (Group 3 = carbonatites) is discussed in section 7.3.

Figure 7.2a. Many silicate igneous rocks, including nephelinites, basanites and basalts, fall in a positive-trending curve in Ca/Si - Mg/Si space (Figure 7.2), excluding two picrite samples with high Mg/Si plotted as part of the basalt-rhyolite data set. Melilitites and kimberlites do not fit this tight trend. The melilitites contain $\text{Ca/Si} > \text{Mg/Si}$, while kimberlites plot in the high Mg/Si - low Ca/Si part of the graph.

The melilitites and kimberlites show considerably more scatter than any of the other rock types shown in Figure 7.2a. The scatter of the melilitite and kimberlite data, compared to the relatively tight clustering of the more SiO_2 -rich samples, probably reflects the contribution that mantle dolomite ($(\text{CaMg})\text{CO}_3$) and magnesite (MgCO_3) make to the parental melts which form these samples. However, Dalton & Presnall (1997) reported that the Ca/Mg ratio of a mantle melt is unaffected by the transition from dolomite to magnesite as the dominant source carbonate phase, and thus the Ca/Mg ratio of the carbonate contribution to carbonatite and kimberlite melts should be similar, which is not seen in Figure 7.2a. The high MgO content of most kimberlites (Mitchell, 1986), and hence high Mg/Si ratios in Figure 7.2a, may also reflect the high content of olivine xenocrysts in these samples.

It is possible that the scatter of the melilitite and kimberlite data reflects inhomogeneous crustal contamination of these rocks, but the tight grouping of the nephelinites and basanites extruded under similar conditions does not support this idea. Given that many of the melilitite samples used in Figure 7.2a are from oceanic environments, and that all analyses used in Figures 7.1 and 7.2 have been recalculated volatile-free to further reduce data scatter, it seems that crustal contamination is unlikely.

Figure 7.2b shows that the Turiy melilitolites [C.5.118, C.190.105, WKT, TUR II] and pyroxenite [C.90.29] have higher Ca/Si ratios than the melteigite [C.AG.122], ijolite [C.IJ.14], and olivine melteigite [C.AG.48A]. The melilitolite Ca/Si ratios are > 0.5 for all but one sample, while the melteigite, ijolite and olivine melteigite Ca/Si ratios

Figure 7.2a Atomic % Ca/Si vs Mg/Si for the Turiy silicate samples and a variety of mostly extrusive silicate rocks from around the world. Data calculated volatile-free. List of data references available from the author.

Figure 7.2b Atomic % Ca/Si vs Mg/Si for the Turiy silicate samples and a variety of mostly extrusive silicate rocks from around the world. List of data references available from the author.

Turiy Group 1 = pyroxenite [C.90.29], melilitite dyke [TU 107], melteigite [C.AG.122]

Turiy Group 2 = melilitolites [C.5.118, C.190.105, WKT, TUR II]

Turiy Group 4 = olivine melteigite [C.AG.48A], ijolite [C.IJ.14], turjite [TU 119], fenite [NC.153.82].

The reasons behind the division of the samples into the groups listed above (Group 3 = carbonatites) is discussed in section 7.3.

vary from 0.34 - 0.46. The Mg/Si values of the Turiy samples show less systematic variation according to rock type. The melteigite [C.AG.122] (0.20) and ijolite [C.IJ.14] (0.20), however, have the lowest values of all the silicate samples (0.20 - 0.68). The elevated Mg/Si ratio of the olivine melteigite (0.68) probably reflects the abundance of megacrystic Mg-rich olivines contained within the sample. The combined Ca+Mg contents are higher for the Turiy melilitolites (24.9 - 38.1 atom %) than the melteigite [C.AG.122] (24.4 atom %) and ijolite [C.IJ.14] (23.4 atom %). The Ca+Mg contents of sample TUR II [melilitolite] (38.1 atom %) and sample C.90.29 [pyroxenite] (38.0 atom %), described earlier in this section as possible cumulates, are almost identical.

The turjite sample, TU 119, has chemical characteristics quite different from any of the other rock types studied, and it is clearly not in textural (Chapters 4 and 5) or chemical equilibrium. On the Ca-Si plot (Figure 7.1b), the data from sample TU 119 (33.4, 26.3) falls within the melilitite field and has a Ca/Si ratio higher than those of all the Turiy melilitolites, due to its high Ca content and low Si content. Its Al/Fe ratio is < 1 , unlike any of the ijolites or melilitolites, most of which have Al/Fe $\gg 1$. The turjite sample [TU 119] is also more enriched in P than any of the other silicate rocks, but is not significantly enriched in Na and K. It has the highest Ti content of any sample analyzed, relatively low Al, and the lowest MgO content of all the silicate samples (2.5 wt.%). It is closest in chemical composition, but not modal mineralogy, to melilitolite sample TUR II. On the basis of the textural and chemical features of its mineralogy, it was suggested by Bell et al. (1996) that turjite might be a natural example of the experimental ijolite-carbonatite reaction carried out by Verwoerd (1978). It is further suggested here, on the basis of petrographic and chemical characteristics, that turjite may represent a reaction product between a magma similar in composition to an ijolite such as sample C.IJ.14 or a fractionated version of the olivine melteigite [C.AG.48A], and a carbonatite. The disequilibrium textures seen in sample TU 119, particularly in the

garnet, suggest that this reaction may not have gone to completion, or that TU 119 is a cumulate representative of the reaction products. Sample TU 120A, a more fine-grained turjite described in Chapter 4 (section 4.8), shows an assemblage that probably reached equilibrium, with only one generation of subhedral zoned garnets, as compared to the subhedral andradite-melanite and anhedral schorlomite seen in sample TU 119. Comparison of the major element composition of sample TU 119 [turjite] with sample C.IJ.14 [ijolite] shows that the turjite is more enriched in Ca, Ti, Fe and P, elements dominant in the Turiy phoscorites and some carbonatites. Reaction of such a carbonatite component with the ijolite sample [C.IJ.14] would produce a composition depleted in elements such as Si, Al, Mg, Na and K with respect to the original ijolite composition, i.e. a melt similar in composition to TU 119. Thus, the major element chemistry is consistent with the ijolite-carbonatite reaction relationship of turjite.

Sample NC.153.82 is a fenitized sample of the Proterozoic charnockites which make up most of the Turiy peninsula country rock. The sample, collected from an area several hundred metres north of the main intrusion, has a completely different major-element geochemistry to any of the silicate samples from the massif. It was analyzed to evaluate the effects of mixing magma from Turiy with the surrounding country rock, and to assess the effect of migrating fluids evolved from the massif. Sample NC.153.82 is rich in SiO_2 , Al_2O_3 and alkalies, poor in MgO , CaO and TiO_2 , and has a high Al/Fe ratio.

7.2b Carbonatites

Three of the four carbonatites analyzed for major and trace element chemistry come from the central complex. Two of them are calcite carbonatites rich in magnetite [C.49.40] which are defined as ferrocarnatites, according to the chemical classification of Woolley & Kempe (1989). The third central complex sample is a dolomite-calcite carbonatite [C.DC], defined as a magnesiocarbonatite using the Woolley & Kempe

(1989) classification scheme. The fourth sample is representative of a south coast carbonatite dyke [S.W58], and can be defined, on the basis of its chemistry, as a calciocarbonatite. It should be noted that there are no Fe-bearing carbonate phases in the two ferrocarbonatite samples. They all show considerable variation in their major-element composition, and in their mineralogy (Chapter 5). Sample C.49.40 is a magnetite-apatite-phlogopite-rich calcite carbonatite and is the only sample analyzed with K_2O (1.26 wt.%) significantly more abundant than Na_2O (0.03 wt.%), and it has the highest P_2O_5 content of any of the samples analyzed. As stated in Chapters 4 & 5, apatite abundance is extremely variable in the carbonatites and phoscorites, and apatite commonly occurs as segregations, veins, and is closely associated or intergrown with magnetite. Dolomite (~5%) and phlogopite with high Mg content (Mg# ~95) occurs in sample C.49.40, which is reflected in the relatively high MgO content of this sample (6.5 wt.%). Sample C.TL.344 is one of the magnetite-rich pyroxene calcite carbonatites from the central massif. Dolomite is rare in this sample and apatite is relatively abundant, with MgO and P_2O_5 contents reflecting the modal mineralogy. Sample C.DC is a late-stage dolomite-calcite carbonatite, also from the central massif. Examination of the C.DC hand-specimen shows that there are no mafic minerals present, but approximately equal amounts of calcite and dolomite (seen clearly when stained with Alizaran Red S), apatite and rare sulphides. Comparison of these observations with the major element chemistry suggest that this sample contains approximately 8% apatite, 46% dolomite (including a small amount of ankerite) and 45% calcite (including 0.5% strontianite). The MnO content of 0.31 wt.% in this sample is relatively low for magnesiocarbonatites (Woolley & Kempe, 1989).

Sample S.W58 is a calciocarbonatite dyke from Turiy's south coast and its relationship to the carbonatites from the central massif is unclear from the major element chemistry. It contains significant apatite, dolomite-rich veins, and its MnO content is the

highest of all the carbonatites (0.64 wt.%).

7.3 Trace element geochemistry

ICP-MS trace element data are presented in Table 7.A. Element ratios and normalized element ratios are listed in Tables 7.B and 7.C respectively. The data are presented graphically in x-y plots and spidergrams. The element ratios and normalized element ratios listed in Tables 7.B and 7.C were used to create several hundred different plots. Only those which showed interesting or systematic variations between rock types are used in this chapter.

Nakamura (1974) was the first to use log-normal plots in which rare earth elements (REEs) were listed in order of decreasing incompatibility (increasing atomic number) along the x-axis. The concentrations in each sample were divided by chondrite abundances, and plotted using a logarithmic y-axis. The chondrite-normalization of the data has the advantage of eliminating the Oddo-Harkins effect, where elements with an even atomic number are naturally more abundant than those with an odd atomic number. Thompson et al. (1983) expanded the normalization concept to include other trace and incompatible elements, listed in order of increasing compatibility, and introduced the idea of normalising the data to the abundances of the elements in “primitive mantle”. Thompson et al. (1983) also coined the term “spidergram” (or “spiderdiagram”) in recognition of the somewhat tentacular patterns which were produced. In a typical spidergram, a variety of incompatible and moderately compatible elements are plotted from left to right in the order of decreasing chemical incompatibility in normal silicate magmas, with a representative selection of elements from different groups within the periodic table. For example, Rb and K are ubiquitous in all spidergrams, but Cs is not always included. Similarly, only certain REE are included. Different authors do not always agree on the order of the elements (i.e. their degree of incompatibility), or on the

Table 7B Trace element ratios (non-normalized)

Sample ID:	C.90.29 pyrox.	C.AG.122 melteigite	TU 107 mel dyke	C.5.118 mel.	C.190.105 mel.	WKT mel.	TUR II mel.	C.49.40 carb.	C.TL.344 carb.	C.DC dol.carb.	S-W.58 carb.	C.AG.48A ol.melt.	C.II.14 ijolite	TU 119 turpite	NC.153.82 fenite
Ca/Si	0.54	0.46	0.64	0.48	0.57	0.54	0.82	1.95	0.74		10.08	0.34	0.42	0.79	0.15
Mg/Si	0.41	0.20	0.30	0.28	0.24	0.22	0.30	8.48	9.48		62.81	0.68	0.20	0.12	0.04
Th/U	8.32	4.66	4.62	11.2	5.28	3.97	10.3	9.75	30.8	5.30	6.84	2.75	2.93	1.17	1.80
U/Pb			0.21	0.45	0.61	0.85					0.09		0.64	0.95	
Hf/Ta	5.50	0.30	0.93	0.05	0.14	0.13	0.12	0.10	2.71		0.33	4.14	0.98	4.53	14.7
Ce/Lu	526	750	1726	11742	7714	8681	16280	858	880	1295	3835	578	619	189	343
Zr/Ta	172	16.4	40.5	2.42	5.75	4.69	6.35	8.6	177	44.1	54.3	161	49.0	224	566
100*Ta/Zr	0.58	6.10	2.47	41.3	17.4	21.3	15.8	11.6	0.57	2.27	1.84	0.62	2.04	0.45	0.18
100*Lu/Ce	0.19	0.13	0.06	0.01	0.01	0.01	0.01	0.12	0.11	0.08	0.03	0.17	0.16	0.53	0.29
K/P	7.21	203	10.7	35.5	12.4	7.89	22.9	0.34	0.06	0.01	0.07	12.6	6.41	2.32	22.9
Rb/K	22.1	21.7	36.0	27.3	20.3	24.7	26.4	25.2	19.9	477	20.2	25.2	26.7	34.7	21.6
Nb/Ta	9.5	10.3	15.3	10.3	14.3	19.1	9.8	5.2	10.5	17.3	129	27.1	15.1	93.6	116.9
Zr/Hf	31.3	54.6	43.7	51.6	41.2	36.7	53.1	83.9	65.4		166	38.9	50.0	49.4	38.4
Zr/Nb	18.2	1.59	2.65	0.24	0.40	0.24	0.65	1.65	16.90	2.56	0.42	5.94	3.24	2.39	4.84
La/Nb	1.62	0.36	0.96	1.07	0.90	0.91	1.33	7.13	6.93	35.2	2.86	1.36	0.70	0.24	1.35
Ba/Nb	3.18	0.99	4.66	6.88	1.45	1.67	1.75	45.6	22.8	182	4.03	11.7	2.92	1.75	50.1
Ba/Th	36.1	10.5	60.3	51.3	20.3	28.4	11.5	258	37.3	270	15.4	125	45.4	114	1199
Rb/Nb	0.21	0.22	0.56	0.42	0.20	0.24	0.11	4.18	0.27	1.33	0.01	0.68	0.46	0.12	3.27
K/Nb	77.4	83.5	129	128	80.4	79.5	34.8	1383	112	23.1	2.60	226	145	29.3	1260
Th/Nb	0.09	0.09	0.08	0.13	0.07	0.06	0.15	0.18	0.61	0.67	0.26	0.09	0.06	0.02	0.04
Th/La	0.05	0.26	0.08	0.13	0.08	0.06	0.11	0.02	0.09	0.02	0.09	0.07	0.09	0.06	0.03
Ba/La	1.96	2.74	4.86	6.44	1.61	1.84	1.32	6.40	3.29	5.17	1.41	8.62	4.15	7.22	37.2
Ce/Sm	10.3	7.93	17.8	22.9	18.9	21.3	21.6	13.5	14.5	19.1	31.6	13.4	12.8	9.01	14.8
Ce/Pb			24	76	80	90					12		39	33	
Sr/P	2.74	13.58	0.95	3.48	1.52	0.88	15.7	0.15	0.16	0.73	1.50	1.78	0.30	0.53	0.56
La/Sm	3.91	2.50	8.65	12.0	9.59	12.2	10.8	6.60	7.41	10.6	18.2	7.26	6.81	4.71	7.52
Zr/Sm	44.0	11.0	23.9	2.65	4.29	3.29	5.25	1.53	18.1	0.77	2.67	31.8	31.3	46.5	27.0
Hf/Sm	1.40	0.20	0.55	0.05	0.10	0.09	0.10	0.02	0.28		0.02	0.82	0.63	0.94	0.70
Sr/Nd	12.4	7.55	14.8	18.2	15.6	18.0	25.5	95.6	52.1	129	43.8	29.2	11.8	30.0	28.2
Sm/Nd	0.17	0.20	0.14	0.13	0.15	0.15	0.13	0.17	0.16	0.15	0.12	0.17	0.18	0.23	0.16
Nb/U	94.6	49.1	59.7	83.4	74.2	67.4	67.8	55.1	50.4	7.9	26.2	29.3	45.6	75.8	43.0
Y/Ho	27.5	27.8	27.5	20.7	23.3	23.9	22.6	26.1	26.0	26.5	27.8	28.0	27.9	28.7	27.7
Ba/Sr	0.11	0.18	0.41	0.55	0.15	0.19	0.07	0.08	0.07	0.06	0.07	0.36	0.43	0.26	1.63
Rb/Sr	0.01	0.04	0.05	0.03	0.02	0.03						0.02	0.07	0.02	0.11
Nb/Sr	0.03	0.18	0.09	0.08	0.10	0.11	0.04				0.02	0.03	0.15	0.15	0.03

Table 7C. Normalised trace element ratios

Sample ID:	C 90.29	C AG.122	TU 107	C 5.118	C 190.105	WKT	TUR II	C 49.40	C TL344	C DC	S W.58	C AG.48A	C JJ.14	TU 119	NC.153.82
	pyrox.	melteigite	mel dyke	mel.	mel.	mel.	mel.	carb.	carb.	dol carb.	carb.	ol melt.	ijolite	turrite	fente
Sr/Ta	0.52	0.11	0.33	0.24	0.26	0.31	0.43	5.85	5.99	90.7	13.7	1.62	0.19	1.16	6.68
Ta/Nb	1.88	1.73	1.16	1.73	1.24	0.93	1.82	3.40	1.70	1.03	0.14	0.66	1.18	0.19	0.15
Zr/Hf	0.84	1.47	1.18	1.39	1.11	0.99	1.43	2.26	1.76		4.48	1.05	1.35	1.33	1.04
Zr/Ti	2.28	2.05	1.93	0.42	0.72	0.56	0.92	1.24	4.75	15.8	18.9	3.05	4.32	7.34	6.37
Ta/Ti	3.75	35.5	13.5	48.7	35.4	33.8	41.1	40.8	7.61	102	98.7	5.37	25.0	9.31	3.19
Sr/Nd	0.78	0.47	0.93	1.15	0.98	1.13	1.60	6.00	3.27	8.07	2.75	1.84	0.74	1.89	1.77
Sr/P	12.4	61.4	4.31	15.7	6.88	3.96	70.92	0.67	0.70	3.30	6.77	8.06	1.36	2.38	2.52
Th/Ta	0.39	0.45	0.55	0.64	0.47	0.52	0.69	0.43	2.98	5.41	15.7	1.18	0.45	0.67	2.28
Th/Tb	3.14	9.46	11.90	29.7	13.4	12.4	25.1	2.07	8.56	3.02	24.9	6.50	8.58	2.40	2.51
Zr/Sm	1.70	0.43	0.92	0.10	0.17	0.13	0.20	0.06	0.70	0.03	0.10	1.23	1.21	1.80	1.04
Sm/Nd	0.54	0.60	0.44	0.40	0.45	0.47	0.41	0.53	0.49	0.46	0.37	0.52	0.55	0.72	0.51
Sm/Hf	0.50	3.46	1.28	13.5	6.69	7.78	7.04	38.2	2.52		43.4	0.85	1.11	0.74	0.99
Yb/Hf	0.05	0.23	0.07		0.12	0.16	0.04	3.42	0.18		2.31	0.14	0.13	0.25	0.24
Nb/La	0.61	2.72	1.03	0.92	1.10	1.08	0.74	0.14	0.14	0.03	0.34	0.73	1.40	4.06	0.73
Ti/Eu	0.99	0.28	0.71	0.39	0.33	0.33	0.36	0.06	0.19		0.01	0.58	0.39	0.28	0.20
Y/Ho	0.95	0.96	0.95	0.72	0.81	0.83	0.78	0.90	0.90	0.92	0.96	0.97	0.97	0.99	0.96
Ta/Yb	29.7	112	114		439	386	1487	21.7	16.0	2.52	10.1	13.6	58.8	6.84	2.16
K/Rb	0.94	0.96	0.58	0.76	1.03	0.84	0.79	0.83	1.05	0.04	1.03	0.83	0.78	0.60	0.96
Ba/Rb	1.41	0.42	0.76	1.49	0.67	0.65	1.44	0.99	7.73	12.5	58.0	1.56	0.57	1.31	1.39
Ce/Th	0.89	0.26	0.67	0.46	0.56	0.67	0.58	1.68	3.13	24.1	14.2	1.51	0.44	0.96	6.86
Rb/Sr	0.23	1.29	1.62	1.11	0.66	0.88	0.15	0.23	0.03	0.02		0.71	2.26	0.61	3.53
Th/P	9.28	261	7.31	42.1	12.4	6.60	113	0.05	0.35	0.20	7.72	5.90	3.23	1.38	0.86
Zr/Sr	1.17	0.54	0.44	0.04	0.08	0.05	0.05	0.01	0.10		0.01	0.35	0.90	0.68	0.30
Nb/Sr	1.03	5.40	2.65	2.41	3.06	3.43	1.27	0.05	0.10	0.01	0.53	0.94	4.45	4.55	0.98
Ce/U	4.88	0.68	1.43	2.05	1.59	1.29	2.20	9.72	8.25	6.03	1.58	0.89	0.73	0.43	1.38
Ce/P	21.3	150	8.84	30.3	14.6	8.44	94.5	0.19	0.37	0.88	6.98	7.50	3.16	1.97	2.59
Sr/Th	1.34	0.24	0.59	0.37	0.56	0.60	0.63	13.6	2.01	16.8	0.88	1.37	0.42	1.72	2.94
Zr/Nb	1.14	0.10	0.17	0.01	0.03	0.02	0.04	0.10	1.06	0.16	0.03	0.37	0.20	0.15	0.30
U/Lu	4.34	44.2	48.8	231	195	270	299	3.56	4.30	8.65	97.8	26.1	34.0	17.9	9.99
Zr/Y	11.5	2.79	6.85	1.52	1.74	1.23	2.65	0.32	3.64	0.18	0.61	6.05	5.94	3.98	3.40
Hf/Ti	2.70	1.39	1.63	0.30	0.65	0.56	0.64	0.55	2.69		4.22	2.91	3.21	5.52	6.15
Ce/Y	16.8	12.6	32.0	82.2	48.1	50.0	68.2	17.9	18.3	28.6	45.4	16.1	15.2	4.84	11.7
Th/U	2.12	1.19	1.18	2.86	1.35	1.01	2.62	2.49	7.87	1.35	1.75	0.70	0.75	0.30	0.46
Mo/Nd	0.10	0.21	0.15		0.18	1.48				0.21	0.07	0.48		0.37	0.36
Ce/Lu	21.2	30.2	69.6	473	311	350	656	34.6	35.5	52.2	155	23.3	24.9	7.61	13.8
Ce/Sm	2.49	1.92	4.31	5.54	4.58	5.16	5.24	3.26	3.50	4.63	7.65	3.26	3.10	2.18	3.59
La/Ce	0.98	0.81	1.26	1.36	1.31	1.49	1.29	1.27	1.33	1.44	1.49	1.40	1.37	1.35	1.31
Ce/Er	20.1	14.8	37.0	90.7	56.1	61.8	79.8	20.1	20.4	30.4	53.6	16.9	17.2	5.11	10.7
Sb/Sm	0.96	0.40	2.27	1.18	0.42	2.37	1.71	1.80	1.40	2.57	0.89	4.39	2.12	0.57	3.91
U/P	4.37	219	6.20	14.7	9.18	6.52	43.0	0.02	0.04	0.15	4.42	8.41	4.31	4.63	1.87

normalization constants that are used. However, in general, the large ion lithophile elements (LILE) are regarded as the most incompatible, followed by two of the most incompatible groups of high field strength elements (HFSE), namely W-Th-U, and Nb-Ta. These are then followed by the LREE, interspersed with Sr, P, Y, then the remaining HFSE group including Zr, Hf and Ti, and finally the HREE and Y.

We have already seen from the isotope data in Chapter 6 that the Turiy intrusion was not derived from a single mantle source, and has maintained open-system conditions during its magmatic evolution. The possibility of post-crystallization alteration, such as fenitization, hydrothermal activity, weathering and deuteritic alteration must also be considered. However, the isotopic data cannot be used to evaluate how and when these processes occurred. Trace element studies enable better characterization of these processes.

Spidergrams can be interpreted in different ways. These include:

a) different degrees of melting of the same source. This will have the effect of producing approximately parallel lines on a spidergram, with the degree of partial melting inversely proportional to the degree of trace element enrichment. The most incompatible elements will also show a greater degree of enrichment than the more compatible elements, i.e. the spidergram pattern will become tilted up to the left (Wilson, 1989). Rocks with parallel spidergram patterns, indicating derivation from a similar source, should also have similar isotopic signatures.

b) magma mixing. Recognition of mixing, whether in the mantle source or at various levels of emplacement within the crust, depends on the identification of the two (or more) magmatic end-members which contributed to the mixture. The spidergram pattern of a sample produced by mixing of two end-members will lie between the spidergram patterns of the two end-members, assuming closed-system conditions once mixing has taken place. The isotopic signature of such a mixed sample should also lie between those of the

end-members.

c) open-system fractional crystallization. Every mineral which crystallizes out of a melt retains certain proportions of each trace element, according to the partition coefficient for that particular mineral-melt pair under the prevailing P-T conditions. Unfortunately, too few partition coefficients are available for the types of mineral-magma combinations seen in the Turiy massif, to fully evaluate the effects that fractional crystallization will have on the trace element signatures. For example, garnet will tend to concentrate the HREEs along with Zr and Hf (Green 1994) and deplete the melt in these elements, while perovskite preferentially incorporates Nb. However, the additional effects of possible silicate-carbonate (or silicate-phoscorite) immiscibility, along with the effects of volatile-melt partitioning, make the effects of open-system fractional crystallization difficult to evaluate.

The following description and discussion of the trace element data obtained from the Turiy samples is divided into two parts. The first evaluates each trace element in turn, grouped according to similarities in chemical characteristics and incompatibility. The element abundances in the samples and relative abundances in various minerals (according to microprobe data listed in Appendix J and partition coefficients from the literature) are then used to evaluate the possible effects of any open-system fractional crystallization on the abundance of each of the elements. For example, given that apatite is the main phosphate mineral encountered in these rocks, it is assumed that variations in the P content of the magmas are due to apatite fractionation, and a Nb anomaly may be indicative of perovskite fractionation. It is important to understand the chemical and partitioning behaviour of each and every element in the spidergram, with respect to the mineralogy of the rocks found in the complex. The second part of this trace element study compares the spidergrams of the individual samples to assess the combined effects of open-system fractional crystallization, magma mixing, contamination, and degree of

partial melting at source, and whether any of the samples are genetically related. The combination of magma mixing, contamination and/or fractional crystallization adds an extra level of complexity to the interpretation of the spidergram.

In all spidergram plots (and some x-y discriminant plots) in this chapter the data have been normalized to primitive mantle (“pyrolite”) using the normalization values and element incompatibility order of McDonough & Sun (1995). Normalized ratios $(X/Y)_N$ have also been quoted throughout much of the text. While realizing that the use of normalized ratios is a somewhat unorthodox method of discussing the trends seen, it has been done to enable easier interpretation of the spidergrams which form the main method of graphical representation of the trace element data. In comparing the positive and negative anomalies, and overall slopes and characteristics of the spidergram patterns, one is essentially comparing normalised values with one another, and thus it seems logical to continue this comparison in the text. Non-normalized values of many trace element ratios are given in Table 7.B, while normalized ratios are listed in Table 7.C.

The data for the x-y discriminant plots have been divided into four groups, based on petrographic, mineralogical, isotopic and/or trace element similarities or relationships between the samples in each of the groups, although it should also be recognized that there can be overlap among the characteristics of the samples in the different groups. The members of Groups 2 and 3 were chosen on the basis of their inclusion in the melilitolite and carbonatite series of rocks, respectively. Group 4 includes the olivine melteigite, ijolite and turjite whose relationship to each other has already been discussed earlier in this chapter. The fenite is included in this group due its noticeable similarity in trace element signature with the turjite, as will be described in a subsequent section. The remaining silicate samples, all rich in pyroxene, form Group 1.

Group 1: pyroxenite, melteigite, melilitite dyke, (C.90.29, C.AG.122, TU 107)

Group 2: melilitolites (WKT, TUR II, C.5.118, C.190.105)

Group 3: carbonatites (S.W58, C.DC, C.TL.344, C.49.40)

Group 4: olivine melteigite, ijolite, turjite, fenite (C.AG.347, C.IJ.14, TU 119, NC.153.82)

7.3a Rare earth elements (REE)

Figure 7.3[a-d] shows pyrolite-normalized REE plots (McDonough & Sun, 1995) for the sample groupings listed above. All samples show varying degrees of LREE-enrichment, and none show any distinct positive or negative anomalies for any one element. Of the silicate rocks, the melilitolites (Figure 7.3b) show the steepest normalized REE patterns ($(\text{Ce/Lu})_N = 311\text{--}656$) and are the most enriched in LREE and depleted in HREE. The melilitite dyke [TU107], shown in Figure 7.3a, has the next highest value ($(\text{Ce/Lu})_N = 70$). The pyroxenite [C.90.29] and the melteigite [C.AG.122] are the only samples to have $(\text{Ce/La})_N > 1$, probably the result of pyroxene control (Green, 1994), and $(\text{Ce/Lu})_N$ values of 21 and 30 respectively. Thus, the overall slope of the REE plots has been characterized by Ce/Lu ratios, rather than La/Lu ratios, to avoid using the depleted La values from the pyroxenite and ijolite samples. The ijolite has $(\text{Ce/Lu})_N = 25$, similar to the value of 23 for the olivine melteigite [C.AG.48A], while the turjite [TU 119] has the lowest $(\text{Ce/Lu})_N$ value of all samples analyzed (7.6) (Figure 7.3d). The remaining silicate sample, the fenite [NC.153.82] has a $(\text{Ce/Lu})_N$ value of 14. The fenite and olivine melteigite [C.AG.48A] appear to have extremely similar REE contents and REE patterns. The fenite [NC.153.82] also has a parallel REE signature to the turjite [TU 119], including HREE, but at lower concentrations. It is possible that the fenitizing fluids which affected sample NC.153.82, were derived from the ijolite-series rocks within the Turiy massif, related to the olivine melteigite [C.AG.48A], ijolite [C.IJ.14] and turjite [TU 119] samples.

Figure 7.3a Mantle-normalized REE spidergrams for the Turiy pyroxenite [C.90.29], melteigite [C.AG.122] and melilitite dyke [TU 107]. Normalization factors from McDonough & Sun (1995).

Figure 7.3b Mantle-normalized REE spidergrams for Turiy melilitolites [C.5.118, C.190.105, WKT, TUR II]. Normalization factors from McDonough & Sun (1995).

Figure 7.3c Mantle-normalized REE spidergrams for Turiy carbonatites [C.49.40, C.TL.344, C.DC (dolomite-calcite carbonatite) and S.W58 (carbonatite dyke)]. Normalization factors from McDonough & Sun (1995).

Figure 7.3d Mantle-normalized REE spidergrams for Turiy olivine melteigite [C.AG.48A], ijolite [C.IJ.14], turjite [TU 119] and fenite [NC.153.82]. Normalization factors from McDonough & Sun (1995).

The carbonatites show two distinct REE patterns in Figure 7.3c. Sample S.W58, from the coastal carbonatite dyke, is the most REE-enriched sample of all of those analyzed. The central massif carbonatites have similar but surprisingly low REE contents, compared to the “average calcite carbonatite” of Woolley & Kempe (1989). The two central massif calcite carbonatites [C.49.40 and C.TL.344] both have $(\text{Ce/Lu})_N$ of 35, the dolomite carbonatite [C.DC] has a value of 52, while the coastal carbonatite dyke [S.W58] has a $(\text{Ce/Lu})_N$ of 154. The $(\text{Ce/Lu})_N$ value for “average calcite carbonatite” = 97 (Woolley & Kempe, 1989). In comparison with data for “average carbonatites” (Woolley & Kempe, 1989), S.W58 matches the REE pattern for “average calcite carbonatite” very closely, as will be discussed further in a later section. The central massif dolomite-calcite carbonatite [C.DC] is less enriched than but parallel to the REE pattern for “average Mg-carbonatite” of Woolley & Kempe (1989), and the remaining central massif calcite carbonatites have significantly lower REE contents than the “average calcite carbonatite” and also lower than the majority of the silicate rocks analyzed from Turiy. Total REE contents of the four carbonatite samples are 249 and 327 ppm [C.49.40 & C.TL.344 respectively], 341ppm [C.DC], and 3900ppm [S.W58]. It can be seen that the dolomite-calcite carbonatite [C.DC] is slightly more enriched in REE than the calcite carbonatites from the central complex, but that the south coast dyke contains more than ten times the REE concentration of the central massif carbonatites. Demaiffe et al. (1997) report values of 255-886 ppm for calcite carbonatites and 120ppm REE for magnesiocarbonatite from Kovdor. Dolomite-rich carbonatites generally tend to be richer in the REEs than calcite carbonatites within the same complex (Woolley & Kempe, 1989).

In terms of overall comparisons, the normalized Ce contents (chosen instead of La to avoid the La-depletion in the pyroxenite [C.90.29] and melteigite [C.AG.122]), decrease in the following order: carbonatite dyke > melilitolites > (ijolites, turjite,

melilitite) > central massif carbonatites > (fenite, olivine melteigite) > pyroxenite. The $(\text{Ce/Lu})_N$ ratios decrease in the order: melilitolites > carbonatite dyke > melilitite > [central massif carbonatites and melteigite] > [ijolite and olivine melteigite] > fenite > turjite.

There are some similarities between the REE data from Turiy and other complexes containing melilitolites. One example is the Iron Hill complex in Colorado. Its REE contents decrease in the order: [magnetite-ilmenite-perovskite segregations] > carbonatites > uncomphagrites (melilitolites) > ijolites > nepheline syenite > pyroxenite (Armbrustmacher & Futa, 1985), showing that, like Turiy, the melilitolites are the most REE-enriched of the silicate rocks within the Iron Hill complex. Comparison of the Ce/Sm and Ce/Y ratios from the Turiy bulk rock samples with REE data from mineral separates from melilitolites and carbonatites from the Oka intrusion (Eby, 1975), a plutonic melilite-bearing complex in Quebec, Canada, are plotted in Figure 7.4. It can be seen that most of the normalized Ce/Sm vs Ce/Y ratios for apatite, calcite, melilite and perovskite from Oka are all relatively similar to one another, while two samples of Ti-andradite from an Oka ijolite have much lower $(\text{Ce/Sm})_N$ vs $(\text{Ce/Er})_N$ ratios. The bulk-rock REE data for the Turiy samples falls between the Oka melilite-apatite-calcite-perovskite group and the Oka garnet data. In particular, the Turiy melilitolites are similar to the Oka melilite-apatite-calcite analyses in Figure 7.4, while the garnet-rich turjite [TU 119] falls closest to the Oka garnet analyses. Two points of interest from a comparison of the Oka and the Turiy REE data in Figure 7.4 are: 1) the Oka melilite separates have similar Ce/Sm ratios to the calcite, perovskite and apatite data, and 2) any combination of garnet with melilite, calcite, apatite or perovskite, will produce Ce/Sm - Ce/Y patterns which have a similar slope, during fractional crystallization.

The degree of partial melting and the composition of the minerals in a magma source will control the REE abundances in a melt but, during open-system fractional

Figure 7.4 Mantle-normalized Ce/Sm vs Ce/Y for the natrocarbonatite (OL 102) from Oldonyo L'Engai (Keller & Spettel, 1995), whole rock melilitolites and mineral separates from melilite-bearing rocks (except garnet separates from ijolite) from the Oka massif (Eby, 1975), and whole rock data from the Turiy massif. Normalization factors from McDonough & Sun (1995).

Turiy Group 1 = pyroxenite [C.90.29], melilitite dyke [TU 107], melteigite [C.AG.122]

Turiy Group 2 = melilitolites [C.5.118, C.190.105, WKT, TUR II]

Turiy Group 3 = carbonatites [C.49.40, C.TL.344, C.DC and S.W58]

Turiy Group 4 = olivine melteigite [C.AG.48A], ijolite [C.IJ.14], turjite [TU 119], fenite [NC.153.82].

Figure 7.5 Ratios of mantle-normalized REE data from mineral separates from the Oka massif (Eby, 1975). Minerals separates from melilite-bearing rocks, except the garnet which comes from an ijolite. Normalization factors from McDonough & Sun (1995).

crystallization (i.e. where the composition of the crystallizing melt may be altered by the addition of new material, or removal of part of the melt other than by fractional crystallization), the Ce/Yb ratio and the total REE content of each sample will be controlled by the REE partition coefficients for each of the individual minerals. It can be seen from Figure 7.5 that, other than the Eu anomalies, normalized perovskite/apatite and melilite/calcite plots of Oka REE data fall close to unity, signifying that each pair of minerals have similar REE patterns. Dawson et al. (1994) reported significantly greater REE concentrations in perovskite than apatite in a jacupirangite sample from Oldonyo L'Engai jacupirangite. The high REE concentrations in perovskite and apatite means that these minerals will be stronger buffers of the total REE content of a magma undergoing fractionation than calcite and melilite, but the REE slope produced by the buffering of the REE contents of a magma by any of these minerals will be similar (Figure 7.5). This may explain why world-wide carbonatite REE normalised plots are so similar (Loubet, 1972; Woolley & Kempe, 1989), given the abundance of these minerals in many carbonatites.

The Ti-andradite garnet REE data from Oka (Eby, 1975) do not have the typical HREE-enriched pattern that is so commonly predicted for (mantle) garnet, but show relative enrichment in the MREE instead. More recent data (Kjaarsgard, 1997), however, suggest that most experimentally-produced Ti-andradite garnet in trace element-doped experimental systems, contains typical HREE enrichment, or at least LREE depletion (Petibon et al., 1997). Thus it is still possible that the high HREE content of the turjite [TU 119] is due to accumulation of HREE-enriched garnet.

Sweeney et al. (1992) showed that the HREE partition coefficients for garnet-carbonate melt systems were 3-9 times higher than for garnet-silicate systems. Thus, if the garnet-rich turjite [TU 119] represents a fractionation product of a reaction between melteigite/ijolite and carbonatite (as suggested earlier), there should be a relationship between the HREE content of the turjite (based on its high garnet content) and the

proportion of carbonatite in the reacting magma. This reaction is obviously difficult to quantify directly, but a comparison may be made between the ijolite sample [C.IJ.14] which may be one of the end-members in the ijolite-carbonatite reaction, and which contains 25 modal % garnet, and the turjite sample [TU 119], the end-product of the reaction, which contains 55 modal % garnet. Both samples have identical LREE normalised values, but the turjite HREE normalised values are ~ 3 times greater than those from sample C.IJ.14 (Figure 7.3d). This implies an unusually large enrichment in HREE in the turjite sample, not dependent on the modal % of HREE-bearing garnet present. The turjite REE data from sample TU 119 is also similar to that of a bergalite from Kaiserstuhl (Hornig, 1988), which also exhibits disequilibrium textures (Dunworth, unpublished data), which might be indicative of silicate-carbonatite magma mixing, and/or reaction.

The low REE contents of the central massif carbonatites, in comparison to S.W58 and world-wide carbonatites, are unusual (Woolley & Kempe, 1989). Using starting materials reasonably close to presumed initial magma compositions at Turiy, Hamilton et al. (1989) showed that the REE partition into an immiscible calcium carbonate liquid at 5-6 kbar, but into a conjugate nephelinite liquid at 1 kbar. Increasing temperature and pressure and polymerization of the silicate liquid promotes partitioning of the REE into the carbonate liquid. The depth of emplacement of the Turiy massif is believed to correspond to an overlying pressure of 3 - 5 kbar (Bulakh & Ivanikov, 1984). If the depth of emplacement of the Turiy carbonatites is in the uninvestigated mid-range of the experiments of Hamilton et al. (1989) in terms of pressure, it is uncertain whether the REE will partition preferentially into a carbonatite liquid or its conjugate silicate liquid. The results of Hamilton et al. (1989), along with the moderate pressure at the depth of emplacement of the Turiy magmas, and the Si-poor (and hence depolymerized) silicate magmas found in Turiy, suggests that the REE could have remained preferentially within

the silicate liquids at Turiy if carbonate-silicate immiscibility has played a part in the evolution of the Turiy carbonatites. This could, perhaps, explain the low REE concentrations in the central complex carbonatites.

While it is common to focus on $\text{CO}_2\text{-H}_2\text{O}$ fluid conditions in a carbonatite environment, it is also important to consider other volatile constituents. F is an important constituent in carbonatite magmas, as shown experimentally by Gittins & Jago (1991), and both F and Cl are important constituents in the natrocarbonatites of Oldonyo L'Engai (Keller & Spettel, 1995). Little work has been done on REE complexing with the halogens in carbonatite systems, and very little is known about the pH of the volatile components of carbonatite magmas (D. Pyle, pers comm 1997). Haas et al. (1995), in a study of REE behaviour in hydrothermal systems, determined pH conditions under the influence of Cl^- (acidic), F^- (neutral) and OH^- (basic) and the effect of pH in a halogen-rich fluid on REE complexing. It was found that LREE preferentially form chloride complexes, HREE preferentially combine with F^- , and that there is no difference in behaviour when OH^- is the dominant anion. Although these results may not appear to be directly applicable to a carbonatite massif, the data set of Haas et al. (1995) at least serves as a first approximation, given that the eruption temperatures of the carbonatites at Oldinyo L'Engai are only 500-600°C (Dawson et al., 1995). If REE-halogen complexing has affected the REE signatures of the Turiy magmas, the F-Cl-OH conditions in the Turiy magmas must be constrained. This is a difficult task, as the only information available is the F-Cl-OH contents of volatile-rich minerals such as apatite, amphibole, cuspidine, fluorite and mica. F is found in apatite and mica in most Turiy magmas, but fluorite is found only in the fenites and metasomatic vein rocks such as TD 1. Trace quantities of a number of mostly unidentified F-rich minerals also occur in the ijolites (Appendix Ji), while F-rich cuspidine occurs in trace amounts in the eastern and southern massif melilitolites (Appendix Jix). Titanite, which occurs in the olivine melteigite,

pyroxenites and ijolites, also contains small amounts of F. Within the apatites, there is an average decrease in F content from fenite (4 wt.%) > ijolite (~2.9 wt.%) > melilitolite (~2.7 wt.%) > [phoscorite = carbonatite (~2.1 wt.%) > pyroxenite (~1.8 wt.%). Trace amounts of Cl (0.3 wt.%) were found in apatite from a central massif melilitolite [C.5.118] and phoscorite [C.19.163], while apatite from sample N.63.4, a pyroxene-calcite carbonatite related to the northern complex melilitolites, contains < 1.8 wt.% Cl. The micas generally have low F contents (Appendix Jxiii). The highest F values occur in micas from the olivine melteigite [C.AG.48A] and from the melteigite [C.AG.122] at 2-3 wt.%. Fluorine values in mica from the turjite chilled margin sample [TQ] are also consistently around 1 wt.% F, along with F contents in micas from some of the more evolved central massif carbonatites. The only mica containing Cl occurs in sample TD2 [aegirine-apatite vein], where Cl is the dominant occupant of the OH site. Cl is also found in cancrinite in sample TUR I [turjaite].

Evaluation of the partitioning behaviour of the halogens implies that F is probably concentrated in the final fluids emanating from the intrusion (i.e. the fenites), that the melteigite-ijolite magmas are richer in F than the melilitolites, while the phoscorites and carbonatites have lower and more variable F contents. Cl is found occasionally in melilitolite minerals, never in ijolite minerals, and in greater abundance in the fenites. If the results of the study by Haas et al. (1995) can be correlated to the F-Cl-OH contents of the minerals within the Turiy samples, then the ijolites, melteigite, fenite, and possibly turjite, should have more HREE-enriched signatures, as opposed to the melilitolites which should be more LREE-enriched. This is, in fact, what is seen. It is possible, therefore, that REE-halogen partitioning has played a role in shaping the REE signatures of the Turiy samples. Two points of note, should, however, be added here. The first is that the F contents listed in Appendix J are minimum values due to F-detection problems with the electron-microprobe during part of this study (data for the samples were

measured for 25 seconds or to 0.5% precision, whichever was attained first). The second is that there is a natural $F\text{-Fe}^{2+}$ avoidance principle in natural micas (Speer, 1984), but given the Mg-rich nature of most of the samples from Turiy, this F-Fe avoidance should not have too great an effect on the partitioning behaviour of fluid species between the micas and their host melts.

Fig. 7.6 shows a plot of absolute abundances of Nd vs Sr for Turiy samples based on the whole-rock ICP-MS and mineral separate data from isotope dilution analyses obtained in this study. Pyroxene, phlogopite and nepheline all have low Sr and Nd concentrations, while apatite and perovskite have the highest combined Sr and Nd contents. Garnet, amphibole and calcite all have moderate Nd contents, but garnet and amphibole have low Sr and calcite has high Sr concentrations. It can be seen that the Nd/Sr ratio of the central massif carbonatites (cluster of three red triangles) is very similar to the calcite values, while the carbonatite dyke [S.W58] (solitary red triangle) has higher combined Sr and Nd contents than any of the other mineral or whole rock data in Figure 7.6. The incompatibility of Sr and Nd are similar (McDonough & Sun, 1995) and hence the Sr/Nd ratios of most of the minerals and whole rocks in Figure 7.6 are relatively constant. The Sr-Nd signatures of the melilitolites are probably controlled by the Sr-Nd contents of perovskite and/or apatite. REE data from mineral separates from the Oka complex (Eby, 1975), along with Sr microprobe data from this study, suggest that melilite may also be rich in REEs and Sr, and hence influence the bulk rock signature under open-system conditions, but melilite was not separated for trace element or isotopic analysis in this study. The pyroxenite Sr-Nd concentrations are controlled by pyroxene and perovskite, and the Sr-Nd signature of the melteigite [C.AG.122] and ijolite [C.IJ.14] fall between the Sr-Nd values of pyroxene, garnet, nepheline and apatite, consistent with their modal mineralogy. Interestingly, the Sr-Nd concentrations of the turjite [TU 119] (3350, 112) (represented by the red circle partially covering melilitolite data (red squares) in

Figure 7.6 Sr and Nd concentration data from mineral separates (obtained by isotope dilution analysis) and whole rock data (obtained by ICP-MS analysis) from the Turiy massif. The hatched lines represent the extremes of Sr and Nd content in the minerals represented by phlogopite - garnet - perovskite - apatite - calcite apices.

Turiy Group 1 = pyroxenite [C.90.29], melilitite dyke [TU 107], melteigite [C.AG.122]

Turiy Group 2 = melilitolites [C.5.118, C.190.105, WKT, TUR II]

Turiy Group 3 = carbonatites [C.49.40, C.TL.344, C.DC and S.W58]

Turiy Group 4 = olivine melteigite [C.AG.48A], ijolite [C.IJ.14], turjite [TU 119], fenite [NC.153.82].

Figure 7.6) fall close to the melilitolites and between two of the samples proposed as reactants in its formation (C.IJ.14 [ijolite] (1190, 101) and S.W58 [carbonatite dyke] (22400, 511)). The silicate whole-rock analyses fall between the extremes in Sr-Nd space defined by phlogopite - [garnet-amphibole] - perovskite - apatite - calcite, and no clear mixing trends or identifiable mineral control can be defined for the silicate samples, assuming crystallization under open-system conditions.

7.3b Light ion lithophile elements (LILE)

The LILE group includes K, Cs, Rb, Ba, Sr and Pb. At Turiy, K and Rb are typically found in micas and nepheline; other K and Rb-bearing minerals include amphibole which is rare, and feldspar which is non-existent in the samples investigated other than the fenite. $(K/Rb)_N$ ratios show little variation among the different samples. All range between 0.6 - 1.0. Ba may be found in mica, nepheline, celsian (in TU 120A [turjite]) and in trace occurrences of barite and Ba-carbonate. $(Ba/Rb)_N$ ratios show variation in the silicate rocks (0.4-1.6) but these are generally lower than those in the carbonatites (1 - 58). Pb is present in < 10ppm levels in all of the samples analyzed except for the carbonatite dyke [S.W58] which contains 163 ppm. No galena or other Pb minerals were found in any of the silicate or carbonatite samples, despite an intensive search in both hand specimens and thin sections.

Spidergrams involving a wide variety of elements normalised to primitive mantle, including the LILE, are plotted in Figure 7.7[a-d]. The division of the samples into four groups, as defined in section 7.3, and used to divide up the REE data plots in Figure 7.3, is also supported by the data plotted in Figure 7.7[a-d]. In terms of inter-sample similarities, the LILE are the most varied compared to the other trace element groups, but, given their high level of mobility and incompatibility, this is hardly surprising. There is a strong correlation in all samples between abundance of mica (and/or nepheline)

Figure 7.7a Mantle-normalized spidergrams for Turiy Group 1 = pyroxenite [C.90.29], melilitite dyke [TU 107], melteigite [C.AG.122]. Normalization factors from McDonough & Sun (1995). See text for discussion of the spidergram patterns.

Figure 7.7b Mantle-normalized spidergrams for Turiy Group 2 = melilitolites [C.5.118, C.190.105, WKT, TUR II]. Normalization factors from McDonough & Sun (1995). See text for discussion of the spidergram patterns.

Figure 7.7c Mantle-normalized spidergrams for Turiy Group 3 = carbonatites [C.49.40, C.TL.344, C.DC and S.W58]. Normalization factors from McDonough & Sun (1995). See text for discussion of the spidergram patterns.

Figure 7.7d Mantle-normalized spidergrams for Turiy Group 4 = olivine melteigite [C.AG.48A], ijolite [C.IJ.14], turjite [TU 119], fenite [NC.153.82]. Normalization factors from McDonough & Sun (1995). See text for discussion of the spidergram patterns.

and the quantities of K and Rb present, suggesting a degree of open-system fractionation, or a K-rich source. This effect is most noticeable within the carbonatites. Sample C.49.40 [calcite carbonatite] contains a significant amount of mica and hence has relatively high K and Rb concentrations while C.DC [dolomite-calcite carbonatite] contains no mica and its K and Rb concentrations are extremely low. $(K/Rb)_N$ ratios for all samples cluster between 0.5-1 except for C.DC which is close to zero. $(Ba/Rb)_N$ values fall between 0.5-1.5 except for three of the carbonatites which have significantly higher values (Figure 7.8a).

7.3c High field strength elements (HFSE)

The HFSE are divided into three groups: W-Th-U, Nb-Ta and Zr-Hf-Ti based on ionic charge and relative incompatibility. They are described below in order of increasing compatibility.

7.3ci W, Th & U

The host minerals for U, Th and W are difficult to identify in the Turiy rocks, but it is believed that garnet may play host to U. Other typical U-bearing minerals include zircon and pyrochlore. Less than five zircon grains ($\sim 5 \mu\text{m}$ diameter) have been found in all of the samples investigated, and (urano-) pyrochlore has only been found in one carbonatite from the central complex, so neither of these is expected to buffer the U and Th concentrations in the majority of the rocks.

For the carbonatites, melilitolites, pyroxenite and melteigite from Turiy, the $(Th/U)_N > 1$. The turjite, ijolite, fenite and olivine melteigite all have $(Th/U)_N < 1$, as does the natro-carbonatite from Oldonyo L'Engai (Keller & Spettel, 1995).

Hauri et al. (1994) showed that in a mantle garnet-clinopyroxene-basalt system, Th was preferentially incorporated into clinopyroxene, while U was preferentially

Figure 7.8 X-Y discriminant plots using mantle-normalized element values. Normalization factors from McDonough & Sun (1995). Dotted lines are 1:1 ratios.

Turiy Group 1 = pyroxenite [C.90.29], melilitite dyke [TU 107], melteigite [C.AG.122]

Turiy Group 2 = melilitolites [C.5.118, C.190.105, WKT, TUR II]

Turiy Group 3 = carbonatites [C.49.40, C.TL.344, C.DC and S.W58]

Turiy Group 4 = olivine melteigite [C.AG.48A], ijolite [C.IJ.14], turjite [TU 119], fenite [NC.153.82].

incorporated into garnet. Assuming that the partitioning effects determined by Hauri et al. (1994) are also applicable to garnet - pyroxene - silicate melt systems at crustal pressures, it is likely, given their modal mineralogies and $(\text{Th}/\text{U})_{\text{N}}$ ratios, that the Th-U signatures of the turjite and ijolite are controlled by fractional crystallization of garnet, and the Th-U signatures of the pyroxenite [C.90.29] and melteigite [C.AG.122] are controlled by fractional crystallization of pyroxene, possibly under open-system conditions, as defined earlier in this chapter. The partitioning data do not appear to apply to the olivine melteigite which contains abundant clinopyroxene and no garnet, but has $(\text{Th}/\text{U})_{\text{N}} < 1$.

Another Th-U mineral site in the Turiy rocks is apatite. A plot of whole-rock $(\text{Th}/\text{P})_{\text{N}}$ vs $(\text{U}/\text{P})_{\text{N}}$ (Figure 7.8b) shows an approximate 1:1 positive correlation. The central massif carbonatites have both $(\text{Th}/\text{P})_{\text{N}}$ and $(\text{U}/\text{P})_{\text{N}} < 1$ and show an enrichment in Th over U, while the melilitolites are spread out along the 1:1 line with slight enrichment in Th. Samples C.IJ.14 [ijolite] (4.3, 3.2) and TU 119 [turjite] (4.6, 1.4) are the only samples to be slightly enriched in U in Figure 7.8b, presumably due to garnet fractionation. The varied Th/U ratios of all the samples (1.2 - 30.8) compared to the strong positive correlation between the $(\text{Th}/\text{P})_{\text{N}}$ and $(\text{U}/\text{P})_{\text{N}}$ values of the samples seen in Figure 7.8b, indicates that either apatite is buffering the Th-U abundances in these rocks, or that the three elements exhibit similar chemical behaviour in these magmas. Another interesting relationship is seen in Figure 7.8c between $(\text{P}/\text{Ce})_{\text{N}}$ vs $(\text{U}/\text{Ce})_{\text{N}}$. A negative relationship occurs in the carbonatites, with $(\text{P}/\text{Ce})_{\text{N}}$ decreasing and $(\text{U}/\text{Ce})_{\text{N}}$ increasing in the order C.49.40 - C.TL.344 - C.DC [dolomite-calcite carbonatite] - S.W58 [carbonatite dyke]. Conversely, the silicate rocks increase in $(\text{P}/\text{Ce})_{\text{N}}$ from the melilitolites (Group 2 in Figure 7.8c), to the olivine melteigite-ijolite-fenite-turjite group of samples (Group 4 in Figure 7.8c). The pyroxenite [C.90.29] and melteigite [C.AG.122] fall outside the trend created by the samples from Groups 2 and 4 in Figure 7.8c. WKT [turjaite], S.W58

[carbonatite dyke] and TU 107 [melilitite] (hidden by the former data points in Figure 7.8c) have virtually identical $(P/Ce)_N - (U/Ce)_N$ values. The data essentially fall into three quadrants based on rock type: a) $(P/Ce)_N > 1$ and $(U/Ce)_N < 1$ = central massif carbonatites; b) $(P/Ce)_N < 1$ and $(U/Ce)_N < 1$ = pyroxenite, melilitolites, melilitite, carbonatite dyke and the fenite, and c) $(P/Ce)_N < 1$ and $(U/Ce)_N > 1$ = melteigite, ijolite, turjite and olivine melteigite. The enrichment in U in the latter is presumably due to open-system fractionation of garnet. The variation in $(P/Ce)_N$ ratios in Figure 7.8c also has implications for magmatic fractionation processes. Because there are no known mechanisms for strongly fractionating Ce from Nd during apatite formation, and given that Figure 7.6 demonstrated that the Nd contents in Turiy apatites only vary by a factor of 10, it is likely that the Ce contents of the apatites will also be relatively constant throughout the massif. Therefore, the wide range of $(P/Ce)_N$ values ($\sim 0.01 - 10$) seen in the bulk-rock samples in Figure 7.8c indicates that the REE signature of the bulk rocks is not solely controlled by the presence of apatite.

W is generally insignificant in all rock types from Turiy (< 1 ppm) other than the turjite (5.3 ppm). (N.B. all samples were crushed in an agate mill, not tungsten carbide, and thus W values are believed to be real). Data from Oldonyo L'Engai (Keller & Spettel, 1995) show that the natrocarbonatite magmas are unusually enriched in W (~ 48 ppm), while all other rock-types are below the detection limit. The experiments carried out by Verwoerd (1978), whose reaction products are similar to the turjite, involved the reactions of a natrocarbonatite - calcite-carbonatite mixture with an ijolite. Although it is also possible to produce many of the trace element characteristics of the south coast turjite dyke [TU 119] by reacting a magma such as C.IJ.14 [ijolite] with a calcite carbonatite such as S.W58 [south coast carbonatite dyke] (Figure 7.9), this still provides no explanation for some of the LILE trace element signatures of the turjite including the W-Th-U spidergram pattern, and the Cs signature (not shown). Both the W content and

the $(U/Th)_N$ ratio of the turjite are the highest of all samples analyzed, and cannot be produced by simple magma mixing of any of the analyzed samples. Given the inclusion of natrocarbonatite as a reactant in the ijolite-carbonatite experiments of Verwoerd (1978), it seems natural to examine the effect that a natrocarbonatite might have on the trace element characteristics of the turjite, a natural product of those reactions. While realising that there are no confirmed occurrences of natrocarbonatite in the Turiy intrusion, the presence of a peralkaline nephelinite in the south coast dyke suite (Dunworth, unpublished data) bearing similarities to the peralkaline nephelinites found at Oldonyo L'Engai, suggests that the possibility that components involved in the genesis of natrocarbonatite may have been involved in the genesis of the Turiy magmas cannot yet be ruled out. Since there is only one true natrocarbonatite so far identified on Earth, trace element data from the Oldonyo L'Engai natrocarbonatite (sample OL 102 from Keller & Spettel, 1995) will be used for comparative purposes in this study. Figure 7.9 suggests that the interaction of the Turiy ijolite sample [C.IJ.14] (containing low W and $(U/Th)_N > 1$), with the source rocks or crystallizing minerals which control the W-Th-U (high W and $(U/Th)_N$) and Cs and Ba signatures of sample OL 102 would be effective in producing the part of the turjite spidergram involving this group of elements.

7.3cii Nb & Ta

Primitive mantle normalized ratios of Nb and Ta both range from 4 - 594 with melilitolites = melteigite and ijolite = carbonatite dyke > melilitite > pyroxenite = olivine melteigite > central massif carbonatites (Figure 7.7). $Nb < Ta$ (normalized values) in all samples except the carbonatite dyke [S.W58], the turjite [TU 119], olivine melteigite [C.AG.48A] and fenite [NC.153.82] in which $Nb > Ta$. Figure 7.8d, a plot of $(Nb/Ti)_N$ vs $(Ta/Ti)_N$, shows the approximate 1:1 relationship between these ratios for most of the samples. The carbonatite dyke [S.W58], turjite [TU 119] and fenite [NC.153.82] lie to the

Nb-rich side of the 1:1 line, while carbonatite sample C.TL.344 lies furthest away from the line on the Ta-rich side. Nb-Ta-bearing minerals found at Turiy include ilmenite, magnetite, perovskite, rutile, titanite and rare pyrochlore. Titanite occurs in trace quantities in the ijolitic rocks, while rutile occurs only in the late-stage vein samples [TD 1 & TD 2]. Green & Pearson (1987), in a study of HFSE partition coefficients, determined that $D_{\text{Nb}}/D_{\text{Ta}}$ increased in the order titanite < rutile < ilmenite < Ti-magnetite, although $\text{Ta} > \text{Nb}$ in all these minerals. In the Turiy rocks, Ta and Nb have not been detected in any silicate phases, but perovskite contains up to 6.6wt.% Nb_2O_5 and minimal Ta in the ijolite sample [C.IJ.14]. Pyrochlore from carbonatite sample C.100.y contains 42-52 wt.% $\text{Nb}_2\text{O}_5 + \text{Ta}_2\text{O}_5$ with $\text{Nb}/\text{Ta} = 4-12$. Thus, most common Nb-Ta-bearing minerals preferentially incorporate Ta over Nb, except perovskite and pyrochlore which prefer Nb. The primitive mantle Nb/Ta ratio according to McDonough & Sun (1995) is 17.8. Under open-system conditions involving fractional crystallization, perovskite is the only mineral found thus far in Turiy with $\text{Nb}/\text{Ta} > 17.8$ that can produce a negative Nb-Ta slope on the mantle-normalized spidergram plots. It is also believed (Rudnick et al., 1993; Green, 1995) that metasomatic carbonate fluids preferentially incorporate Nb with respect to all other HFSEs, and that an enriched Nb-signature is a sign of carbonate metasomatism. Experimental data from immiscible silicate-carbonate melts (Green, 1994) indicate that Nb preferentially concentrates into the carbonate liquid, leaving the other HFSEs in conjugate silicate liquids or mineral phases. The carbonatite dyke [S.W58] has $(\text{Nb} > \text{Ta})_{\text{N}}$ consistent with liquid immiscibility, and typical of the “average calcite carbonatite” of Woolley & Kempe (1989). However the $(\text{Nb}/\text{Ta})_{\text{N}}$ ratios of the central massif calcite carbonatites are not consistent with derivation by immiscibility, as samples C.49.40 and C.TL.344 have $(\text{Ta}/\text{Nb})_{\text{N}} > 1$ and absolute values significantly lower than “average calcite carbonatite” of Woolley & Kempe (1989) for both elements. The

Figure 7.9 Mantle-normalized spidergram for the Turiy ijolite [C.IJ.14], turjite [TU 119], carbonatite dyke [S.W58], and natrocarbonatite [OL 102] from Oldonyo L'Engai (Keller & Spettel, 1995). Normalization factors from McDonough & Sun (1995).

dolomite-calcite carbonatite [C.DC] has $(Ta = Nb = 5)_N$ which again does not correspond to the $(Nb > Ta)_N$ ratio and high normalized Nb and Ta concentration values (~ 900) in the “average magnesium carbonatite” of Woolley & Kempe (1989). The turjite has $(Nb > Ta)_N$, almost identical to the signature of the carbonatite dyke [S.W58]. This could be a good argument for a carbonatite component dominating the Nb-Ta signature of the turjite, if turjite indeed represents an ijolite - carbonatite reaction product.

Comparing the behaviour of elements like Sr with the incompatible HFSE may help to constrain the effect of open-system fractionation of phases such as perovskite and titanite on the trace element compositions of the magmas. Systematic variations in $(Nb/Sr)_N$ occur between different rock types. The melteigite, ijolite and turjite [C.AG.122, C.IJ.14 and TU 119] have the highest ratios (4.5-5.4). These are followed by three of the melilitolite and the melilitite dyke [C.5.118, C.190.105, WKT and TU 107] with values between 2.4 - 3.4. The three samples that are richest in olivine, pyroxene and melilite respectively ((olivine melteigite, C.AG.48A), pyroxene (pyroxenite, C.90.29) and melilite (melilitolite, TUR II)) have values of 1.0 - 1.3. The fenite has a $(Nb/Sr)_N$ value of 0.9. The $(Ta/Sr)_N$ ratios are more varied than the $(Nb/Sr)_N$ ratios, but show tight clusters within the melilitolite & melilitite group (2.3-4.2) and carbonatites (0.01-0.17). The highest values occur in the melteigite [C.AG.122] ($(Nb/Sr)_N = 9$) and ijolite [C.IJ.14] ($(Nb/Sr)_N = 5.3$).

A comparison of the behaviour of the LILE and most incompatible HFSE can be made using the Th/Ta ratio. Variations in $(Th/Ta)_N$ are minor within the majority of the silicate rocks and carbonatite sample C.49.40 (0.4 - 0.7). The olivine melteigite [C.AG.48A] rises to 1.2, and the fenite [NC.153.82] to 2.3. The greatest variations are seen in the remaining three carbonatites with $(Th/Ta)_N$ ratios 3.0 - 15.7. Very little variation is seen in $(Nb/La)_N$ ratios among the different rocks. All samples vary between 0.03 - 1.0 with three exceptions. The ijolite [C.IJ.14] has a value of 1.4, the melteigite

[C.AG.122] a value of 2.7, and the turjite [TU 119] has the highest value of 4.1.

7.3ciii Zr, Hf, Ti

The remaining HFSE (Zr, Hf, Ti), more compatible than either W, Th and U, or Nb and Ta, show some interesting variations in the Turiy samples. Of particular note, is the behaviour of Zr and Hf in terms of their enrichment in the turjite, pyroxenite and melilitite dyke, their significant depletion in the melilitolites, ratios of $(\text{Zr/Hf})_N$ close to unity in all silicate samples, and fractionation ($\text{Zr} \gg \text{Hf}$) in the carbonatites. Ti concentrations are relatively constant in the silicate rocks, and constant, but with a much lower value, in the carbonatites.

A plot of $(\text{Hf/Ti})_N$ vs $(\text{Zr/Ti})_N$ (Figure 7.10) shows a 1:1 linear relationship for the silicate Turiy samples. Zr and Hf can be fractionated within respect to each other by carbonatite immiscibility (Green 1994; Pettibon et al., 1997) and pyroxene fractionation (Hart & Dunn, 1993). Figure 7.10 reveals that the melilitolites (Group 2) are the only samples with both ratios < 1 . All other samples, except carbonatite sample C.49.40, have both ratios > 1 . Partition coefficient measurements show that melilite excludes Zr (Nagasawa, 1980), but the behaviour of Hf in melilite is unknown. Given the similar Zr and Hf concentrations in the melilitolites, it is likely that melilite excludes both Zr and Hf. There are few other minerals in the melilitolites that could include these elements during open-system fractional crystallization. Sample TUR II, for example, contains only melilite, nepheline, magnetite, perovskite and phlogopite. With the exception of perovskite, none of these minerals is likely to retain Zr or Hf, and thus there must be a residual cumulate, melt or fluid from the melilitolites which contains these two elements. The turjite and pyroxenite are the only samples analyzed which show significant enrichment in Zr and Hf with respect to similarly incompatible trace elements, and the two elements are probably controlled by garnet and pyroxene respectively. Analyses in

Figure 7.10 X-Y discriminant plot using mantle normalized element values. Normalization factors from McDonough & Sun (1995). Dotted line represents 1:1 ratio.

Turiy Group 1 = pyroxenite [C.90.29], melilitite dyke [TU 107], melteigite [C.AG.122]

Turiy Group 2 = melilitolites [C.5.118, C.190.105, WKT, TUR II]

Turiy Group 3 = carbonatites [C.49.40, C.TL.344, C.DC and S.W58]

Turiy Group 4 = olivine melteigite [C.AG.48A], ijolite [C.IJ.14], turjite [TU 119], fenite [NC.153.82].

Appendix Jxi show significant Zr contents in some of the garnet from the ijolite (and turjite) series rocks, whereas the trace amount of garnet produced in the melilitolites by the breakdown of melilite or its reaction with late-stage fluids are Zr-free.

Comparison of Zr and Hf values with other elements within the spidergram provides some interesting correlations. $(\text{Sm}/\text{Zr})_N$ and $(\text{Sm}/\text{Hf})_N$ ratios lie between 0.5 - 1.5 for many of the silicate samples. Comparison of the REE data between the different samples suggests that the higher $(\text{Sm}/\text{Zr})_N$ and $(\text{Sm}/\text{Hf})_N$ ratios of between 5.0 and 13.6 in the melilitolites are mostly due to the depletion of Zr and Hf. The melteigite [C.AG.122] ratios are 2.32 and 3.46 respectively, which is slightly higher than the values from the majority of the silicate samples, but not as high as the values from the melilitolites. The carbonatites, other than sample C.TL.344, have high ratios (10 - 44).

$(\text{Zr}/\text{Sr})_N$ ratios for all rock types are < 1 , except for the pyroxenite sample [C.90.29] which is marginally higher. $(\text{Zr}/\text{Nb})_N$ ratios follow a similar pattern, with all values < 0.4 except the pyroxenite [C.90.29] and carbonatite sample C.TL.344 which have values slightly over 1.

One of ratios to show the greatest variations in the Turiy rocks is $(\text{Zr}/\text{Y})_N$. Y is similar to the HREE in terms of incompatibility, and, as result, is relatively depleted in the melilitolites, compared to other rock types. The $(\text{Zr}/\text{Y})_N$ ratios for the melilitolites range from 1.2 - 2.7. The melteigite, ijolite and turjite values range from 2.8 - 5.9, while the olivine melteigite and melilitite dyke have values of 6.1 and 6.9 respectively. The pyroxenite has the highest value at 11.5, and the fenite has a value of 3.4. The carbonatite values are generally low (0.2 - 0.6), except for sample C.TL.344 which has a value of 3.7.

There is a wealth of experimental partitioning data available in the literature for Zr and Hf, involving both variations in melt composition (carbonate-silicate) and common mineral phases (e.g. garnet, pyroxene, oxides). Sweeney et al. (1992) found

that Zr partitioned into garnet in carbonate melts, but not silicate melts. Hart & Dunn (1993) showed that clinopyroxene preferentially excludes Zr, with respect to Hf and Ti. Only five microcrystalline zircon grains have been found in all of the samples analyzed from the massif, and baddelyite, zirsinalite and calzirtite have each been found in one carbonatite. These trace minerals, which are not always in textural equilibrium with their host rock (e.g. baddelyite overgrowths on an apatite vein in sample C.6.15 [carbonatite]) are not expected to be dominant in controlling the Zr contents of most Turiy magmas. Perovskite is another mineral that can also contain Zr (Mitchell & Steele, 1992).

Nielsen et al. (1994) found that the D_{Zr} was inversely proportional to the Al concentration in magnetite, but is directly proportional to the Fe/Mg ratio in magnetite, and that D_{Ti} of magnetite is not affected by magnetite composition. The negative correlation between Al - HFSE was attributed to the immiscibility gap in the spinel series that occurs between $(Fe,Mg)_2TiO_4$ - $(Mg,Fe)Al_2O_4$ which also marks a structural change between normal-inverse spinels (Waychunas, 1991). The melilitolite magnetites from Turiy are relatively rich in Al (Chapter 5.3ai; Appendix Jxvi) and thus are likely to exclude Zr.

There are two explanations for the unusual depletion of Zr and Hf in the melilitolites. The first assumes that the melilitolite magmas were already depleted in Zr and Hf and the present contents simply reflect the magma source, and/or parent magma, and the second, that the melilitolite magmas originally contained higher concentrations of Zr and Hf which were lowered during fractionation of the melilitolitic melt. The pyroxenite [C.90.29] has similarly enriched Zr and Hf signatures to those of the melteigite [C.AG.122] and ijolite [C.IJ.14], although the pyroxenite has $(Zr/Hf)_N < 1$ quite different to any of the other silicate or carbonatite samples. If the pyroxenite represents a cumulate from an as yet unidentified residual liquid, then this would indicate that the pyroxenite preferentially retained Hf, which is consistent with Zr and Hf partition

coefficients for pyroxene (Hart & Dunn, 1993). Such open-system behaviour would produce a residual magma poor in Zr and Hf, and with $(Zr/Hf)_N$.

The spidergrams for sample TU 107 [melilitite dyke] and sample C.5.118 [central massif melilitolite] (Fig. 7.11a) are approximately parallel, except for the Zr and Hf data which show depleted signatures in sample C.5.118. Given their overall similarity in trace element compositions (along with SiO_2 content and Mg#, Table 7.A) it is likely that the two magmas share a similar magma source, and that Zr and Hf (along with the HREE) were possibly removed from the melilitolite during open-system fractional crystallization, either into a cumulate product (garnet or pyroxene-rich) or into a residual liquid (e.g. carbonatite or possibly phoscorite). Hornig (1988) reported high Zr contents in phoscorites from Kaiserstuhl, and attributed this to the presence of baddelyite and zircon inclusions. The two central massif calcite carbonatites [C.49.40, C.TL.344] from Turiy (which are transitional to phoscorites, in terms of modal mineralogy but not isotopic composition) show $Zr > Hf$, but both elements are still depleted with respect to Sm and Ti (the neighbouring elements on the spidergram, defining the boundaries of relative incompatibility of Zr and Hf). Green (1994) found that Hf partitioned very strongly into the silicate member of conjugate immiscible silicate-carbonate liquids. This fits with the $Zr \gg Hf$ pattern seen in all four carbonatites analyzed, but there appears to be no silicate magmas with $(Zr/Hf)_N < 1$, unless the pyroxenite [C.90.29] represents a cumulate from such a melt. The nature of the Zr-Hf pattern which is produced in a carbonatite which has *fractionated* from a silicate parent is unclear at this time, and thus it is difficult to compare the possibilities of fractionation versus immiscibility, in order to explain the petrogenesis of the Turiy carbonatites.

Rudnick et al. (1993) suggested that Zr enrichment within a mantle melt could occur as a result of carbonate metasomatism within the mantle source, and that a decreasing Ti/Eu ratio within any mantle melt was also indicative of carbonate

metasomatism. Figure 7.8e shows a direct correlation between increasing $(\text{Zr/Hf})_N$ and decreasing $(\text{Ti/Eu})_N$ with increase in carbonate content of the Turiy samples. Andrade et al. (1997), however, in a study of Zr/Hf ratios in carbonatites, have warned against using this ratio as a reliable guide to carbonatite metasomatism. Although there are variations in $(\text{Zr/Hf})_N$ and $(\text{Zr/Ti})_N$ ratios, the $(\text{Zr/Hf})_N$ and $(\text{Zr/Ti})_N$ ratios remain close to unity in most samples, and are highest in the carbonatite dyke [S.W58] (~ 10).

7.3d Other trace elements

Sr, P, Y are also important components of the spidergrams used in this study. The Sr-Nd concentrations, shown in Figure 7.6 and discussed in section 7.3a, indicate the partitioning behaviour of various Sr-Nd-bearing minerals in the different samples. Y, like the HREE, behaves as a moderately incompatible element and shows little variation in concentration from the HREE. Figure 7.8f shows a plot of $(\text{Nb/Sr})_N$ vs $(\text{Y/Ho})_N$. There is a gap in the $(\text{Y/Ho})_N$ values between the silicate samples in groups 1 & 4 and group 2, which is filled by the values from the carbonatites. Nash (1972) reported an apparent Y fractionation between the carbonate and silicate rocks at the Iron Hill massif in Colorado, with Y enriched in the silicate rocks but depleted in the carbonatites. With the exception of the melilitolites, this is also what is seen in Figure 7.8f. The low $(\text{Y/Ho})_N$ values from the melilitolites are related to the steep HREE slope in the melilitolites, and reflect the steep REE slope of these samples. Figure 7.8f also shows a wide variation in $(\text{Nb/Sr})_N$ values < 1 from the carbonatites, although relatively constant values > 1 from the silicate rocks, indicating that perovskite fractionation may play some role in controlling the Nb and Sr systematics of the silicate rocks, but little role in the carbonatites. Sr and P are both more enriched in the carbonatites than the silicate rocks, but do not correlate directly with one another in most of the Turiy rocks, indicating that apatite is not the dominant mineral control on Sr partitioning in any of the samples analyzed. Zr is also a common

trace element in apatite (de Mulder et al., 1986), but neither Zr nor Hf concentrations show any correlation with the P content of the Turiy samples indicating that apatite was not important in controlling the Zr-Hf signatures of these rocks. Jones et al. (1995) found that P and F were two of the elements most strongly partitioned into carbonate liquids in carbonate-silicate immiscible systems, and the solution of P also increased the stability field of the carbonate liquid, as well as the Si content of the carbonate liquid. The Turiy carbonatites (and phoscorites) are both rich in (fluor-) apatite and hence P and F, although this is not proof that they have been produced by immiscibility, as suggested by the results of Jones et al. (1995). De Mulder et al. (1986) pointed out that apatite settling in a magma chamber (or separation of a phosphate melt) would explain both negative P anomalies in spidergram patterns of fractionated magmas, and concomittant REE fractionation patterns. Separation of phosphate melts within Turiy is certainly possible, given the apatite veins seen in the phoscorites and some carbonatites. However, as seen in Figure 7.6, there are a number of minerals, in addition to apatite, that may control the REE signatures of the rocks, and hence a direct link between P content and REE content within the samples is difficult to ascertain, as concluded from Figure 7.8c.

Mo, in the same group in the periodic table as W, lies between Ce and Sr in terms of relative incompatibility on a spidergram (Keller & Spettel, 1995). Mo shows a negative anomaly on the spidergram for all samples (not shown) except for WKT which has a flat Ce-Mo-Sr pattern. Fitton (1995) found that there was a directly proportional relationship between Mo and Nb in basalts from the Basin and Range volcanic province. This relationship is also seen in the Turiy carbonatites, but not in any of the silicate rocks analyzed. W, Mo and Sb are strongly enriched in the natrocarbonatite at Oldonyo L'Engai (Keller & Spettel, (1995). The enrichment of W and natrocarbonatite-like LILE patterns in the turjite has already caused speculation that the magmatic components which control the W-Th-U signature of a natrocarbonatite magma might have been involved in

the petrogenesis of the turjite, but this is not matched by the behaviour of Mo and Sb in the latter.

Sb, a member of Group V of the periodic table, lies between Hf and Ti in incompatibility. Relative to Sm (chosen because of the varied Hf and Ti concentrations in these rocks), $(\text{Sb}/\text{Sm})_N$ ratios decrease in the order: (olivine melteigite, fenite = 4) > (dolomite-calcite carbonatite, WKT [melilitolite], melilitite dyke and ijolite = 2.3) > (TUR II [melilitolite], calcite carbonatites = 1.6). The $(\text{Sb}/\text{Sm})_N$ ratio is close to unity in the pyroxenite, C.5.118 [melilitolite] and the carbonatite dyke, but less than 1 in the turjite [TU 119], melteigite [C.AG.122] and C.190.105 [melilitolite]. There is no consistent variation in the $(\text{Sb}/\text{Sm})_N$ ratio within the Turiy samples, based on rock-type.

7.4 Spidergram comparisons

In the previous section the chemical behaviour of the trace elements used in the spidergrams, and their mineral partitioning were examined, in order to evaluate the effects of possible open-system fractional crystallization within these magmas. Spidergrams plotted in Figure 7.7 were primarily grouped according to similarities in petrology in an attempt to understand the trace element variations within a single magma type, or group of closely related magma types (e.g. the olivine melteigite - ijolite - turjite samples). Having constrained the basic characteristics of the trace element signatures of the different rock types, it is now important to establish whether any of the different rock types share a common source or are the products of magma mixing, while bearing in mind the effects that open-system fractional crystallization may have on the sample signatures. Thus, the spidergrams are now plotted in different combinations, and, where appropriate, compared to spidergrams from related rocks from elsewhere on Earth. Similarities and variations between samples which appear to share similar sources, or are related by magma mixing or fractionation are discussed in the following section.

7.4a Pyroxenite-melilitolite-melilitite

The pyroxenite [C.90.29], melilitolites, and a melilitite dyke from the south coast [TU 107] show remarkable parallelism in their trace element spidergrams, with normalized ratios increasing in the order pyroxenite < melilitite < melilitolites, as seen in Figure 7.11a. Because the trace element characteristics for the melilitolites are similar (Figure 7.7b), only one melilitolite from the central massif has been plotted in Figure 7.11a. The greatest variations in abundance are proportional to incompatibility of the elements (i.e. large differences in the LILE, and overlapping patterns in the HREE). The melilitite is closer to the melilitolite than the pyroxenite in terms of incompatible element signatures (i.e. the left end of Figure 7.11a) but all three have similar compatible HFSEs and HREE signatures. Exceptions to this generalization in the pyroxenite signature [C.90.29] are $(\text{Ce/La})_N > 1$, a relative enrichment in W, and lack of enrichment in Sb. The melilitite dyke [TU 107] shares the W-depletion and Sb-enrichment of the melilitolite, but the Zr-Hf enrichment of the pyroxenite. The negative Zr-Hf anomaly remains unique to the melilitolites. However, the gross similarity between the trace element signatures of the three rock types suggests that these samples shared a similar source, and that the majority of their trace element patterns reflect source characteristics and not the result of open-system fractional crystallization, although the LILE-depletion and Zr-Hf enrichment of the pyroxenite suggests that fractional crystallization has probably played a role in its magmatic evolution. Fractionation may also be responsible for the Zr-Hf anomaly in the melilitolites. The isotopic signatures outlined in Chapter 6 indicate that some of the satellite complex melilitolites may have been subjected to crustal contamination. However, mixing calculations indicate the trace element contents of the melilitolites are high enough to make minor amounts of crustal contamination unnoticeable on the spidergram plots. This is shown by sample C.190.105 which has an

Figure 7.11a Mantle-normalized spidergrams of the Turiy pyroxenite [C.90.29], central massif turjaite [C.5.118], melilitite dyke [TU 107] and the type-locality olivine melilitite [U3] from Urach, Germany (Wilson et al., 1995). Normalization factors from McDonough & Sun (1995).

Figure 7.11b Mantle-normalized spidergram of a Turiy pyroxenite [C.90.29], turjaite [C.5.118], turjite [TU 119], along with pyroxenites [HB001 (J) and HB008-1 (J)] from Jacupiranga (Huang et al., 1995), and pyroxenite C-35-63 (C) from the Cargill Complex (Sharpe, 1987). Mantle normalization factors from McDonough & Sun (1995).

uncontaminated isotopic signature, and whose spidergram patterns closely match those of the other melilitolites in Figure 7.7b, two of which have isotopic signatures that suggest they have been affected by contamination.

A useful comparison can be made between the spidergram plots of the Turiy melilitolites and that of the type-locality olivine melilitite (U3) from the Tertiary-Quaternary Urach volcanic field in the Rhinegraben, Germany (Wilson et al., 1995). Figure 7.11a shows that sample U3 has a very similar spidergram pattern to the melilitite dyke from Turiy [TU 107], but has higher Ba, Th, U, P and HREE contents, and lower K and Rb contents. The REE spidergram pattern of the type-locality olivine melilitite U3 is very similar to the REE spidergrams of the central complex carbonatites, along with ijolite sample C.IJ.14 (not shown). Wilson et al. (1995) postulated, on the basis of isotopic and trace element characteristics, that the trace element signature of U3 originated in the asthenospheric garnet-dolomite stability field, and the isotopic characteristics suggested that the source of the U3 magma originated from a mantle component known as HIMU, indicative of the involvement of a mantle plume. Given the steep Ce/Lu slopes in the Turiy pyroxenite-melilitite-melilitolite series, as well as the low absolute abundances of the HREE in these samples, garnet is also likely to have been residual within their magma source. The additionally steep slope in the three Turiy samples, compared to sample U3, begs the question as to whether additional HREE fractionation has also occurred, perhaps by fractionation of pyroxene or even garnet during crystallization. This HREE-depletion effect is most noticeable in the melilitolites (Figure 7.7b), along with the depletion in Zr and Hf, which may also be attributed to garnet fractionation.

The issue of the relationship between the Turiy pyroxenite-melilitolite-melilitite samples, with respect to their mantle source, warrants discussion. Weaver (1991) listed 10 trace element ratios (Zr/Nb, La/Nb, Ba/Nb, Ba/Th, Rb/Nb, K/Nb, Th/Nb, Th/La and

Ba/La), of well-established mantle reservoirs, defined by Zindler & Hart (1986), including N-MORB, HIMU, EM I, EM II, and Primitive Mantle, along with continental crust. Comparison of the values of these trace element ratios from the melilitite dyke [TU 107] (Table 7.B) with the values from Weaver (1991) shows that the trace element signature of TU 107 is most closely matched with that of the HIMU component. Exceptions to this are ratios involving La due to the enrichment of this element above estimated concentrations in HIMU-derived magmas, and the Zr/Nb ratio which is closest to HIMU but slightly lower than typical values for this reservoir. Given the extremely varied contents of Zr (and Hf) in the Turiy rocks, this latter variation is hardly surprising. Thus, this correlation between the pyroxene - melilitolite - melilitite trace element signatures with the HIMU signature, coupled with related similarities already seen in the isotope data (Chapter 6), suggest that a mantle plume has indeed been involved in the genesis of at least some of the Turiy magmas.

Figure 7.11b shows the close parallelism between the trace element signatures of the Turiy pyroxenite and the pyroxenites from Jacupiranga (Huang et al., 1995). A spidergram pattern from a hornblende pyroxenite sample from the Precambrian Cargill carbonatite complex in Canada (Sharpe, 1987) lies parallel to the Turiy and Jacupiranga data, but with a greater degree of trace element enrichment. Huang et al. (op. cit.) calculated the trace element composition of a magma in equilibrium with the Jacupiranga pyroxenites using pyroxene partition coefficients. The resulting pattern did not match any of the samples that were analyzed from Jacupiranga. Similar calculations for the Turiy pyroxenite [C.90.29] were unsuccessful due to the presence of small, but significant amounts of amphibole and perovskite in the sample. An attempt to adjust the calculation to account for the mineralogical variations in sample C.90.29 was also unsuccessful due to difficulties in constraining the trace element characteristics of perovskite.

7.4b Olivine melteigite-ijolite-turjite-carbonatite

Similarity, in terms of spidergram patterns, also occurs between the olivine melteigite [C.AG.48A], ijolite [C.IJ.14] and turjite [TU 119] (Figure 7.12a). The overall normalized spidergram patterns are relatively flat and exhibit no significant enrichment or depletion in any one group of trace elements. The isotope data from the most mafic member of this group of samples, the olivine melteigite [C.AG.48A], lies closest to the EM I mantle component, as discussed in Chapter 6. Comparison of the values of a group of trace element ratios from a number of mantle components (HIMU, EM I, N-MORB, etc) (Weaver (1991), with the values of the same ratios from the olivine melteigite [C.AG.48A] (Table 7.B), shows that the olivine melteigite fits most closely with the trace element characteristics of the EM I component, matching the isotopic correlation between the olivine melteigite and the enriched end of the Kola Carbonatite Line, already seen in Chapter 6. The exception to this, as with the match between the melilitite dyke [TU 107] and HIMU described in the previous section, are ratios involving La, as this element is also enriched in the olivine melteigite above the values expected from EM I source magmas. Thus, the fit of the Turiy silicate rock data, in isotopic space, along the Kola Carbonatite Line, added to the fit of the trace element data from the Turiy samples to the two end-members proposed for the formation of the similar, but modern, East Africa Carbonatite Line, suggests that the HIMU and EM I components also control the slope of the KCL. Kramm (1993) proposed that the enriched end of the KCL was controlled by the EM I component, a finding which is confirmed by the olivine melteigite isotope and trace element signatures. The data from the “depleted” group of Turiy rocks have now enabled constraints to be placed on the mantle component which controls the depleted end of the KCL, which is now believed to be HIMU, which thus provides strong evidence of the existence of a mantle plume underlying the Kola Peninsula, contributing to the

Figure 7.12a Mantle-normalized spidergram of the olivine melteigite [C.AG.48A], ijolite [C.IJ.14], turjite [TU 119], carbonatite dyke [S.W58], and a natrocarbonatite from Oldonyo L'Engai [OL 102] taken from Keller & Spettel (1995). Normalization factors from McDonough & Sun (1995).

Figure 7.12b Mantle-normalized spidergram of the turjite [TU 119] and the fenite [NC.153.82]. Normalization factors from McDonough & Sun (1995).

magmas of the Kola Alkaline Province.

A correlation between the olivine melteigite, ijolite and turjite has already been proposed on the grounds of their petrology and mineral chemistry (Chapters 4 & 5), and major element chemistry (this chapter), and the suggestion has been made that the turjite composition is representative of the reaction between the fractionation products of the olivine melteigite [C.AG.48A] or the ijolite [C.IJ.14] and a carbonatite (Fig. 7.9). The turjite [TU 119] is enriched in W, U, Nb, Sr, Zr, Hf, Y and HREE, and depleted in Rb, K and Ta relative to the ijolite [C.IJ.14]. A summary of the trace element characteristics of this fractionation - reaction relationship in Figure 7.12a also includes two carbonatites, sample S.W58 [carbonatite dyke] and sample OL 102, the natrocarbonatite from Oldonyo L'Engai (Keller & Spettel, 1995). The spidergram patterns of both the ijolite and turjite suggest that they could represent different stages in the reaction process between a residual fractionated melt from a sample such as the olivine melteigite [C.AG.48A], and a carbonatite with a signature somewhere between that of the carbonatite dyke [S.W58] and a carbonatite similar to that from Oldonyo L'Engai (Keller & Spettel, 1995). The turjite sample TU 119 may contain cumulus garnet due to open-system crystal fractionation. This is reflected in the enriched Zr-Hf and high HREE signature of the turjite. The Sb content of the turjite is slightly lower than would be expected from the carbonatite - melteigite/ijolite reaction. The possible influence of components involved in the production of the Oldonyo L'Engai natrocarbonatite in the genesis of the turjite may be reflected in the most incompatible elements, especially the W-Th-U triplet on the spidergram. The rest of the turjite signature can, as a first approximation, be attributed to mixing between S.W58 [calcite carbonatite dyke] with C.IJ.14 [ijolite], or the fractionated olivine melteigite [C.AG.48A].

7.4c Turjite-fenite

Mention should also be made of the similarity between the spidergram patterns of the turjite and fenite (Fig. 7.12b). The spidergrams are essentially parallel to each other, with the turjite showing more enriched trace element signatures than the fenite except for a strong enrichment in K, Rb and Ba in the fenite. The Zr-Hf enrichment, $Nb > Ta$, $U > Th$, and HREE enrichment patterns seen in the turjite are also seen in the fenite but to a slightly lesser extent.

Other than the enrichment in alkalis and alkali earth elements in the fenite (e.g. Na, K, Rb, Ba), there is a noticeable parallelism between the spidergram patterns of these two samples. One possible explanation for this similarity is that both samples share a common source. Wilson et al. (1995) pointed out that parallel trends on spidergrams between different magmas represented different degrees of melting of a common source. A geological map of Kola published in Kogarko (1987) shows Proterozoic “granitoid” rocks in the vicinity of Turiy, which include charnockites directly surrounding the Turiy massif (Bulakh & Ivanikov, 1984). The modal mineralogy of the fenite [NC.153.82] is dominated by alkali feldspar, aegirine-augite and amphibole but no discernable nepheline or quartz, which is typical of fenite assemblages (Bulakh & Ivanikov, 1984).

The original trace element composition of the charnockites around Turiy have probably been affected by fluids emanating from the massif. The present day $^{87}Sr/^{86}Sr$ of the fenite of 0.7058 (Chapter 6) is not representative of typical charnockites (de Mulder, 1986). Mixing calculations carried out in Chapter 7 indicate that the original isotopic composition of the fenite [NC.153.82] has been significantly altered by the magmatic metasomatic fluids emanating from the massif. Such an isotopic model, however, does not explain the close similarity in trace element patterns (particularly HREE and Zr-Hf) between the turjite [TU 119] and fenite [NC.153.82]. An alternative model for the origin of the turjite, based on the similarities in trace element data between the turjite and fenite,

would be to consider the turjite as a rheomorphic fenite. However, the very different isotopic signatures, major element chemistry, petrography and mineral chemistry characteristics of both samples, as described in earlier chapters, almost certainly invalidate this idea.

It is most unlikely that the turjite is responsible for producing the fenitizing fluids which affected fenite sample NC.153.82, as the turjite only occurs in the form of dykes on the south coast, and the fenite sample was obtained from a drill core, north of the central complex. However, Haas et al. (1995) showed that F preferentially complexes with the HREE in hydrothermal systems, while Keppler (1993) showed that F dissolves HFSE minerals in SiO₂-rich systems. Since the fenite contains fluorite, and the turjite samples contain F-bearing minerals such as apatite, it is possible that F-dominated element partitioning behaviour is being demonstrated in Figure 7.12b. Until more information is obtained, further interpretation of the trace element similarities between the turjite and fenite samples would be somewhat premature.

7.4d Melilitolite-melteigite-ijolite

An unresolved problem, in terms of field relationships, petrology, mineralogy, isotope geochemistry, and trace element geochemistry, is the link between the ijolites and other rock types at Turiy. While there seems to be a clear link between the olivine melteigite [C.AG.48A], the ijolite [C.IJ.14] and the turjite [TU 119], the origin of sample C.AG.122 [melteigite] and its link to the other silicate rocks within the central massif has not yet been discussed. Figure 7.13a includes spidergrams for a central massif turjaite [C.190.105], melteigite [C.AG.122] and the ijolite [C.IJ.14]. The most incompatible elements (left hand end of the spidergram) are very similar, in terms of both pattern and concentration, for all three samples. This is considered significant, given that these elements are so incompatible and mobile. Variations in the patterns, most of which can

Figure 7.13a Mantle-normalized spidergram of a central massif Turiy turjaite [C.190.105] and the central massif melteigite [C.AG.122] and ijolite [C.IJ.14]. Normalization factors from McDonough & Sun (1995).

Figure 7.13b Mantle-normalized spidergram of the type-locality turjite [TU 119], turjaite [WKT], bergalite [K 217] and okaite [OK 7]. Bergalite and okaite data from Hornig (1988). Normalization factors from McDonough & Sun (1995).

be attributed to open-system fractional crystallization, start with a slightly elevated Ta content in the melteigite [C.AG.122]. This sample is also slightly depleted in La (due to fractionation of clinopyroxene) and Sr, with respect to the other two samples. The P contents are very variable (melteigite < melilitolite < ijolite) presumably due to apatite fractionation. The melteigite [C.AG.122] also mimics the Zr-Hf depletion of the melilitolite [C.190.105], but to a lesser degree. The ijolite [C.IJ.14] is relatively enriched in Zr-Hf-Sb, although it mimics the Zr-Hf-Sb pattern of the melilitolite [C.190.105] at higher concentration, while the melilitolite shows strong Zr-Hf and HREE depletion with respect to the melteigite and ijolite, presumably due to garnet fractionation at some point in its magmatic history.

The close link between these three samples in terms of their trace element chemistry, although not their petrology or mineral chemistry, are also interpreted in terms of their major element chemistries given in Table 7.A. The major differences between the melilitolite and the melteigite-ijolite pair are in SiO₂, Al₂O₃ and CaO contents which are approximately 2-3% lower, and 1% and 3-4% higher respectively in the melilitolite. There are a few minor differences between the melteigite [C.AG.122] and ijolite [C.IJ.14] - the ijolite contains approximately 1.5% more SiO₂ and 1.5% and 1% less Fe₂O₃ and CaO respectively, than the melteigite. As expected from Figure 7.1, a decrease in SiO₂ and increase in CaO content tends to favour the production of melilite, and this is borne out by the major element chemistry of these three samples.

The unequivocal link between the ijolite and turjite, discussed earlier, has an added implication if there is also a petrogenetic link between the turjite [TU 119] and melilitolite [C.190.105]. This model receives support from the presence of relict melilite in turjite sample TQ, some strong similarities in mineral compositions between some of the turjite samples and melilitolites, as seen in Chapter 5, and antipathetic Zr-Hf and HREE signatures in the two rock types. One possibility would be that the turjite

represents a residual liquid from the fractionation of the melilitolites, or that it represents the product of a melilitolite - carbonatite reaction, as opposed to the ijolite - carbonatite reaction suggested earlier.

Comparison is made in Figure 7.13b between the trace element compositions of the type-locality okaite from Oka and bergalite from Kaiserstuhl (Hornig, 1988), with the type-locality turjaite [WKT] and turjite [TU 119] from Turiy. Bergalite is considered by Le Maitre (1989) to be the hypabyssal equivalent, and okaite the hauyne-bearing equivalent of turjaite. The spidergram patterns of the okaite and turjaite are very similar to each other, including the significant depletion in Zr, although the okaite does not share the extreme HREE depletion seen in the turjaite. The Kaiserstuhl bergalite shows some similarities in trace element pattern with both the turjaite (in the middle part of the spidergram), and the turjite (in the most incompatible and compatible elements plotted on the spidergram). However, further interpretation of the similarities between the spidergrams of these four type-locality rocks would be inappropriate without inclusion of trace element data from other rock types associated with the Kaiserstuhl bergalite and Oka okaite, but it can be seen that there are some interesting similarities between these samples.

7.4e Carbonatites

The four carbonatites selected for analysis exhibit very different petrological and major-element chemistry characteristics as outlined earlier. Similarly, their trace element signatures are also varied (Figure 7.14). Sample S.W58, one of fourteen carbonatite dykes on the south coast of the Turiy peninsula (Bulakh & Ivanikov, 1984; Bell et al., 1996), exhibits the most enriched signature of all samples analyzed. Sample S.W58 has an almost identical trace element signature to the “average calcite carbonatite” of Woolley & Kempe (1989) (Figure 7.15a). The central massif dolomite-calcite

carbonatite [C.DC] is close to the “average Mg-carbonatite” of Woolley & Kempe (1989) (Figure 7.15b) and also shows a similar pattern to S.W58, but at lower degrees of enrichment. The lack of Nb in sample C.DC is the one notable exception to the match (Figure 7.14). The other carbonatites analyzed are three central massif samples [C.49.40, C.TL.344 and C.DC] which are much less enriched in trace elements than S.W58 (Figure 7.14). Their REE patterns and contents are similar to that of the pyroxenite [C.90.29], i.e. lower than most of the silicate rocks. Of particular note are the relatively low abundances of Nb-Ta and Zr-Hf, and $(\text{Ta/Nb})_{\text{N}} > 1$ in all three central massif carbonatites which is unusual in most carbonatites (Figure 7.15). All three samples have strong positive peaks at Sr and P (presumably due to calcite and apatite accumulation). There are also strong inter-sample variations in K in the central massif carbonatites, demonstrated by sample C.49.40 which is rich in phlogopite and has relatively high K and Rb contents, while C.DC, which has no mica, has low K and Rb contents. Similar large variations are seen in the most compatible of the HFSEs with relatively high Zr-Hf-Ti contents in sample C.TL.344, minimal contents in sample C.DC [dolomite calcite carbonatite], and intermediate values in sample C.49.40. Sample S.W58 is significantly enriched in Th, U, Pb and REE compared to all other samples. The reasons for the unusual trace-element signatures of samples C.49.40 and C.TL.344 may be due to the fact that they are poor in calcite (40-45%), and rich in apatite and magnetite, i.e. grading into phoscorite, in terms of their modal mineralogy. However, the isotopic signature of sample C.49.40 falls within the carbonatite group in Figure 6.4 and should not be classified as a true phoscorite, as discussed in Chapter 4. Pyrochlore has only been found in one carbonatite (C.100.y, Appendix. Ji). The trace element signatures of the three central massif carbonatites are depleted relative to those of sample TU 107 [melilitite dyke] in all elements except Sr and P (Figure 7.17).

Figure 7.14 Mantle-normalized spidergrams of the Turiy carbonatites [C.49.40, C.TL.344, C.DC, S.W58] and a natrocarbonatite from Oldonyo L'Engai [OL 102] (Keller & Spettel, 1995). Normalization factors from McDonough & Sun (1995).

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Figure 7.15a Mantle-normalized spidergrams of the carbonatite dyke [S.W58] and the “average calcite carbonatite” of Woolley & Kempe (1989). Normalization factors from McDonough & Sun (1995).

Figure 7.15b Mantle-normalized spidergrams of the central massif dolomite calcite carbonatite [C.DC], the “average Mg-carbonatite” of Woolley & Kempe (1989), and a Mg-carbonatite from Iron Hill (Keller & Spettel, 1995). Normalization factors from McDonough & Sun (1995).

Simonetti et al. (1997) compared values of various trace element ratios from the 1993 Oldonyo L'Engai carbonatite flows with those from MORBs and OIBs and related rocks, in an attempt to see if trace element signatures are similar to those of known mantle components. It is useful to compare the Turiy carbonatite data with the trace element signatures of the various mantle components, and with the new Oldonyo L'Engai data from Simonetti et al. (1997), as Oldonyo L'Engai is the only active carbonatite on earth. Non-normalised ratios for the Turiy samples shown in Table 7.B, and comparison of these ratios with the data from Simonetti et al. (1997), reveals some significant differences between the Turiy carbonatites and the Oldonyo L'Engai natrocarbonatites, as seen in the spidergram comparisons in Figure 7.14. Simonetti et al. (1997) report $Ba/Sr > 0.7$ for the natrocarbonatites and < 0.7 for most calciocarbonatites. The Turiy carbonatites all have values between 0.06-0.08. The Nb/U ratios of the 1993 natrocarbonatite flows are 7.8 - 16.5, while Hofmann et al. (1986) quoted values of 47 ± 10 for MORBs and OIBs. The Turiy Nb/U ratios for the two central complex calcite carbonatites [C.49.40 and C.TL 344] are 50.4 - 55.1, consistent with the mantle-derived basalt values proposed by Hofmann et al. (1986). However, the Nb/U ratio of the dolomite calcite carbonatite [C.DC] is 7.9, while the value for the carbonatite dyke [S.W58] is 26.2. The Turiy carbonatites have low Nb values, ranging from 3-11ppm for the central complex samples, but the carbonatite dyke [S.W58] has a much higher value of 391ppm. Simonetti et al. (1997) also report low Nb concentrations in the natrocarbonatite samples (18-104ppm), while Woolley & Kempe (1989) suggest that typical carbonatites contain 500-1200ppm Nb. The Th/U values of the natrocarbonatites are all < 1.0 (Simonetti et al., 1997), and range from 2.5-4.0 in chondrites, oceanic basalts and primitive mantle (Sun & McDonough, 1989). Values for the Turiy carbonatite samples fall between 5.3 and 30.8. The Th/Nb ratios of most Turiy whole rock silicate

Figure 7.16 Mantle-normalized spidergrams of a central massif carbonatite [C.49.40], carbonatite dyke [S.W58], central massif melilitolite [C.190.105], pyroxenite [C.90.29] and turjite [TU 119]. Normalization factors from McDonough & Sun (1995).

Figure 7.17 Mantle-normalized spidergrams of the central massif carbonatites [C.49.40, C.TL.344, C.DC] and the melilitite dyke [TU 107]. Normalization factors from McDonough & Sun (1995).

samples are 0.02 - 0.13, similar to the Th/Nb ratios considered characteristic of HIMU oceanic island basalts (Weaver, 1991). However, the Turiy carbonatites have Th/Nb ratios significantly higher which fall between 0.18 and 0.67, although the Oldonyo L'Engai natrocarbonatites have Th/Nb ratios between values similar to the HIMU value quoted above, and that of primitive mantle (0.117, Weaver, 1991). The Sm/Nd ratios of the Turiy carbonatites (0.12 - 0.17) are similar to the value of 0.15 considered typical of carbonatites by Bell & Blenkinsop (1987a,b), unlike the Sm/Nd ratios of the natrocarbonatites that range from 0.070 to 0.081 (Simonetti et al., 1997).

This comparison between the trace element signatures of some important silicate rocks and mantle reservoirs, the Turiy carbonatites and the Oldonyo L'Engai natrocarbonatites show that the values from the Turiy carbonatites are quite different to those of natrocarbonatites, and to known mantle reservoirs.

7.5 Summary

The open-system conditions suggested by the isotope data in Chapter 6 are corroborated by the trace element data. Fractional crystallization has played a significant role in the evolution of these samples, although there may have been a partial transition from open system to closed system conditions during the transition from magma emplacement to crystallization, as suggested by the consistency of the isotope data within each rock type (Chapter 6), and the consistency of the trace element data within individual rock types (e.g. the melilitolites). Despite the open-system effects, the trace element data confirm the contribution of the HIMU and EM I mantle components to the sources of pyroxenite - melilitolite - melilitite group and olivine melteigite - ijolite - turjite group of rocks, respectively, and provide strong evidence that a mantle plume was involved in the derivation of the Turiy magmas. The theory that the turjite is the product of a reaction between an ijolite and a carbonatite is supported by the trace element data.

The sources of the carbonatites are still unclear, but the carbonatite dyke from the south coast bears no resemblance, in terms of trace element signature, to the carbonatites from the central complex. The carbonatites from the central complex are less enriched in incompatible trace elements than most calcite-rich carbonatites.

CHAPTER 8

DISCUSSION

8.1 Introduction

In previous chapters, various aspects of the geological history and geochemistry of the Turiy Massif have been described and comparisons have been drawn with other massifs. The primary aim of this chapter is to devise a model of evolution for the Turiy massif. Results from studies of other carbonatites and SiO_2 -undersaturated rocks, along with experimental work, are examined in order to place the results obtained from the Turiy rocks into a wider context. Tectonic settings of carbonatites are evaluated, and this is followed by a discussion of the characteristics of the Turiy magma sources, ideas regarding melt production and magma emplacement, and a discussion of the petrogenesis of particular groups of rocks from within the massif.

8.2 Tectonic setting

The Turiy massif lies within the Kandalaksha Graben, a continental rift developed within the Baltic Shield craton. The carbonatites at Turiy are part of an alkaline ultrabasic massif, emplaced at the same time as a suite of diamond-bearing kimberlite pipes in the nearby Terskii Coast, and lying just outside the major walls of the Kandalaksha graben structure (Kalinkin et al., 1993). Thus, according to Barker (1989), the carbonatites at Turiy appear to overlap the world-wide groupings of “cratonic kimberlite-carbonatite” and “continental rift-alkaline massif-carbonatite”, as well as the “hypabyssal” group represented by the carbonatite dykes on the south coast of the Turiy Peninsula. Most Paleozoic carbonatites in the Kola Peninsula are plutonic and Turiy is one of the few such carbonatites to be associated with melilitolites. Another example is the Kovdor Massif, also located along the Kandalaksha Graben. Turiy also bears many similarities, both in terms of structure and composition, to the Kaiserstuhl carbonatite

complex located just within the Rhine graben in Germany. Like Turiy, the Kaiserstuhl carbonatite itself, lying just within the flanks of a graben, is dominated by sovitic carbonatites, and contains melilite-bearing rocks, including type-locality bergalite, believed to be the hypabyssal equivalent of turjaite (Keller, 1997). Turiy, however, is an intrusive complex, while Kaiserstuhl is extrusive-hypabyssal (Keller et al., 1990).

What initiated magmatism within the Kandalaksha Graben during the Late Proterozoic and Devonian times is difficult to establish. Melting within the mantle could be initiated by active or passive rifting (Sengor & Burke, 1978) or by influx of volatiles causing depression of the mantle solidus. No available geophysical data allows calculation of the β -factor (crustal stretching parameter) to enable any kind of estimation of crustal stretching, and the likelihood of crustal doming or sagging during the Devonian. Woolley (1989) stated that the majority of carbonatites world-wide occur in areas of continental crustal doming and thus it is likely that they are influenced by active rifting and mantle upwelling (McKenzie & Bickle, 1988).

The sheer quantity of alkaline magmatism which is believed to have been emplaced within the Kola Peninsula during the Devonian (Kramm et al., 1993) suggests that a plume may be responsible for initiating much of the magmatic activity. Wilson et al. (1995) explained the isotopic signatures of the Tertiary - Quaternary magmatism occurring across Europe by means of a mantle plume, and suggested that a plume might be broken up into a series of small "plumelets" based on the geophysical results of Hoernle et al. (1995) and Glahn et al. (1995) which showed a sub-horizontal "sheet" of low-velocity (volatile-enriched) material within the asthenosphere beneath Europe from which small pulses of enriched material might ascend to the base of the lithosphere. The nature of the scattered Tertiary-Quaternary magmatism beneath Europe is similar to the twenty or so Devonian intrusions scattered across the Kola Peninsula. These tectonic features, coupled with the isotopic signatures of many of the massifs, suggests that a mantle plume

may well be involved in the Kola magmatism.

Whether a plume or “plumelets” rise from the asthenosphere beneath the Kola lithosphere, or whether the magmatic activity is initiated by passive rifting, the emplacement of the magmas subsequently produced will be controlled by the tectonic properties of the overlying brittle Mechanical Boundary Layer (MBL) (Wilson et al., 1995), and magmas will migrate towards and be preferentially emplaced along zones of pre-existing weakness. Dunbar & Sawyer (1989) showed that major rift zones (and associated magmatic activity) tend to follow pre-existing lithospheric sutures and zones of weakness. Milanovsky (1992), in a study of Russian aulacogens, described activation of many such rift structures during the Late Proterozoic and reactivation during the Devonian. The reason for this reactivation is unclear, but would seem to fit with the bimodal Precambrian - Devonian magmatism seen in Turiy (Table 6.D), the Precambrian phase identified by pyroxenite sample C.AG.347. Whether such magmatic bimodality also exists in the Kovdor massif, which was also emplaced within the Kandalaksha Graben, is unclear. The ancient suture zone represented by the Kontozero Graben, trending south west - north east in the Kola Peninsula, was also reactivated during the Paleozoic, as seen by the emplacement of the Khibina and Lovozero intrusions and other massifs (Arzamastsev, 1994). Kravchenko et al. (1997) showed a Moho relief map for the Siberian platform, similar to those published for Europe (e.g. Glahn et al., 1992). These maps indicate topographic highs in the Moho associated with rift zones, kimberlite intrusions and ring structures, and were thought by Kravchenko et al. (1997) to be related to mantle convection cell boundaries.

Thus, although the geophysical data currently available seems to suggest reactivation of ancient rifts as a reason for the geographical distribution of the massifs within the Kola Peninsula, it cannot fully constrain the causes of that reactivation, and subsequent magmatic activity. Further constraints may be provided by the use of

geochemical data to discern the presence or absence of a mantle plume in the history of the Turiy rocks.

8.3 Magma sources and melt production

The relationship between carbonatites and lithospheric geodynamics is still a matter for debate. Carbonatites do not occur in the oceanic crust except where that crust is part of an abnormally thick lithosphere (e.g. Canary Islands, Cape Verde Islands). Evidence of carbonatite metasomatism has been recorded from xenoliths erupted within ocean island basalts (e.g. Hauri et al., 1993), while primary calcite has been found by Pyle & Haggerty (1994) and Kogarko et al. (1995) in Group II mantle xenoliths (Frey & Prinz, 1978). These findings suggest that carbonate melts exist in the mantle beneath both the continents and the oceans, but that the carbonate is only concentrated into carbonatite magmas within the brittle continental lithosphere. In this environment, preservation of very small quantities of partial melt are possible because the geotherm tangentially approaches the mantle solidus and the mechanical boundary layer (MBL) is brittle enough to allow preservation and eruption of carbonatite melts derived from, or held within, the lithospheric thermal boundary layer (TBL) and asthenosphere beneath (Dalton, 1993). These formation and preservation conditions are also shared by other magmas rich in volatiles, possibly including kimberlites (Dalton, 1993). Such melts upwell diapirically through the mantle beneath the oceanic crust, which is a much slower process than dyke emplacement through the brittle continental cratonic lithosphere. Hence, such melts are unable to maintain their disequilibrium route to the surface in the oceanic lithosphere and are equilibrated with their wall-rocks, either within the TBL, or within the MBL (Figure 8.1) (Dalton, 1993). The implications of Figure 8.1 for the Turiy magmas include the fact that the preservation and eruption of carbonatites and kimberlites is to be expected in areas of cratonic rifting, and that the physical or chemical death of

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these and related magmas within the lithosphere (Bell, 1997) creates a source of volatile-enriched material that may be tapped over time by later phases of rifting and melting, or entrained within magmas ascending from below, during subsequent phases of magmatic activity. Given the Precambrian-Devonian bimodal ages seen within the pyroxenites at Turiy, “retapping” or lithospheric contamination may well have occurred in this area, emphasized by the variety of isotopic signatures seen among the different rock types.

The critical questions that need to be answered with respect to magma production at Turiy, are:

- a) what are the mantle sources of the Turiy magmas? In Chapter 7 it has already been shown that several sources are involved in the production of the Turiy magmas.
- b) where are the sources of the Turiy magmas? Melting of different compositions of different components at different pressures and depths within the mantle will produce magmas of different compositions. Are we dealing with a stratigraphic or geographical variation in source locations, or are the variations solely due to differences in the characteristics of metasomatic fluids?
- c) what initiated melting? Was melting caused by an increase in temperature of the source rocks (mantle plume), or a decrease in pressure (lithospheric stretching), or due to volatile-rich metasomatism of the source(s) causing a depression of the mantle solidus?
- d) did all the melts reach the surface, and are the melts which did get through representative of the melts which attempted to reach the surface?

In order to understand melt generation within the mantle, it is important to know the composition of the source material. Major phases present in the upper mantle are olivine, clinopyroxene, orthopyroxene, garnet and spinel. Carbon is present in the mantle in form of reduced carbon (CH_4), elemental carbon (diamond and graphite), and oxidized carbon in the form of CO_2 and carbonate (magnesite, dolomite and calcite). Minor phases stable in the mantle environment are rare, but important hosts for a variety of trace

elements. HFSE phases such as rutile and titanite are believed to be stable in the mantle in subduction-zone environments. Trace-element-rich phosphate veins, possibly representing frozen melts, have been found between the grain boundaries in mantle xenoliths (Rosenbaum et al., 1997). H_2O , F and K are contained within phlogopite and amphibole. The stability of each one of these phases is dependent on the pressure-temperature-chemical composition (P,T,X) parameters that govern each part of the mantle.

If the temperature of the mantle geotherm is elevated, or the pressure decreased, in volatile-absent conditions, the first melt from a mantle peridotite will be produced within the TBL, at what is effectively the base of the tectonic plate (Latin et al., 1990). However, in the presence of large quantities of volatiles (H_2O , CO_2) the shape of the mantle solidus changes dramatically (Figure 8.2), as first documented by Wyllie & Tuttle (1960), but the point of the first intersection of the volatile-rich or anhydrous mantle solidus and the geotherm will still be within the TBL. One of the important implications of this finding is that the distinction between lithospheric and asthenospheric melts becomes somewhat ambiguous at this point, as the volatile-rich trace melts building up within the lithospheric TBL are presumably derived over time from the underlying asthenosphere (Wilson et al., 1995), and the evolution of any trapped melt within the lithosphere during this time will also affect its geochemical signature. Thus, suggesting definite, physical locations for the mantle sources which contributed to the Turiy magmas may not be particularly appropriate.

Higher up within the lithosphere the shape of the (H_2O -) CO_2 -saturated peridotite solidus creates a sub-horizontal ledge on which melts derived from within the TBL may be trapped (Figure 8.3), releasing fluids and vapours which may metasomatize the surrounding wall-rocks within the low-porosity MBL (Wyllie, 1987). The majority of trace melts rising from the base of the TBL likely die within the lithosphere either due to

Figure 8.2 Sketch of the main features of the phase equilibria of the upper mantle within a cratonic environment (after Wyllie, 1987 & 1989 and references within, and Wilson et al., 1995). The lithosphere-asthenosphere boundary lies within the TBL essentially between points A and B, and is defined as the intersection of the tangents to the advective and conductive geotherms (Latin et al., 1990). Peridotite = dry peridotite solidus. For a peridotite saturated in volatiles, melt will be generated between points A - B, where the solidus first crosses the geotherm at the base of the TBL. Position of oxygen buffers from Haggerty (1987). Abbreviations: Hb = hornblende; Mag = magnesite; Do = dolomite; V = vapour; WM = wüstite-magnetite; IW = iron-wüstite; NNO = nickel-nickel oxide; FMQ = fayalite-magnetite-quartz.

thermal death (Spera, 1984) or chemical death (Bell, 1997). However, those melts created where the geotherm crosses the volatile-saturated solidus in Figure 8.3 ascend adiabatically to the surface. Depending on the slope of the geotherm (and temperature of the melts), the melts may either reach the ledge (thin lithosphere) or ascend directly to the surface (thicker lithosphere). The extent of the ledge will also depend on the $\text{H}_2\text{O}:\text{CO}_2$ ratio (Wyllie, 1987). The role of F in changing the slope of the peridotite solidus has not yet been investigated in systems relating to carbonatite formation, although low-pressure experiments in F-bearing carbonatite systems suggest that F can considerably lower the liquidus temperature of carbonatite systems (Jago & Gittins, 1991).

If adiabatic diapiric upwelling were to cause melts to cross the anhydrous peridotite solidus, basaltic magmas would be produced (Wyllie, 1987). However, melts which do become trapped under the ledge (near Q in Figure 8.3) almost certainly include carbonatites, nephelinites and other related alkalic magmas, possibly including those which subsequently produced the Turiy rocks. These melts crystallize and release vapours which metasomatise the overlying mantle at approximately 60-100km depth in a cratonic environment (Haggerty, 1987). If the melts crystallize beneath point Q on Figure 8.3, then the vapours will be H_2O -rich since free CO_2 is not stable in the carbonate stability field, but if the melts crystallize above point Q, the $\text{CO}_2:\text{H}_2\text{O}$ ratio will match that of the system (Wyllie, 1987). The release of volatiles from the "ledge" will fracture the overlying MBL, and allow the escape of magmas from the ledge environment to the surface. Complications to this basic outline of the processes involved in melt production include variations in the depth of the geotherm-solidus intersection, variations in the $\text{CO}_2:\text{H}_2\text{O}$ ratio causing variations in the shape of the solidus, variations in the adiabatic path of any diapiric material rising through the lithosphere, the unknown effects of ancient volatile-rich pockets trapped in the MBL, etc.

Locating the mantle sources of the silicate magmas found at Turiy and along the

Figure 8.3 Sketch based on Figure 8.2, showing the fate of melts derived from various parts of the mantle for a CO_2 -saturated peridotite in a cratonic environment. Melts derived directly from the asthenosphere may ascend directly to the surface. Others, generated between “a” and “b” may bypass the ledge at Q and head to the surface, or head directly to the peridotite solidus (anhydrous) and produce basaltic melts. Some melts will reach the horizontal ledge at Q, freeze, and metasomatise the surrounding lithosphere, as shown by the “clouds”. Release of vapours in this environment will cause crack propagation within the brittle MBL, allowing the ascent of the low degree partial melts previously stored beneath the ledge. This is believed to be the area of derivation of most carbonatites and nephelinites. (Wyllie, 1987 & 1989).

Terskii Coast requires knowledge of the magmas derived from them and the partition coefficients pertaining to the mineral-melt compositions involved. The samples believed to be originally derived from the “asthenospheric source” (the pyroxenite-melilitolite-melilitite group) have steep REE patterns indicative of residual garnet in the source, and are highly enriched in incompatible elements while retaining depleted isotopic signatures. This is thought to be characteristic of a source which has been metasomatized a short time (geologically speaking) before melt formation and extraction (e.g. Rudnick et al., 1993). The negative K anomaly in this group of isotopically depleted samples may also be indicative of residual phlogopite in the source, given that phlogopite is stabilized in the mantle in the presence of excess carbonate (Rogers et al., 1992). The similarity between the trace element characteristics of this “depleted” group of samples and the type-locality olivine melilitite from the Urach volcanic province in Germany, which is also believed to be derived originally from an asthenospheric source underlying a rift zone (Wilson et al., 1995), and is associated with the Kaiserstuhl carbonatite, supports an asthenospheric derivation for the isotopically depleted group of Turiy silicate rocks. Both the Turiy and Rhine graben melilitites share almost identical trace element signatures to the HIMU trace element signature, as characterized by Weaver (1991) on the basis of data from oceanic basalts. The isotopically enriched olivine melteigite group at Turiy shows no evidence of garnet retention (relatively flat HREE patterns), suggesting that the proposed source for the olivine melteigite group lies within the spinel stability zone in the mantle. The trace element ratios of the olivine melteigite sample [C.AG.48A] are almost identical to the EM I trace element signature (Weaver, 1991). In trace element terms, therefore, this appears to create a HIMU - EM I mixing line between the two groups of silicate rocks, in terms of trace element signatures, but, as has been discussed in the previous chapter, the isotopic signature of the depleted group of samples is not representative of the HIMU component. This implies that the HIMU isotopic and trace element signatures

are decoupled from one another, as has been suggested by Rudnick et al. (1993). Since the HIMU isotopic component is often associated with a mantle plume signature (Hofmann, 1997), this also implies that a HIMU trace element signature is not necessarily representative of plume-derived melts.

The trace element signature of the olivine melteigite pipe [C.AG.48A], which is similar to modern-day EM I basalts, is also similar to the signature of the Terskii (Group 1) kimberlites. It is difficult to interpret the significance of this relationship. The Terskii kimberlite signature, lying close to the bulk Earth – CHUR intersection, may provide a possible “enriched” end-member for the KCL (Kramm, 1993) excluding the solitary data point on the KCL from Vuoriyarvi which falls within the enriched quadrant of the Sr-Nd anti-correlation plot (Kramm, 1993), and which is not in agreement with recent, more depleted isotope data from the Vuoriyarvi carbonatites (Demaiffe et al., 1997). Trace element data from the Terskii kimberlites and melilitite pipes (Beard et al., in prep.), do not show extreme degrees of trace element enrichment. Barring enrichment in Ba, the data ranges from 350x to 3x primitive mantle, for elements ranging from Rb to Y, and the highest values are from Nb and La. The level of enrichment is lower than that typically expected for kimberlites, and is similar to the trace element data from the olivine melteigite series rocks from Turiy (also ~350 to 3 times primitive mantle).

8.4 Carbonatite magmas - primary or secondary?

Turiy contains a variety of carbonatites, although their origin is still not clear. The possibility that at least some of them may represent pristine or fractionated “primary carbonatites” cannot be ruled out, but the characteristics of so-called “primary carbonatites”, necessary for their identification, have not yet been fully established. Characteristic features of mantle xenoliths found in areas of carbonatite intrusion include a bias towards wehrlite compositions (olivine-clinopyroxene) indicating that the

orthopyroxene typically present in mantle lherzolite, has been removed by reaction:

It is important to note that this implies that the carbonate present within the carbonate stability field has melted, migrated upwards and subsequently metasomatized the peridotite above to create wehrlite (Dalton & Wood, 1993). Most initial “primary carbonatite melts” never make it to the surface, but metasomatise the wall-rocks that form the plumbing system for later magmas to reach the surface (Spera, 1984) (Figure 8.3). Only when this metasomatism has been completed, and the wall-rocks are saturated in the carbonate component (i.e. reaction (i) has reached equilibrium) is the possibility of “chemical death” eliminated, and “primary carbonate melts” can migrate further up into or within the lithosphere. These criteria are believed to be hard to meet during the early stages of genesis of an alkaline massif such as that seen on the Turiy peninsula.

The next question which must be answered is whether the initial melts created within the carbonate stability field are solely carbonatitic, whether they are a mixture of carbonate and silicate material (Minarik, 1997), or whether the initial carbonatite melts migrate upwards as previously described, into a zone of increased partial melting, and mix with and/or react with a silicate melt. If the latter scenario is the case, then the carbonate melt must again saturate its silicate host melt, prior to forming a solitary carbonatite melt which can migrate to the surface. This subsequent carbonatite melt may be derived from a carbonate-saturated silicate melt by liquid immiscibility or crystal fractionation and will be called a “secondary carbonatite” in this discussion. It is likely that so-called “secondary carbonatites” represent most carbonatites seen at the surface today.

The carbonatites at Turiy are mostly late-stage, i.e. they are the last magmas to be intruded within the central complex, as is common in many carbonatite-silicate complexes (Barker, 1989; Kogarko et al., 1995). If they have been derived from silicate

magmas by immiscibility, fractionation or differentiation, then they will all be “secondary carbonatites”. However, the possibility that they could represent “primary carbonatites” cannot yet be ruled out, on the grounds that the plumbing system will already have been established by the earlier carbonated silicate magmas, - a theory borne out by the location of the carbonatites in the centre of the central complex at Turiy, suggesting that they were the last magmas to come straight up the main vent. Similarly, the wall-rocks will already have reacted with the preceeding magmas, and reaction (i) above should have reached equilibrium. What is more important at this point in time, is the ability to *extract* the very small percentages of melt needed to produce a “primary carbonatite”, directly from its source. A more substantial degree of partial melt would have been needed to establish the original plumbing system, hence we see carbonate-rich *silicate* magmas as the first phases of the intrusion. However, as the magmatic activity dies down and the solidus and geotherm move closer to one another, the degree of partial melt decreases, and ultimately, the final melt to be produced will be similar in composition to the first melt produced (but not extracted) from the source - i.e. a carbonatite (Dalton, 1993). The extraction of such a carbonatite melt to the surface at the end of magmatic activity is now permissible as it will migrate rapidly through the established plumbing system, and will be chemically buffered by the existing wall-rock compositions which have been metasomatized by previous generations of carbonate melt. The minimum melt extracted will be equal to the maximum melt left behind in the source at the end of eruptive activity (Minarik & Watson, 1995). This is not to say that the central complex carbonatites at Turiy are necessarily “primary”, but that this mechanism provides a means for their eruption on the surface, now that experimental studies (Wallace & Green, 1988; Dalton & Presnall, 1997) have proved that it is possible to actually produce “primary carbonatites” within the mantle. Geochemical criteria may thus aid the resolution of the “primary” vs “secondary” carbonatite debate at Turiy. Eggler (1989) stated that the criteria for

recognising primary carbonatites included a $Mg\# > 68$, finite quantities of silicate components, compositions representative of near-liquidus melts, and “appropriate trace element concentrations”. Most of the Turiy central complex carbonatites are low in SiO_2 , three out of the four carbonatites analysed for whole-rock major and trace element contents have $Mg\#s > 62$, but their $Ca\#s$ ($Ca/(Ca+Mg)$) range from 0.84 – 0.91 which appears to be too high for liquidus compositions of primary carbonatites (Dalton & Presnall, 1997).

In the light of earlier discussion regarding the influence of passive versus active rifting in the production of the intrusions lying along the Kandalaksha Graben, the possible influence of a mantle plume on the Turiy magmatism should be considered. It has been common in recent studies to attribute many isotopic and trace element signatures of modern low- SiO_2 melts (Wilson et al., 1995) and carbonatites (e.g. Bell & Simonetti, 1996) within relatively modern rift environments to a “mantle plume”, and, as was discussed earlier, the melilitite-melilitolite-pyroxenite group at Turiy show trace element but not isotopic signatures consistent with derivation from a plume source. If a plume has been involved in the genesis of the Turiy magmas and the development of the Kandalaksha Graben as an active rift zone, then it will not only have affected the chemistry of the magmas, but also the ambient geotherm in the mantle, the solidus-geotherm relationship, and the porosity-permeability characteristics of the underlying asthenosphere. The latter effects are not well constrained at this time, except that it is assumed that the higher temperatures of the plume (100-300°C above the ambient mantle temperature) can only increase the degree of melting and hence increase the likelihood of magmas reaching the surface.

It is unclear whether a magma chamber underlay the Turiy intrusion. The Kola Peninsula cross-section shown in Figure 2.2 indicates that a magma chamber feeding the intrusion may have lain beneath the Proterozoic charnockite-granitoid body that encloses

the intrusion. Cumulates are common beneath such intrusions and other igneous occurrences in rift zones. Certainly, fractional crystallization has occurred in many of the rock-types seen within the massif, but whether this occurred during magma ascent with the crystallization of material which subsequently lined the magma conduits, or whether this occurred due to crystal settling within a magma chamber, is unknown. The variation in isotopic composition between the different rock types also implies the magmas at Turiy are not solely derived by differentiation of one large batch of magma. This is consistent with the findings from volcanic centres such as Oldonyo L'Engai and Shombole (Bell & Petersen, 1989; Bell & Dawson, 1995).

One problem with involving a magma chamber at Turiy is the emplacement of the pyroxenites, which represent the first phase of magmatism. A mechanism to explain why a magma chamber would erupt a cool, dense accumulation of pyroxene crystals from (presumably) the base of the chamber, as the first phase of magma intrusion, in preference to the hot, convecting magma above, is lacking in the geological literature. Alternative explanations include the idea that the pyroxenites represent the base of an older intrusion which lay above the current level of the Turiy Massif and has since been eroded, or that the pyroxenites in fact represent the > 2 kbar equivalent of nephelinite or melilite-nephelinite magmas, as experimentally determined by Tilley & Yoder (1968) which has already been proposed in an earlier chapter and which would be consistent with the suggested depth of the intrusion (Bulakh & Ivanikov, 1984).

If there was a single magma chamber that underlay the current level of the intrusion, the contents of which fractionated to produce the different magma types seen in the intrusion, the range of isotopic values must have been caused by new melts entering the deep magma chamber from the mantle, mixing with the material already present, and then erupting the next magma pulse. This mechanism is similar to the model proposed by Simonetti & Bell (1994) for the Napak magmas. However, it should also be remembered

that bulk mixing of the most mafic silicate magma types at the depleted and enriched ends of the isotopic spectrum (pyroxenites and olivine melteigite) will not produce the magmas with intermediate isotopic values (ijolites, melilitolites), without significant fractionation prior to eruption. Crustal contamination has only affected rocks from the satellite complexes, and it is believed that this happened during the emplacement of the magmas at their current level, and not within a deep magma chamber feeding the whole of the massif. The distinct and consistent groupings of isotopic values according to rock-type, suggests that, once emplaced, the magmas evolved with a reasonable degree of closed-system behaviour.

Immiscibility is not indicated by the mineralogical, geochemical or isotopic data. If a carbonatite had been derived as an immiscible liquid, under closed-system conditions, from a carbonated silicate magma, there should be a consequent decrease in the Ca content of some of the silicate rocks. Rather, Ca-saturation of the silicate rocks is the norm, indicating possible fractionation of CaCO_3 -oversaturated melts to produce carbonate liquids. The lack of overlap between the carbonatites and silicate rocks in terms of their isotopic data indicates that if closed-system immiscibility or fractionation has occurred, then it occurred early in the history of the magmas, and that one or other of the conjugate silicate-carbonate liquids has evolved under open-system geochemical conditions in order to change its isotopic composition.

Another, unresolved issue is the origin of the ijolite bodies which form the outer part of the complexes. The theory that they erupted prior to the intrusion of the melilitolites has not been proven (or disproven) in this study, and their relationship to the ijolite-urtite body which encloses the carbonatite stock in the central complex is also uncertain. What is known of the contact zones between the ijolite and melilitolite bodies suggests that the contacts are usually gradational (Ronenson et al., 1978) indicating contemporaneous intrusion, and possibly reaction between the two magma types. A

contemporaneous relationship between the outer ijolite ring and the main melilitolite bodies means that the ijolites may also have intruded as ring dykes around the outside of the melilitolites while the ijolite body enclosing the carbonatites and phoscorites in the middle of the central complex ascended contemporaneously through the main central vent. This implies that there might have been just one generation of ijolite intrusion, rather than the two-stage pre-melilitolite and post-melilitolite intrusion sequence previously described (Bulakh & Ivanikov, 1984; Bell et al., 1996). However, the modal mineralogy of the ijolite “ring dykes” and the ijolites surrounding the carbonatite stock within the central complex are dissimilar. The mafic minerals in the outer ijolite bodies are dominated by pyroxene, whereas the central ijolite body is dominated by garnet $\pm$ phlogopite, indicative of a (pyroxene-)ijolite - carbonatite reaction (Verwoerd, 1978).

Debate has surrounded the issue of crystal settling in carbonatites ever since the first viscosity measurements were made on carbonatite magmas (Trieman & Schedl, 1983; Pinkerton et al., 1995 and refs. within). Their incredibly low viscosity, should, in theory, mean that crystal settling is an inescapable phenomenon in carbonatites, and phoscorites have frequently been put forward as classic “carbonatite cumulates” (Andersen, 1987). However, the isotopic signature of the Turiy phoscorites suggests that this is not the case, and phoscorites should generally be considered as separate melts in their own right (Philpotts, 1967). The method of intrusion of such low-viscosity melts was a topic of recent debate at the 1997 GAC-MAC Symposium on Carbonatites. The general consensus was that such magmas were intruded as “crystal mushes” which may be consistent with the intrusion mechanisms, relationships and textures for plutonic carbonatites, but inconsistent with textures seen in extrusive carbonatites such as Kaiserstuhl (Keller et al., 1990) and Oldoinyo L’Engai (Pinkerton et al., 1995), and does not resolve the issue of how and where the plutonic “crystal mushes” were formed. It

should be remembered, however, that dense minerals (e.g. magnetite) will only settle within a magma if they have actually crystallized prior to eruption or intrusion. Treiman & Schiedl (1983) showed that if (turbulent) convection was fast enough within the carbonatite melt, crystallization would occur to a significant extent prior to settling. Additionally, Treiman & Schiedl (1983) stated that calcite is easily deformed, leading to “gneissic textures” and strained extinction. These features are seen in particular in the northern complex carbonatites, although the claim by Treiman & Schiedl (1983) that primary calcite textures are probably never seen due to the ease of deformation, is not believed to be true in Turiy, where a range of clearly-defined textures are seen in the carbonatites, and described in detail in Chapter 5.

8.5 Melt structure

Different rock types at Turiy contain very variable proportions of network-forming and network-modifying cations. The major network-forming cations are traditionally listed as Si, Al, Fe^{3+} , Ti and P, whilst Ca, Mg, Fe^{2+} , alkalies, and also Al and Fe^{3+} are accepted as network-modifying cations. It should be noted that Al, Fe^{3+} and Ti can be present in both forms in a melt, depending on a variety of chemical and physical parameters. In addition, the alkalies (and less commonly, the alkaline earth elements) can occur in a pseudo network-forming role, in charge-balancing the substitution of trivalent cations for tetravalent Si [$\text{NaAl} = \text{Si}$].

Hess (1980), in a comprehensive review of silicate melt-structures, listed three types of Si-O bonds: O^0 (Si-O-Si), O^- (Si-O-M) and O^{2-} (M-O-M). At the point where SiO_2 mole % represent less than a third of the cations in the melt, the number of O^{2-} (free oxygens) exceeds the number of O^0 (bridging oxygens) for the first time (Figure 8.4). This 33% Si corresponds to approximately the minimum percentage of Si present in natural silicate magmas, commonly represented by extrusive melilitites and many of the

Figure 8.4 Sketch of oxygen co-ordination as a function of SiO_2 content (after Hess, 1980). The critical boundary at 33 mol.% SiO_2 below which “free” oxygens (M-O-M) are more abundant than “bridging” oxygens (Si-O-Si) represents the cut-off for natural, homogeneous silicate magmas, as seen in Figure 8.1. K = equilibrium constant:

$$K = \frac{[\text{O}^-]^2}{[\text{O}^\circ][\text{O}^{2-}]}$$

melilite-bearing rocks at Turiy (Table 7.A). Masson et al. (1970) claimed that melts with $< 33\%$ SiO_2 were equivalent to ionic liquids, dominated by monomeric silicate structures. There has been a tendency in the past to assume that a true carbonatite contained $< 10\%$ SiO_2 and that compositions with $> 10\%$ SiO_2 had assimilated silicate phases (Woolley & Kempe, 1989). However, Dalton & Presnall (1997) showed experimentally that carbonatite melts with up to 25% SiO_2 could be legitimately produced by melting within the dolomite and magnesite stability fields in the mantle. Thus the existence of an apparent “no-go” zone for natural magmas derived under equilibrium conditions between ~ 10 - 33% Si has now been thrown into doubt. Perhaps mantle melts formed primarily of a silicate framework “end” at 33% Si, and Si-bearing carbonatites dominate at Si values $< 33\%$? If this is the case, then it becomes all the more critical to understand the nature of the transition across the 33% Si boundary, and how it affects the evolution of Si-poor magmas. Figure 8.5 shows the relationship between Si content of each of the bulk rock analyses of the silicate rocks listed in Table 7.A and the Si content of all of the silicate phases contained within these samples, as determined in Chapter 5 and listed in Appendix J. This diagram shows very clearly that the bulk-rock Si content of the melilitolites is the lowest of all the silicate rock-types analyzed from Turiy (except the turjite sample TU 119 which has already been described as a reaction product, and is believed to represent disequilibrium crystallization conditions), and these bulk-rock Si values fall very close or below the Si content of the all of the major phases precipitating out of the magma. In theory, the melilitolites can cross the 33% Si boundary by fractional crystallization of any silicate phases precipitating out of the magma. It is only the precipitation of non-silicate phases (magnetite, apatite, perovskite, calcite) in these samples that maintains the bulk-rock composition at $> 33\%$ Si. The other silicate samples (ijolite, olivine melteigite) contain a greater abundance silicate mineral phases with Si contents less than that of the bulk-rock composition, and will thus easily fractionate to Si-rich compositions, as

indicated by a petrographic study of the samples. If such fractionation occurs, Figure 8.5 shows us that the ability of a silicate magma to cross the Si = 33% boundary (from higher to lower values) due to fractional crystallization of silicate phases, is most likely to occur in the melilitolites at Turiy. Figure 8.5 shows that fractionation of the major silicate phases from any of the other silicate magmas within Turiy is likely to maintain the composition of the residual silicate melt at > 33% Si. Thus, it is feasible that a melilitolite could cross the 33% Si boundary and break down into a number of structurally separate liquids or melt components, such as a carbonate phase (removing Ca, Sr, Na, K, $\pm$ Mg, Fe), a phosphate phase (removing P, Ca, REE), an oxide phase (Fe, Ti), a sulphide phase (S, Fe, Cu) (where the carbonate, phosphate, oxide and sulphide phases are the dominant components of the phoscorites and carbonatites), leaving the remaining melt relatively enriched in Si (e.g. an ijolite). This is essentially a potential group of immiscible liquids created by a fractional crystallization process. There is evidence that this process may have happened, at least in part, in the northern complex carbonatites at Turiy, where magnetite-melilite-calcite assemblages are seen associated with melilitolites and melilite-bearing carbonatites. The origin of the phoscorites in the central complex may also be related to such a process, but a suitable parental silicate magma has not yet been identified.

The abundant melilitolites at Turiy rocks an excellent opportunity to try to answer another question that has not yet been fully answered: “why does melilite form?” The dimeric structure of melilite is unusual in the mineral world. Does the silicate melt structure govern the formation of dimeric melilite or is alkali-saturation in melilitites responsible for the formation of a mineral more CaO-rich than clinopyroxene? Experimental evidence thus far points to both factors being important (Yoder, 1973; Yoder, 1976; Lloyd, 1985), while Mysen (1976) points to the fact that the presence of Fe^{3+} in an SiO_2 -undersaturated melt stabilizes the solubility of CO_2 in that melt (which

Figure 8.5 Relationship between the Si content (atom %) of the silicate whole-rock samples and the silicate minerals contained within them. Bulk rock SiO_2 contents from Table 7.A (converted to atomic Si and plotted along the x-axis) and mineral compositions from Appendix J. The size of each of the crosses within the boxes corresponds to the relative abundance of each mineral within the sample. The location of each cross corresponds to the average Si content of that particular mineral. It can be seen that the melilitolites and turjite are the only samples where the bulk rock Si contents are approximate equal to or lower than all the Si content of the minerals contained within, with the possible exception of nepheline. Hence, the melilitolites (and possibly the turjite, which is itself believed to be a disequilibrium reaction product) are the most likely silicate magmas analyzed from Turiy to fractionate to Si-poor residual liquids.

artificially enhances polymerization). Subsequent saturation in Fe^{3+} (removed in the form of oxide phases) initiates a reversal in this process with the release of CO_2 , depolymerization of the melt and formation of melilite. The introduction of excess CO_2 and Ca into the melt can be linked to the presence of carbonate phases whose presence and influence are seen clearly in the Turiy massif.

In the investigation of silicate melt structures, it is common practice to mix Si with one or two other element oxides $\pm$ volatile components, and to note the changes in the structural units present in the silicate melt, as function of varying pressure, temperature and chemical composition. One of the features of mineral chemistry seen in Chapter 5 was the almost exclusive appearance of $^{\text{vi}}\text{Al}$ in minerals associated with melilite. It is therefore important to understand how $^{\text{vi}}\text{Al}$ is created in aluminosilicate melts. Mysen et al. (1985) demonstrated that Al was always in tetrahedral co-ordination in Na-Ca-Al-Si-O melts. Addition of extra Al (increasing the $\text{Al}/(\text{Al}+\text{Si})$ ratio), and increasing the $\text{M}^+/(\text{M}^++\text{M}^{2+})$ (i.e. $\text{Na}/(\text{Na}+\text{Ca})$ ratio merely increased the amount of Al in tetrahedral co-ordination, forming 3-D aluminosilicate melts. However, it should be noted that the experimental charges used to determine these characteristics did not contain the low amounts of Si pertinent to the Turiy melilitolites, and it is not impossible that the aluminosilicate behaviour will change under conditions of low silica activity. For example, Table 7.A shows that the melilitolites (containing $^{\text{vi}}\text{Al}$ -bearing minerals) have the highest Al/Si ratio of any of the rock-types analysed, indicating that $^{\text{vi}}\text{Al} \propto \text{Al}/(\text{Al}+\text{Si})$ ratio at Turiy. Additionally, the mineral melilite has the lowest Al : Si ratio of all the silicate minerals found in the Turiy rocks, according to the compositions listed in Appendix J, and precipitation of melilite will cause the Al/Si of the residual melt to increase. NMR-IR studies by Poe et al. (1992) suggest that the Al/Si ratio may be proportional to the $^{\text{vi}}\text{Al}/^{\text{iv}}\text{Al}$ ratio, contrary to the results of Mysen et al. (1985) quoted above. Mysen et al. (1985) explained their results in terms of Al preference for sites with

the lowest possible T-O-T angle, i.e. tetrahedral co-ordination. Na (or any other alkali or alkaline earth cation) is needed to associate with Al in any tetrahedral site where it is substituting for Si, in order to maintain charge-balance ($\text{NaAl} = \text{Si}$). The effectiveness of this charge-balancing mechanism, and hence abundance of such cation pairs, is dependent on the identity of the charge-balancing cation available and the effectiveness decreases in the following order: $\text{K} > \text{Na} > \text{Li} > \text{Ca} > \text{Mg}$ (Navrotsky et al., 1982). Increasing pressure or temperature do not create $^{\text{vi}}\text{Al}$. Mysen et al. (1981) reported that if $\text{Al} > [\text{alkaline earth} + \text{alkali cations}]$, then $^{\text{vi}}\text{Al}$ can be formed in the form of octahedral AlO_6 triclusters. Such a scenario does not appear to occur in Turiy, according to bulk rock chemistry (Table 7.A), and thus another mechanism for creating $^{\text{vi}}\text{Al}$ must be sought in order to explain the mineral chemistry of the melilitolites. The presence of Fe^{3+} in tetrahedral co-ordination also requires the presence of an alkali or alkaline earth cation for charge-balance, like Al, but formation of $^{\text{iv}}\text{Al}$ is energetically more favourable than creation of $^{\text{iv}}\text{Fe}^{3+}$. Another possible way to try to move Al into octahedral co-ordination is to remove the cations which provide its charge-balance in tetrahedral co-ordination. Mysen et al. (1980) and Mysen & Virgo (1980, 1985) report that addition of H_2O , CO_2 or F to a melt will cause the preferential formation of NaOH , Na_2CO_3 or NaF complexes within the melt, thereby destabilizing Al in tetrahedral co-ordination, due to inability of the melt to provide a charge balancing mechanism for the Al-Si substitution ($\text{NaAl} = \text{Si}$), and causing the conversion of $^{\text{iv}}\text{Al}$ to $^{\text{vi}}\text{Al}$. Taylor & Green (1987) report similar behaviour using CH_4 instead of H_2O , with the creation of minerals rich in $^{\text{vi}}\text{Al}$ under mantle conditions, and enlargement of the olivine stability field due to depolymerization. If the results of Mysen et al. (1980) and Mysen & Virgo (1980, 1985) can be extrapolated to the formation of $^{\text{vi}}\text{Al}$ -bearing minerals in the melilitolite magmas, as the results of Taylor & Green (1987) indicate, then it would suggest that the melts were particularly enriched in dissolved volatiles at the point of crystallization. The effects of Tschermak-related charge-balance

mechanisms on Al co-ordination are not well constrained, although formation of Al-Ti complexes (Mysen et al., 1980) and Al-P complexes (Mysen et al., 1981) have been reported. There may also be an association between Al-Fe³⁺ in alkali-free systems

The behaviour of another network-forming cation, Ti, is also important as Ti occurs in a variety of important silicate (garnet, pyroxene) and non-silicate (magnetite, perovskite) minerals in the Turiy rocks. In melts, Ti occurs in both tetrahedral and octahedral co-ordination, as TiO₂ clusters, as Si-Ti substitutions in sheet structures (Mysen & Neuville, 1995), and forms complexes with Al in aluminosilicate melts: $\text{AlSi} + \text{Ti} = \text{Si} + \text{AlTi}$ (Mysen et al., 1980). Increasing Ti content within the melt leads to decreased polymerization within the Ti-bearing structural units. The fraction of ^{vi}Ti in the melt systems investigated never increases above about 6% total Ti, and the ^{vi}Ti : ^{iv}Ti ratio is temperature sensitive (Mysen & Neuville, 1995). In the presence of Na, changes in polymerization may occur, including formation of dimeric silicate units and ^{vi}Ti. One of the interesting points of the results of Mysen & Neuville (1995) is the dissimilarity between the dominance of ^{iv}Ti in melts, and the exclusive ^{vi}Ti co-ordination in silicate minerals, such as garnet (Locock et al., 1995), where the latter assumption has been used in the calculation of mineral formulae in Appendix J. Thus, the experimental results on the structural setting of Ti in a silicate melt appear to be at odds with its known co-ordination state in silicate minerals.

The ubiquitous, but extremely variable abundance of apatite in the Turiy rocks also requires explanation. Apatite formation is dominantly controlled by P and Ca saturation within the corresponding melt. Mysen et al. (1981) found that P occurs mostly in tetrahedral co-ordination within discrete structural units within silicate melts in the form of M-PO₄ units, although P-O-Si bonds may be possible in depolymerized melts. The removal of M cations (alkali, alkaline earth or Al) from the silicate part of the melt to form the phosphate complexes results in the increased polymerization of the residual

silicate melt. Mysen et al. (1981) suggest that the enrichment of P in late-stage liquids in magmas may result in the formation of immiscible phosphate liquids, an interesting observation in the light of apparent apatite veining in some carbonatites and phoscorites at Turiy.

Fe can occur in two oxidation states as well as two different co-ordination states which makes its behaviour in melts more challenging to quantify. It also provides a link between oxygen fugacity and its ability to vary between network-forming and network-modifying roles within the melt. In natural silicate minerals, Fe^{3+} can occur in both octahedral and tetrahedral co-ordination, while Fe^{2+} occurs almost exclusively in octahedral co-ordination except for minerals such as melilite. In melts, Fe^{2+} is always octahedrally co-ordinated and acts as a network-modifier (Mysen et al., 1984). Of particular relevance to the melilitolite-ijolite relationship at Turiy, in terms of the varying abundance of iron oxide and iron silicate minerals between the two rock-types, is the fact that the co-ordination state of Fe^{3+} varies depending on the oxygen fugacity and degree of polymerization of the melt. Fe^{3+} is in octahedral co-ordination for $\text{Fe}^{3+}/\text{Fe}_{\text{TOTAL}} < 0.4$, and in tetrahedral co-ordination for $\text{Fe}^{3+}/\text{Fe}_{\text{TOTAL}} > 0.5$. There is a transition zone between $\text{Fe}^{3+}/\text{Fe}_{\text{TOTAL}} = 0.4-0.5$ where Fe^{3+} changes from octahedral to tetrahedral co-ordination (Mysen et al., 1984). Like Al, Fe^{3+} will preferentially enter tetrahedral co-ordination in a fully polymerized silicate melt, causing subsequent depolymerization of the silicate structure (Mysen et al., 1984). When Fe^{3+} is in tetrahedral co-ordination, it needs to be stabilized by an alkali cation (normally Na) to provide charge-balance for the tetrahedral $\text{NaFe}^{3+} = \text{Si}$ exchange reaction, as mentioned earlier. The change in the $\text{Fe}^{2+} : \text{Fe}^{3+}$ ratio is also linearly dependent on the NBO/T ratio for any given alkaline earth cation in the system. The depolymerization of the melt results in the oxidation of Fe^{2+} to Fe^{3+} (Mysen et al., 1984) although the addition of Fe^{3+} to an Fe-free melt can also result in increased polymerization, since Fe^{3+} will exist in tetrahedral co-ordination under those

conditions. Changing temperature of the melt also has an effect on the oxidation state of Fe with Fe^{2+} becoming more abundant with increasing temperature (Mysen et al., 1980). The links described here between the oxidation and co-ordination states of Fe with melt polymerization may be critical factors in the evolution of the Turiy magmas, particularly in the melilitolites and ijolites, given the variations in their Fe partitioning behaviour.

Having looked at the behaviour of individual network-forming cations in silicate systems, looking at melt behaviour with combinations of these cations may be additionally informative. A melt system applicable to the Turiy phoscorites is the Ca-Fe-P-Si-O- CO_2 system investigated by Mysen (1992). Of interest here is the fact that solution of phosphate ions in a silicate melt increases the $\text{Fe}^{3+}/\text{Fe}^{2+}$ ratio and enhances the stability of Fe^{3+} . This relationship between P and Fe may also relate to the interesting intergrowth textures between magnetite and apatite in the Turiy carbonatites and phoscorites, as described in Chapter 4.

Much has also been written in the literature regarding the relationship of the various silicate melt components to each other, where equilibrium reactions can be written between the different silicate components listed in Table 8.A. There are many such reactions, but one of particular interest to this study is the reaction:

Extrapolated to mineral compositions, this translates into: 2 pyroxene = melilite + mica, which represent the dominant ferro-magnesium silicate components of the ijolites and melilitolites respectively. Nepheline is common to both rock-types and does not feature in the silicate equation. The reasons for moving the equilibrium position to the left or right of this reaction within the Turiy magmas are not clear at this time, but warrant further investigation.

Of interest in the interpretation of the petrogenesis of the Turiy carbonatites is the finding that CO_2 preferentially combines with Ca rather than Mg in alkaline earth -

bearing silicate melts (Holloway et al., 1976), in terms of a positive correlation between the amount of dissolved CO_2 and the $\text{Ca}/(\text{Ca}+\text{Mg})$ ratio of the melt investigated. Mysen & Virgo (1980), in a study of CO_2 -bearing aluminosilicate systems, suggest that this is due to the activity of Ca^{2+} in the melt rather than the degree of polymerization of the melt, suggested by Holloway et al. (1976). This $\text{Ca} > \text{Mg}$ preference for CO_3^{2-} anions was borne out by the study of Dalton & Presnall (1997) who showed experimentally that the $\text{Ca}:\text{Mg}$ ratio of melts derived from the dolomite and magnesite stability fields within the mantle was not noticeably different, despite a significant change in $\text{Ca}:\text{Mg}$ ratio in the carbonate source ($(\text{CaMg})\text{CO}_3 - \text{MgCO}_3$). This is in contrast to Wallace and Green (1988) who found that the Mg content of primary carbonatite melts increased with increasing pressure. The results of Holloway et al. (1976) and Dalton & Presnall (1997), regarding preferential combination of Ca with carbonate ahead of Mg, may well explain why most carbonatite complexes contain early calcite carbonatites, followed by later dolomite-calcite and dolomite carbonatites, as opposed to the typical viewpoint that Mg-bearing phases must always be the first to crystallize, and are a sign of a "primary" melt. Certainly, for silicate phases, where there is $\text{Mg} - \text{Fe}^{2+}$ substitution and exchange, this may be true, but in carbonatites the Ca-Mg exchange in the carbonate phase is controlled primarily by the activity of Ca in the melt. This fact may also be of importance to the $^{\text{vi}}\text{Al}$ feature of the melilitolite magmas at Turiy, as preferential removal of CO_2 to combine with the alkaline earth elements, which are so abundant in the melilitolites, will create a melt rich in dissolved $\text{H}_2\text{O} \pm \text{F}$ which can subsequently complex with the alkali metals and destabilize $^{\text{iv}}\text{Al}$. The precipitation of iron oxides will also cause release of dissolved CO_2 into the melt (Mysen et al., 1976) as mentioned earlier, and this will enhance the removal of Ca and Mg from any charge-balancing roles with tetrahedral Al.

It is obvious that this summary of some of the melt structure investigations carried out by a small number of researchers over the last few years is a very brief look at an

extremely complex topic. However, there are a number of extremely interesting points relevant to our understanding of the Turiy rocks to be drawn from the material above:

1. P, Ti and Fe^{3+} do not easily substitute for Si in silicate network structures, but prefer to remain as discreet complexes within the silicate melts, possibly existing as immiscible liquids. Coincidentally, two of these network forming cations, P and Fe^{3+} , are concentrated in the Turiy phoscorites.
2. The addition of CO_2 , H_2O or F to alkali or alkali-earth-bearing aluminosilicate melts causes the removal of charge-balancing cations from the exchange $(\text{Ca})\text{NaAl} = \text{Si}$, causing conversion of $^{\text{iv}}\text{Al}$ to $^{\text{vi}}\text{Al}$. It is also possible that high Al/Si and peralkaline melts may contain $^{\text{vi}}\text{Al}$. The “volatile - alkali” scenario seems the most realistic explanation for the creation of the $^{\text{vi}}\text{Al}$ -bearing minerals unique to the melilitolites at Turiy.
3. There is a direct relationship between melt saturation in Fe^{3+} , solubility of CO_2 in the melt, and the formation of melilite. This correlates to the significant presence of magnetite in the Turiy melilitolites.
4. The equilibrium reaction $2\text{Si}_2\text{O}_6^{4-} = \text{Si}_2\text{O}_7^{6-} + \text{Si}_2\text{O}_5^{2-}$ goes from right to left in the ijolite rocks and from left to right in the melilitolites, but the cause of the shift in the equilibrium position is not yet clear in the Turiy rocks.
5. CO_2 , dissolved as carbonate anions in a silicate melt, will preferentially combine with Ca instead of Mg. This may explain the sequence of carbonatite formation seen in many carbonatite massifs, in terms of calcite carbonatites preceeding dolomite-bearing carbonatites.

8.6 Petrogenesis

Given all the information accumulated and discussed thus far in this thesis, it should be possible to identify chemical signatures and magmatic histories for each of the groups of rock-types that have been identified. The assumption is made that the samples within each group have shared a similar source, and/or crystallization pathways during their history, although this does not mean that they are necessarily co-magmatic, or represent conjugate liquids.

8.6a The “depleted” silicate magmas

The first grouping to be discussed is the “depleted” group of silicate rocks. The term “depleted” refers to the position of the isotope data from the Turiy samples on the Sr-Nd isotopic anti-correlation plot (Figure 6.4). On the basis of this first criteria, the group includes the pyroxenites, and melilitolites, discounting, for the moment, the possible effect of crustal contamination on some of the melilitolite samples analyzed from the satellite complexes. The similarity in overall trace-element spidergram patterns corroborates the pyroxenite-melilitolite association, and, on the basis of trace element data, the melilitite dyke TU 107 is included in this “depleted” group. Similarities in mineralogical compositions between the pyroxenites and melilitolites provides a third piece of evidence linking these samples within the “depleted” group. It should be emphasized that the variations in ages obtained between the samples in this group, as mentioned above, means that they cannot all be comagmatic, despite their similarities in chemical characteristics, but that likely share a similar source, and/or crystallization histories.

The melilitolites have initial $^{87}\text{Sr}/^{86}\text{Sr}$ ratios of 0.70362-0.70407 and initial $^{143}\text{Nd}/^{144}\text{Nd}$ ratios of 0.51243-0.51233, when back-corrected to 373 Ma. This main grouping does not include an apatite analysis from the eastern complex melilitolite body

from sample E.133.30 with ratios of 0.70374 and 0.51208. Mineral separates from pyroxenite sample C.90.29, also back-corrected to 373 Ma, have initial $^{87}\text{Sr}/^{86}\text{Sr}$ ratios of 0.70365-0.70367 and initial $^{143}\text{Nd}/^{144}\text{Nd}$ ratios of 0.51235-0.51231. It has therefore been suggested that sample C.90.29, for which no isochron information is available, may be comagmatic with the other Devonian rocks of the massif, and may share the same source as the melilitolites, given that its isotopic ratios are the same as those of the most depleted melilitolites. The mineral separates from pyroxenite sample C.AG.347, from which a 588 Ma age was obtained from an Rb-Sr isochron, as explained in Chapter 7, have an initial $^{87}\text{Sr}/^{86}\text{Sr}$ ratio of 0.70343 and initial $^{143}\text{Nd}/^{144}\text{Nd}$ ratios of 0.512224-0.512345. The data from the calcite separate from this sample, which was not used in the isochron determination, has initial ratios of 0.70346 and 0.512100. As seen in Chapter 7, the epsilon Sr and Nd values for both pyroxenite samples [C.AG.347 & C.90.29] creates a vertical line in $\epsilon_{\text{Sr}}(\text{T}) - \epsilon_{\text{Nd}}(\text{T})$ space, with $\epsilon_{\text{Sr}}(\text{T})$ values ranging from -5.64 to -5.87 and $\epsilon_{\text{Nd}}(\text{T})$ values from 3.06 to 8.01. The epsilon values for the melilitolites are $\epsilon_{\text{Sr}}(\text{T})$ -0.22 to -7.9 and $\epsilon_{\text{Nd}}(\text{T})$ 2.66 to 4.59, not including the $\epsilon_{\text{Nd}}(\text{T})$ value of -2.1 from the apatite separate from sample E.133.30. It can, therefore, be seen that the melilitolites have much more variation in $\epsilon_{\text{Sr}}(\text{T})$ space than $\epsilon_{\text{Nd}}(\text{T})$ space, contrary to the trend shown by the data from the pyroxenites. The melilitolite data may be explained by crustal contamination, as discussed in Chapter 7, but the vertical range in the pyroxenite data is harder to explain. (There are no isotope data for sample TU 107 [melilitite dyke] which is also included in this “depleted” group of samples).

There appear to be three possible explanations for the vertical trend in the pyroxenite data:

- 1) that the spread in $\epsilon_{\text{Nd}}(\text{T})$ values is due to experimental error. This is believed to be unlikely, given the consistency of the rest of the data and the blank values quoted in Appendix E.

- 2) that the pyroxenite samples have been derived from a depleted source whose isotopic characteristics evolved parallel to those of bulk Earth but not CHUR over time. This scenario requires the rather complex, multiple evolution of an older, depleted reservoir (Allegre, 1979), but is, in theory, possible.
- 3) that the rocks have crystallized under open-system conditions, given that the spread of $\epsilon_{\text{Nd}}(\text{T})$ values are seen from mineral separates taken from just two hand samples. This scenario requires that the samples exhibited open-system behaviour with respect to the REE, but closed-system behaviour with respect to the Rb-Sr system, during crystallization. This implies REE-fractionation within the crystallizing melt during the sequential crystallization of the various minerals - a fractionation effect, possibly involving volatile phases, which did not affect the Rb-Sr system.

The pyroxenite and melilitolite data overlap in an area around $\epsilon_{\text{Sr}}(\text{T})$ -5.5 to -8 and $\epsilon_{\text{Nd}}(\text{T})$ +3 to +5. While recognising that the phoscorites (and some carbonatites) are more depleted than these silicate rocks, the pyroxenite and melilitolites represent the most depleted silicate magmas, and for now, will be taken to represent the original isotopic signature of the “depleted group”.

The possibility that crustal contamination was involved in the production of enriched $^{87}\text{Sr}/^{86}\text{Sr}$ isotopic signatures in samples from the satellite complexes was discussed in detail in Chapter 7. It was proposed that the contamination occurred as the result of partial melting and selective dissolution and assimilation of wall-rock charnockites during the formation of the satellite complexes, and the plumbing systems beneath, as proposed by Andersen (1987) to explain similar contamination effects in the Fen carbonatite complex. The Nb/U ratio was used by Hofmann et al. (1986) and Simonetti & Bell (1997) as an indicator of crustal contamination. This ratio is 47 ± 10 in MORBs and OIBs (Hofmann et al., 1986) and approximately 10 in average continental crust (Weaver & Tarney, 1984). The Nb/U ratios at Turiy range from 9 in the dolomite-

calcite carbonatite [C.DC] to 96 in the pyroxenite [C.90.29]. The Nb/U values for the melilitolites range from 67 to 83, while the turjite Nb/U = 76 and the fenite = 43. These values, along with absolute Nb and U values, are not consistent with crustal assimilation in the samples analyzed, most of which come from the central complex. However, the wide variety of Nb/U values also suggest that the Nb/U ratio is not a good indicator of crustal contamination in carbonatitic magmatic systems.

The trace element signatures of this pyroxenite-melilitolite-nephelinite group of rock-types have been compared and discussed in Chapter 7 (section 7.4a). The pyroxenite trace element data from sample C.90.29 are very similar to the data from the olivine-phlogopite-pyroxenites from Jacupiranga (Fig. 7.11b) (Huang et al., 1995). However, the pyroxenites at Jacupiranga have enriched isotopic signatures (just to the right of the bulk earth line) and are not associated with melilitolites, unlike the Turiy pyroxenites which have depleted isotopic signatures and are related, both spatially, chemically and mineralogically to the melilitolites.

The Turiy pyroxenite sample C.90.29 (believed to be Devonian in age) is depleted in LILE compared to the melilitolites, but shares a similar trace element pattern for the more compatible elements. The higher concentration of the trace elements in the melilitolites may be due either to a smaller degree of partial melting of the same mantle source, or a greater degree of metasomatic enrichment at some point in the magmatic history, if metasomatic fluids or melts were responsible for the trace element signatures of the pyroxenites and melilitolites. However, in addition to possible variations in the degree of partial melting and/or metasomatism of the mantle source, some fractional crystallization has almost certainly taken place to create the very low Zr-Hf-HREE signatures of the melilitolites, probably created by garnet and/or pyroxene fractionation.

The overall, enriched, trace-element patterns but depleted isotopic signatures of the samples in the depleted group suggest initial derivation from a source with a depleted

isotopic signature, which may have been previously metasomatised by trace element-rich fluids or melts. It has been suggested that this source lies at the base of lithosphere, within the garnet-dolomite stability field (Wilson et al., 1995), and that the metasomatic fluids may have been asthenospherically-derived carbonatites. Barring effects of fractional crystallization, the trace element spidergrams bear an overall resemblance to the trace element characteristics of HIMU, a mantle component that many have attributed to asthenospherically-derived magmas related to mantle plumes (e.g. Weaver, 1991; Wilson et al., 1995). Comparison of a variety of trace element ratios from the melilitite dyke [TU 107] from the south coast, with typical values for HIMU-derived OIBs (e.g. St Helena, Tubaii, etc) (Weaver, 1991), shows that melilitite sample TU 107 matches HIMU very closely in terms of trace element characteristics, with some similarities to N-MORB. More importantly, such HIMU signatures, which exhibit depletion in HREE and K, and enrichment in LREE and the incompatible HFSE, are typically associated with carbonatitic metasomatism of the mantle source (Rudnick et al., 1993). This is linked to the fact that the ratios from Weaver (1991) which show least correlation between HIMU and sample TU 107 are related to an unusually high degree of enrichment of La in sample TU 107, where enrichment in the LREE is commonly associated with carbonatite metasomatism. Due to the depleted isotopic signatures of the members of this group of samples, and enriched trace element signatures, it is suggested that the mantle source of the “depleted group” is located within the garnet-dolomite-phlogopite stability field in the asthenosphere, contains residual phlogopite and garnet (depleted K and HREE), and was subjected to carbonatite metasomatism (high trace element contents consistent with a carbonate-rich host fluid) shortly prior to magma formation and eruption.

The REE patterns of the melilitolites should also be considered in the light of the results of Haas et al. (1995) discussed in Chapter 7, where it was shown that the LREE complex preferentially with Cl^- , the HREE with F^- , and there is no partitioning preference

for the OH^- group. It is possible that halogen-rich fluids have also played a part in the production of the exceptionally high Ce/Lu ratios of these magmas during fractional crystallization. If this is the case, then it would suggest that the melilitolites may be enriched in Cl-bearing fluids (LREE rich), while the loss of an F-HREE-rich fluid would explain the extreme HREE depletion in these samples.

The nature of the petrogenetic relationship between the pyroxenites and the melilitolites can be explained in one of three ways:

1) the Precambrian pyroxenites represent the base of an older intrusion which would have overlain the location of the Devonian intrusion, and has since been eroded away during crustal uplift. Given the similarity in isotopic signature between the Precambrian pyroxenite sample C.AG.347 and the Devonian(?) pyroxenite sample C.90.29 and the most depleted melilitolites, it is likely that the magma which produced sample C.AG.347 may have tapped a similar mantle source earlier in the geological history of the area. This model helps to explain why the Devonian central complex samples show no signs of crustal contamination in their isotopic signatures, as the Precambrian magmas may have created a plumbing system which became “insulated” from the country rock which could be re-utilized by the Devonian magmas. The satellite complexes, along with their upper crustal plumbing systems, appear to have been created during the Devonian episode of intrusion and incorporated crustal material during that time.

2) Devonian pyroxenites, possibly represented by sample C.90.29, represent cumulates associated with the Devonian magmatism. This implies that a dense, cool cumulate was intruded during the first phase of magmatism which seems unlikely as the isotopic data infer the lack of a convecting magma chamber beneath Turiy. The pyroxenite bodies, which almost certainly follow the stockwork pattern of intrusion shown by the other rock types in the central complex (Ronenson et al., 1978), therefore cannot represent cumulates from the stratigraphically equivalent melilitolite bodies within the massif.

Despite isotopic similarities, the melilitolites do not fit the trace element criteria for a conjugate liquid for the pyroxenites, and the mineralogical compositions described and discussed in Chapter 6 do not provide evidence for fractionation relationships between the two rock types.

3) both suites of pyroxenites represent a high-pressure version (>2 kbar) of a melilite-nephelinite magma, as suggested by the experimental data of Tilley & Yoder (1968), and supported subsequently by the experimental results of Yoder (1973), Brey & Green (1975, 1977) and Brey & Ryabchikov (1994). This suggestion explains the absence of cumulus textures, the overlapping mineral compositions between the melilitolites and pyroxenites, the similarities in the trace element data (although not the low LILE contents of the pyroxenites), and the isotopic similarities. This model creates problems in massifs where pyroxenites occur in association with olivinites, as, unlike the pyroxenite-nephelinite link, there is no obvious low-pressure equivalent of olivinites.

All these models explain some of the features associated with the pyroxenites. The isotopic data also show that the magmas in the isotopically depleted group of Turiy silicate rocks have not maintained closed-system conditions during their history which makes estimates of conjugate relationships, if they exist, difficult to quantify.

Amongst the myriad of mineralogical data plotted, described and discussed in Chapter 6 (and listed in Appendix J), some essential criteria stand out with respect to the melilitolites and pyroxenites. The most obvious characteristic is the behaviour of Al in the Al-bearing minerals in these samples. The melilitolites, pyroxenites and melilite-bearing/associated carbonatites, are the only samples which contain ^{vi}Al in minerals such as garnet and mica. Al contents are high in pyroxene and magnetite. The most likely reasons for this are related to bulk rock Al/Si ratios which are high in the pyroxenites and melilitolites, and the role of volatiles (OH^- , F^-) in removing alkalis from their role of charge-balancing tetrahedral $\text{NaAl} = \text{Si}$ exchange within the melt, hence forcing Al into

octahedral co-ordination. Other distinctive features from the mineral composition data are that Si is more abundant than Sr in apatite from the melilitolites and pyroxenites, and perovskite from these samples contains REE in excess of that needed to charge-balance Na substitution. The high REE concentrations in perovskite from these samples also reflects the high REE content of the bulk rock analyses.

A number of the silicate rocks at Turiy contain cancrinite (Appendix Jvi), which is commonly assumed to be secondary in origin. Melilitolite sample TUR I contains no nepheline but 16% cancrinite and virtually unaltered melilite. Melilite is one of the first minerals to be extensively altered due to post-magmatic hydrothermal activity, and yet this is not seen in sample TUR I. The study of the nepheline-calcite-H₂O system by Watkinson & Wylie (1971) showed that the reaction of nepheline and calcite produced cancrinite under hydrous conditions, and it was thus postulated by Bell et al. (1996) that cancrinite can form as a primary product of the crystallization of a Ca-rich, SiO₂-poor silicate magma. Cancrinite occurs in both the melilitolite and ijolite rocks, and, given the Ca-saturated nature of most of these rocks, it is suggested that cancrinite may be a primary mineral in many of the samples, especially when the rocks show no sign of alteration.

8.6b The “enriched” silicate magmas

Samples in this group have the most enriched isotopic signatures in the Sr-Nd anti-correlation diagram. The majority of their $^{87}\text{Sr}/^{86}\text{Sr}$ values lie between 0.70344-0.70416 with $^{143}\text{Nd}/^{144}\text{Nd}$ values between 0.51217-0.51230. They fall near the bulk earth - CHUR intersection, and include the olivine melteigite [C.AG.48A], several of the ijolite series rocks, including samples C.AG.122 and C.IJ.14, and the turjite, represented by sample TU 119. It should be noted that these samples fall close to the isotope data from the Terskii Coast kimberlites ($^{87}\text{Sr}/^{86}\text{Sr} = 0.70385\text{-}0.70402$ and $^{143}\text{Nd}/^{144}\text{Nd} = 0.51207\text{-}$

0.51224), and that the analyses from the turjite [TU 119] and west coast ijolite [WKM] show more enriched $^{87}\text{Sr}/^{86}\text{Sr}$ but similar $^{143}\text{Nd}/^{144}\text{Nd}$ signatures compared to the central complex ijolite series samples, indicating possible crustal contamination. The data from these samples are also similar to some of the most enriched data points on the KCL (Kramm, 1993), not including the solitary point from Vuoriyarvi which falls within the enriched quadrant. Data obtained more recently from the Vuoriyarvi carbonatites (Demaiffe et al., 1997) fall within the depleted quadrant.

In contrast to the relatively enriched isotopic signatures of this group of samples, their trace element spidergram patterns (Chapter 7) are relatively flat and unenriched. This decoupling between enriched isotopic signature and relatively depleted trace element signature of this group of samples, is the opposite of what was seen in the pyroxenite-melilitolite-nephelinite group (depleted isotopic signature, enriched trace element signature). This suggests that the “enriched group” magmas are either derived from a different source, were subjected to a different type of metasomatism, or both, compared to the “depleted group”. The obvious source location would be within the continental lithosphere, within the spinel stability field, and a likely candidate for a metasomatising agent would be a kimberlitic melt (Wyllie, 1989), given the similarity in isotopic composition to the Terskii Coast kimberlites. The trace element compositions of the kimberlite samples have not been analyzed in this study, but a general spidergram of the Terskii kimberlites and olivine melilitites (Beard et al., 1996) shows that the samples are not strongly enriched in trace elements, and have a similar degree of enrichment to the isotopically enriched group under discussion. This isotopically enriched group of samples are also relatively enriched in the alkali elements, consistent with derivation from a phlogopite or amphibole-bearing source. Kimberlite metasomatism to create such a source, possibly in the base of the lithosphere, is consistent with the model of Dautria et al. (1992). However, the trace element ratios of the olivine melteigite sample

[C.AG.48A], the most primitive sample within the “enriched group” also match those of the EM I mantle component, according to the data compilation provided by Weaver (1991). As discussed in the previous chapter, this provides confirmation of the identity of the enriched end-member forming the KCL, proposed by Kramm (1993), and also raises tantalizing questions regarding the relationship, if any, between the EM I mantle component and kimberlitic melts, given the trace element and isotopic similarities between the “enriched group” and the Terskii kimberlites described above, as suggested by Grundfelder et al. (1986). The EM I component was defined on the basis of isotopic signatures of oceanic basalts by Zindler & Hart (1986). Any link between oceanic EM I and kimberlites is difficult to constrain as the geochemical fate of kimberlites beneath the oceanic crust and their influence on other upwelling magmas is poorly constrained at the present time.

Mineralogical similarities among the members of the isotopically “enriched” group are also strong. All samples are dominated by pyroxene, garnet and mica as the main mafic phases, and nepheline as a common ground-mass constituent, other than the turjite samples which contain analcime or calcite as the main groundmass phase. Magnetite is rare in all samples except the olivine melteigite, and Fe^{3+} is normally contained within the silicate phases such as garnet, aegirine-augite or $^{\text{iv}}\text{Fe}^{3+}$ -phlogopite. In contrast to the melilitolites and pyroxenites, there is no $^{\text{vi}}\text{Al}$ present in the silicate phases. The apatite compositions are distinguished by enrichment in $\text{Sr} > \text{Si}$ (unlike the “depleted” group which have $\text{Si} > \text{Sr}$) while perovskite, which is rare in these samples, is normally relatively depleted in REE. Other variations and correlations have been discussed in Chapter 6.

The two groups of ijolites described by Bulakh & Ivanikov (1984) and Ronenson et al. (1978) can only be distinguished on the basis of modal mineralogy. Although it is hard to distinguish the two groups on the basis of mafic index, it is possible to distinguish

them on the basis of the ratios of garnet : mica : pyroxene contents. The first group, represented by samples C.AG.122, C.ND.500 and E.123.106, has ratios of 1 : 0 : 5 (i.e. pyroxene >> garnet). Samples C.90y and C.IJ.14 have ratios of 3 : 0 : 2 and 5 : 1 : 2 (i.e. garnet > pyroxene). Both of these groups contain little mica. In general, the samples with the most phlogopite also contain small amounts of magnetite. The modal mineralogy, along with observations from additional drill-core samples, suggest that some of the main central complex samples share a similar mineralogy or crystallization conditions with the eastern ijolite ring body (represented by sample E.123.106), in being enriched in pyroxene. The ijolite body surrounding the carbonatite complex contains samples enriched in garnet, which fits in with the idea of garnet-rich ijolites representing a reaction product between a pyroxene-rich ijolite and a carbonatite (Verwoerd, 1978). The model of Verwoerd (1978) was used to explain the origin of the turjite dykes, which appear to represent the end-products of this reaction. The other two ijolite samples analyzed are WKM, the west coast micro-ijolite which is not in textural equilibrium, and sample C.170.38 which contains 40% cancrinite. Both of these samples contain equivalent amounts of garnet + mica : pyroxene.

The turjite also belongs to this group of “enriched” rocks. Its $^{87}\text{Sr}/^{86}\text{Sr}$ isotopic signature is more enriched than those of the central massif ijolites and the olivine melteigite, which may be a result of crustal contamination. Its genesis, in terms of an ijolite-carbonatite reaction, has already been discussed in previous chapters. However, some models would suggest that ijolites or nephelinites and carbonatites are conjugate liquids (Kjarsgaard & Hamilton, 1989), and thus the reason why an ijolite would *react* with a carbonatite to form the garnet-rich turjite (and possibly garnet-rich ijolites, like sample C.IJ.14) is unclear. If this was a metasomatic reaction, a pre-existing ijolite might react with later carbonatite fluids and hence be “metasomatized”. However, the trace element signature of the turjite sample TU 119 does not substantiate this argument.

The extreme enrichment in HREE does not indicate LREE-rich carbonatite metasomatism, but more of garnet accumulation from a melt such as ijolite sample C.IJ.14, which, in turn, shows a fractionation relationship with the olivine melteigite [C.AG.48A] described earlier.

It is useful to make comparisons between the main characteristics of the mineralogy of the isotopically “enriched” and “depleted” groups. Table 8.A below summarizes the variation in mineral composition and abundance between the melilitolites and ijolites, of the minerals which show systematic variations in composition, particularly with respect to Al content and co-ordination, as outlined in Chapter 5.

Table 8.A Comparison of melilitolite and ijolite mineralogy

Mineral	Notable characteristic	Ijolite series rocks	Melilitolites
pyroxene	abundance	15-70%	<4%
	aegirine	aeg ij > mel.	aeg mel < ijolite
	Tschermak	Tsch. ij < mel.	Tsch. mel. > ijolite
garnet	abundance	5-25%	<4%
	^{vi} Al content	none present	none present
mica	abundance	0-12%	2-35%
	^{vi} Al content	none present	always present
	Al : Si & Mg #	ijolite < mel.	mel. > ijolite
apatite	Si : Sr	Si < Sr	Si > Sr
magnetite	abundance	0-7%	3-30%
	Al	ijolite < mel.	mel. > ijolite

[Note that “always” or “<, >” are used here to express approximately 90% confidence levels; mel. = melilitolite, and Tsch. = $\text{CaTiAl}_2\text{O}_6 + \text{Ca}(\text{Fe}^{3+}\text{Al})_2\text{SiO}_6$ components].

This table shows that there are some very distinct differences in the mineral compositions of minerals common to the two rock-types. These characteristics include low pyroxene abundance in the melilitolites, with low aegirine and high Tschermak

component contents. ^{vi}Al is never present in the garnet or mica compositions from the ijolites, and magnetite is a ubiquitous and abundant constituent of the melilitolites.

Fractionation trends in some silicate rocks can be monitored by the Mg#. The relationship between the Mg# of the bulk rock chemistry of the silicate samples and the Mg#s of the minerals within is shown in Figure 8.6 for a selected group of samples, where Mg# of the minerals is defined as $100\text{Mg}/(\text{Mg}+\text{Fe}^{2+})$, while the Mg# of the bulk rocks is defined as $100\text{Mg}/(\text{Mg}+\text{Fe}_{\text{TOTAL}})$, where $\text{Fe}_{\text{TOTAL}} = \text{Fe}^{2+}$. This figure shows that, for each sample, the variation between the bulk rock Mg# and the mineral Mg#s can be substantial, and emphasizes the need for caution when interpreting the Mg# of bulk rock analyses in Si-depleted rocks. The partition coefficients needed to produce the mineral Mg#s from the bulk rock values in Figure 8.6 are unrealistic, due to the presence of considerable quantities of Fe^{3+} in the minerals and hence the bulk rock compositions which does not readily substitute for Mg in the ferro-magnesian minerals. This implies that Mg# of the bulk rocks, which is based on $\text{Fe}_{\text{TOTAL}} = \text{Fe}^{2+}$ should not be used as a strict indicator of “primary” or “differentiated” magmas. Thus the Mg# of the bulk rocks in Figure 8.6 is a minimum estimate of the true $100\text{Mg}/(\text{Mg}+\text{Fe}^{2+})$ in the samples.

8.6c Phoscorites and carbonatites

The origin of the carbonatites and phoscorites, and their relationship with the silicate rocks in Turiy remains uncertain. The phoscorites have the most depleted isotopic signatures of all samples investigated from Turiy, and their isotopic signatures show no overlap with any of the other rocks in the massif. They are closest in Sr-Nd space to the central massif carbonatites. The trace element signatures of the phoscorites have not been determined, although sample C.49.40, whose bulk rock and trace element characteristics were determined and described in Chapter 8, is close to a phoscorite in modal composition although not in texture.

Figure 8.6 Relationship between the Mg# (x-axis) of silicate whole-rock samples and their silicate minerals. Bulk rock Mg#s from Table 7.A and mineral compositions from Appendix J. The thickness of each of the mineral lines within the boxes corresponds to the relative abundance of each mineral within the sample. The length of each line corresponds to the range in Mg# of that particular mineral. It can be seen that the Mg#s of the minerals are often commonly substantially different from those of their host rocks, due to the influence of Fe^{3+} which does not substitute for Mg. This diagram emphasises the need to treat the significance of bulk-rock Mg#s where Fe_{TOTAL} is assumed to be Fe^{2+} , with caution in Si-undersaturated rocks

The texture of the phoscorites examined in this study is distinctive. They contain coarse-grained segregations of equigranular crystals of magnetite, phlogopite and apatite. Apatite commonly occurs as veins in these samples and also forms an interlocking texture with magnetite. Calcite, and occasionally dolomite, form most of the interstitial material. Forsterite is the dominant silicate mineral in the phoscorites from the Kovdor massif, but forsterite has not been found in the phoscorites from Turiy that were examined in this study. The textures described above from the Turiy phoscorites are not indicative of cumulus settling. The equigranular crystal segregations and apatite veining may be indicative of flow differentiation or immiscibility between silicate - [P-Fe] - carbonate melts, given the clusters of phlogopite - magnetite-apatite - calcite crystals and veins found within the phoscorites.

The modal mineralogy tables in Appendix Biii are, at times, inconsistent with the true mineralogy of the phoscorites, since the modal analyses are taken from thin section mounts which are not necessarily representative of the overall modal mineralogy, given the heterogeneity of the specimens.

Studies of melt structure summarized in section 8.6 suggest that Fe^{3+} , Ti and P will form discrete units within silicate melts, preferring, on the whole, not to substitute for Si in silicate or aluminosilicate structural units. The presence of discrete segregations of Fe-Ti and P-rich structural units in a melt with low SiO_2 -content where separation of such units from a co-existing silicate melt will be relatively easy, is thus not surprising. Whether the separation of these discrete melt units occur as a result of immiscibility or crystal fractionation is harder to judge, and the composition of the parent melt or conjugate silicate liquid is also uncertain. The isotopic signatures, as well as major element chemistry of the silicate rocks, suggest that if the phoscorites are derived by magmatic differentiation, the melilitolites are the most likely candidates to reflect the parental magmas. Fractionation of silicate phases within a melilitolite magma will cause

the residual liquids to become depleted in SiO_2 . Only the precipitation of non-silicate phases such as magnetite and apatite can prevent the melilitolite bulk-rock compositions from migrating below the critical 33% Si boundary discussed earlier. The fact that melilitolites contain substantial magnetite $\pm$ apatite suggests that they are possible candidates for parental magmas, if the phoscorites are produced by fractional crystallization, but the melilitolite phlogopite compositions, in particular, are rich in ^{vi}Al , and are vastly different to those of the phoscorite micas. Pure liquid immiscibility between the two magma types is therefore not a viable option, unless a Fe^{3+} -P-Ca-Si-Mg-K liquid separates out from the melilitolites and subsequently fractionates and crystallizes the phases seen in the phoscorites. Isotopically, it can be seen from Figure 6.4 that the phoscorites could, in theory, be the most isotopically depleted end-member of the melilitolite trend.

An alternative possibility is that the phoscoritic magmas are somehow related to the ijolite series rocks which surround the carbonatites and phoscorites within the central complex, but the isotopic signatures appear to rule this out.

The most likely explanation, however, is that the phoscorites are derived from a depleted mantle source and may be linked to the origin of the melilitolites (and/or pyroxenites). They are the most depleted rocks, isotopically, seen in Turiy, and may have been intruded after a period of quiescence following the emplacement of the silicate magmas at Turiy.

There are three suites of carbonatites seen at Turiy, based on field relationships, and isotopic, trace element, and mineralogical variations which are summarized below. The isotopic signatures of the suites of carbonatites, as well as the phoscorite isotopic data, are plotted in Figure 8.7

Sample S.W58 is representative of a suite of fourteen carbonatite dykes located on the south coast of the peninsula, and is fine-grained and slightly banded. Isotopically it

Figure 8.7 $\epsilon_{\text{Sr}}(\text{T})$ vs $\epsilon_{\text{Nd}}(\text{T})$ data for the Turiy carbonatites and phoscorites, and the Terskii Coast kimberlites. Data from Tables 7.(A, B, C). The data fall into distinct groups. The data from the central massif samples falls below the KCL, while data points from the south coast dyke (S.W58) and northern carbonatites falls above the KCL. Sample C.54.x represents a late-stage vein within the central massif. "Terskii Kimbs." = Terskii Coast kimberlite whole rock and mineral separate data.

plots within the depleted quadrant in Figure 8.7, with a similar $\epsilon_{Nd}(T)$ value to those of the phoscorites, but with a more enriched $\epsilon_{Sr}(T)$ value, assuming a Devonian age. It lies to the right of the KCL and above the melilitolite data. Its trace element pattern is almost identical to that of the “average calcite carbonatite” of Woolley & Kempe (1989).

The carbonatites from the northern complex include melilite-bearing carbonatites which grade into calcite-rich melilitolites. Data from these samples plot to the right of the KCL in Figure 8.7, within the top right-hand quadrant. They are similar to the phoscorites in terms of $\epsilon_{Nd}(T)$ values, but are much more enriched in $\epsilon_{Sr}(T)$ values. The one exception is sample N.63.4 which is a pyroxene-calcite rock whose origin, as discussed in Chapter 6, is believed to relate to the reaction: melilite + CO_2 = diopside + calcite. The mineral composition of the pyroxene in this sample is very similar to that of the melilitolite-pyroxenite suite in Figure 5.5e, and its isotopic composition overlaps that of sample N.64.19, a melilite and monticellite-bearing calcite carbonatite. The stable isotope values of the northern complex carbonatites are enriched in $\delta^{18}O_{SMOW}$ compared to those of the carbonatites from the central complex (Table 6.E). The low $\delta^{18}O_{SMOW}$ and $\epsilon_{Sr}(T)$ values of the central complex carbonatites and phoscorites do not form a mixing pattern with the higher $\epsilon_{Sr}(T)$ values and $\delta^{18}O_{SMOW}$ of the northern carbonatites. The trace element compositions of the northern carbonatites have not been analyzed.

There are no dolomite-bearing carbonatites in the northern complex, and, according to Kapustin (1980), the modal and mineralogical compositions of the northern carbonatites would be considered representative of “early carbonatites”. Sample N.64.19 is dominated by calcite, melilite, magnetite and ^{vi}Al -bearing mica. Samples N.187.100 and N.187.134 contain ^{vi}Al -bearing garnet and mica along with pyroxene and abundant calcite. Magnetite is only found in the melilite-bearing samples. Sample N.187.100 contains a vein of microcrystalline quartz of unknown origin, and unusual rounded phenocrysts of calcite exhibiting strained extinction patterns, in a fine-grained,

interlocking calcite matrix (Figure 4.10d). Calcite deforms easily, and the deformation seen in the phenocrysts presumably occurred during the eruption. This deformation is also shared, to a lesser extent, by some of the coarse-grained carbonate phases in the central complex carbonatites and may also relate to tectonic forces operating as a result of the uplift which has occurred since the Devonian.

The occurrence of ^{vi}Al -bearing silicate phases in these carbonatites, along with the occurrence of melilite, suggests that these carbonatites are associated with the nearby melilitolites and, like the melilitolites from the satellite complexes, have an isotopic signature with depleted $\epsilon_{\text{Nd}}(\text{T})$ values but enriched $\epsilon_{\text{Sr}}(\text{T})$ values. These carbonatites may represent early fractionation products from the melilitolites, according to the ternary phase diagram of Nielsen (1980), based on the experimental results of Watkinson & Wylie (1971). It was used by Nielsen (1980) to explain the production of carbonatites in the melilitolite-bearing Gardiner Complex in Greenland. The lack of magnetite in samples N.187.100 and N.187.134, along with the presence of garnet, pyroxene and mica is similar to the mineralogy of many of the ijolite series rocks. However, the micas in these carbonatites are ^{vi}Al -bearing, the pyroxenes have high Mg#s and low aegirine content, and the garnet compositions also contain ^{vi}Al and very low concentrations of Ti. These are features all held in common with the mineral phases present in the melilitolites. The ijolite micas are $^{iv}\text{Fe}^{3+}$ -bearing, their pyroxenes are relatively rich in aegirine and have moderate Mg#s, and their garnet compositions contain no ^{vi}Al and are rich in Ti. So although the modal mineralogy of the non-carbonate phases in these carbonatites resembles the mafic mineralogy of the ijolite series rocks, the mineral compositions are quite different.

The third, largest, and most varied suite of carbonatites, are represented by carbonatites from the central complex. They have lower $\epsilon_{\text{Nd}}(\text{T})$ and higher $\epsilon_{\text{Sr}}(\text{T})$ values (Figure 8.7) than the phoscorites, although the latter enrichment is more moderate than

the former depletion. They are the only carbonatites in Turiy to lie below and to the left of the KCL. Their $\delta^{18}\text{O}_{\text{SMOW}}$ stable isotope signatures fall within the “primary igneous carbonatites” field of Hoefs (1987), but their $\delta^{13}\text{C}_{\text{PDB}}$ values are higher, as discussed in Chapter 7. It was noted in Chapter 7 that the stable isotope signatures of the central complex carbonatites (and phoscorites) were very similar to some carbonatites from the Kovdor, Fen and Alnö complexes, as well as carbonatite dykes near Kandalaksha, and may suggest a similar mantle source for the carbon.

Three of the central complex carbonatites from Turiy (C.49.40, C.TL.344 and C.DC) were analyzed for their trace element content. The first two samples are borderline carbonatites, in that they have low carbonate contents ($> 40\%$), but are texturally homogeneous and have a varied mineralogy. The third sample is a dolomite-calcite carbonatite which contains no silicate phases, and accessory apatite and sulphide minerals. The trace element signatures of these three samples are remarkably similar to each other, but quite different to that of the carbonatite dyke sample from the south coast [S.W58] (Figure 8.7). The three central complex carbonatites analyzed have low degrees of trace element enrichment in comparison with sample S.W58 and silicate whole rocks from the massif. This feature is also seen in Kovdor (Demaiffe et al., 1997) and is unusual, compared with most carbonatites (Woolley & Kempe, 1989). Specific details of the spidergram patterns were outlined in Chapter 7 (section 7.4e). The reasons for this lack of trace element enrichment are unclear, and may, in part, be related to inadequate sampling of the central complex carbonatites. However, the Sr and Nd concentrations in calcite separates from a wide range of carbonatites, as determined by isotope dilution (Tables 6.A and 6.B), suggest that many of the central complex carbonatites have similar Sr and Nd enrichment characteristics, assuming that the partition coefficients are constant throughout. The unusual lack of trace element enrichment in the spidergram patterns has to be incorporated into any overall model devised to explain their petrogenesis.

Mineralogically, the central complex carbonatites contain some interesting variations. Figure 8.8 is a ternary plot summarizing the modal mineralogy of the Turiy phoscorites and carbonatites (excluding the south coast carbonatite dyke S.W58). It can be seen that the northern carbonatites are rich in silicate phases and poor in non-silicate, non-carbonate phases such as magnetite and apatite. The central carbonatites appear to fall into two groups within this diagram. One group is carbonate-rich and forms a cluster of points near the carbonate apex. This group contains < 10% silicate phases, and < 30% non-silicate, non-carbonate phases. It includes the two dolomite-rich samples C.DC and C.100a. An apparent compositional gap, which may be due to inadequate sampling, gives way to a band of points which are located between 40-58% carbonate. If sample C.54.x is ignored, the group lying between 40-58% carbonate contains < 30% silicate phases but is rich in non-silicate, non-carbonate phases (17-50%). This group includes two of the central complex carbonatite samples which were analyzed for major and trace elements (C.49.40 and C.TL.344). The third sample to be analyzed for trace elements was sample C.DC which falls into the carbonate-rich group. The phoscorites have also been plotted in this diagram, along with the calcite-melilitolite sample N.64.18m. The major element data from samples C.49.40 and C.TL.344, coupled with estimates of major element compositions based on mineral compositions and modal mineralogy from other samples, suggests that none of the central massif carbonatites contain > 10% SiO₂. In contrast, the northern massif carbonatite samples are estimated to contain approximately 11-25% SiO₂ which is considered high for true carbonatites (Woolley & Kempe, 1989), although textural heterogeneities in these samples suggest that these estimates should be treated with some caution.

A link between the Mg# of the carbonatite micas and the modal mineralogy of the carbonatites can also be demonstrated by correlating the data shown in Figures 8.8 and 8.9. The latter diagram shows that the mica data can be divided into three groups. The

first group, a small cluster of mica analyses from sample C.49.40 have Mg# 90-95 and $^{iv}\text{Fe}^{3+}$ contents of ~ 1.3 apfu. The rest of the carbonatite mica analyses have $^{iv}\text{Fe}^{3+}$ contents between 0-0.5 apfu, and can be further divided into two groups with Mg# > 90 and Mg# < 90 respectively. Almost all the mica analyses with Mg# < 90 come from samples with < 20% “100-silicate-carbonate” phases. The “100-silicate-carbonate” apex of Figure 8.8 is essentially correlated with the magnetite and apatite content of the samples. Thus, there appears to be a link between low magnetite and apatite contents and low Mg# of the micas from the same samples. This is presumably due to low $f\text{O}_2$ of the melt, preventing the crystallization of magnetite, which implies a relatively low $\text{Fe}^{3+}/\text{Fe}^{2+}$ ratio, and hence lower Mg# of the melt and ferro-magnesian minerals crystallizing from it. There are two points to note here. Firstly, the samples which fall into this group are the later, carbonate-rich carbonatites, and secondly, this implies that there is a definite link between modal mineralogy and mineral composition within these samples (mica Mg# is inversely proportional to magnetite + apatite content). This in turn suggests that the mineral phases in these samples crystallized in equilibrium with each other, and that these samples originally represented actual melt compositions, rather than a mixture of cumulus phases and spatially unrelated transporting liquids.

Given the lack of any evidence for silicate-carbonate immiscibility to produce the the central carbonatite samples, other options to explain the production of the central complex carbonatite $\pm$ phoscorite magmas should be examined. Since the isotopic compositions of the silicate magmas at Turiy are different from the central complex carbonatites, a direct fractional crystallization model cannot be invoked without open-system behaviour, such as mixing of a liquid fractionated from a silicate parent, with new carbonate material derived from a mantle source. Alternatively, given the isolated natures of their isotopic signatures and unusual trace element patterns, the central complex carbonatites may be primary, derived directly from the mantle. Unfortunately,

Figure 8.8 Ternary diagram based on the modal mineralogy of the carbonatites and phoscorites. Data from Appendix Biii. Groupings the same as in Figure 8.7, with the addition of two melilitolite points (N.64.18 & TUR I). The minerals representing the [100-silicate-carbonate] apex are dominated by apatite and magnetite.

Figure 8.9 Mica compositions for the carbonatites. The central complex carbonatite analyses with high $^{iv}Fe^{3+}$ are from sample C.49.40. The analyses with low $Mg\#$ are from the samples containing < 20% [100-silicate-carbonate] component in Figure 8.8 (i.e. low in apatite and magnetite). Sample C.54.x represents a late-stage carbonatite vein within the central massif.

there are no realistic criteria yet established for the recognition of primary carbonatite magmas.

Two models devised to explain the location of the central complex phoscorite and carbonatite data on the Sr-Nd anti-correlation plot are illustrated in Figure 8.10. The first model assumes three component mixing between magmas from a depleted asthenospheric reservoir, an ancient carbonatite trapped within the lithosphere, and an enriched mantle component, possibly residing within the spinel stability field. An episode of magmatic activity in the area during the Precambrian, supported by the Rb-Sr isochron from pyroxenite sample C.AG.347, is assumed to have included the formation of a carbonatitic melt, initially derived from the asthenosphere (reservoir A in Figure 8.10), which was frozen within the lithosphere, probably trapped beneath the mantle solidus ledge at point Q as illustrated in Figure 8.3. This carbonatite is assumed to have evolved isotopically under closed-system conditions within the lithosphere, with $Rb/Sr \sim 0$ and $Sm/Nd = 0.15$ (Bell & Blenkinsop, 1987a,b), as shown by X in Figures 8.10 a&b. During the Devonian, fresh melts were extracted directly from the same asthenospheric reservoir (A), intruded into the crust, and lie at the depleted end of the KCL. Mixing of magmas with similar Sr/Nd ratios derived from (A) and EM I (the latter possibly lying within the spinel stability field at the base of the lithosphere) created the Devonian Kola Carbonatite Line (Kramm, 1993). However, towards the end of Devonian activity, the final melts are likely to have become volatile-rich, cooler, and represented smaller percentages of partial melt, as discussed in section 8.4. It is therefore possible that these melts were not able to reach the surface (or crustal magma chambers) directly but encountered the CO_2 -rich mantle solidus ledge shown in Figure 8.3. If the part of the lithosphere which contained this "ledge" were already saturated in volatiles from a previous (Precambrian) magmatic event, as outlined above, further incursion of carbonatite melt from the asthenosphere would cause over-saturation of this part of the lithosphere in carbonatite melts and

Figures 8.10a & b Sketch of the evolution of a carbonatite (X) extracted from a depleted asthenospheric reservoir (A) at t_1 and frozen within the continental lithosphere. The carbonatite is assumed to have a Rb/Sr ratio ~ 0 and Sm/Nd $= 0.15$ (Bell & Blenkinsop, 1987), both of which are lower than the values assumed for A and for bulk Earth – CHUR (Faure, 1986). At t_2 , direct extraction of a second depleted melt from reservoir A is accompanied by remelting of the frozen carbonatite (X) derived from A at t_1 . The epsilon values are represented by the vertical distance between the A or X reservoirs and bulk Earth-CHUR. At t_1 , the epsilon values for both A and X are identical. At t_2 , $\epsilon_X \text{ Sr} > \epsilon_A \text{ Sr}$ (negative values), and $\epsilon_X \text{ Nd} < \epsilon_A \text{ Nd}$ (positive values). The extraction of melts such as X from a large asthenospheric reservoir can, in theory, occur during repeated phases of magmatism. Thus, another carbonatite melt X_2 could be extracted from A at t_2 and evolve in a similar fashion to that of X from t_1 to t_2 . This is essentially a step-wise process. The model assumes that the carbonatite extracted from the asthenospheric reservoir is frozen under closed-system conditions within the lithosphere, probably trapped under the solidus ledge at Q in Figure 8.3. The implications for Turiy, where t_1 is assumed to be ~ 588 Ma and $t_2 \sim 373$ Ma, are outlined in Figure 8.10c.

Figure 8.10c A schematic Sr-Nd diagram to illustrate the results of three-way mixing between end-members A, X and EM I at 373 Ma, in order to produce the trend, marked by an arrow, of the central complex phoscorites and carbonatites, as shown in Figure 8.7. Point A represents a carbonatite melt derived from an asthenospheric reservoir at 373 Ma, point X represents a carbonatite derived from the same asthenospheric reservoir during the Precambrian, which was trapped under closed-system conditions in the lithosphere (probably underneath the solidus ledge at point Q, shown in Figure 8.3) and remelted at 373 Ma, and EM I represents the enriched component defining the end of the Kola Carbonatite Line (KCL) (Kramm, 1993). The creation of point X is explained in Figures 8.10a & b above.

Figure 8.10d An alternative model to produce the central complex carbonatite and phoscorite data lying below the KCL, as shown in Figure 8.7, involves two-component mixing between a carbonatite with high Sr/Nd formed from a parental magma derived from an asthenospheric reservoir lying at the top of the KCL represented by A, with either a melt with trace element and isotopic characteristics similar to the olivine melteigite “OM” [C.AG.48A] or one of the Terksii kimberlites “TK” [K77A]. The A-OM and A-TK mixing lines are shown on the diagram. See text for further explanation.

metasomatic fluids, causing remelting of the Precambrian trapped carbonatite (X) and mixing with the new Devonian melts (A), which could escape to the surface through crack propagation (Wyllie, 1989). At some point in this sequence of events, the third component, assumed to be EM I, also becomes involved in the mixing process. The direction of the “arrow” in the mixing triangle in Figure 8.10c suggests that the carbonatites from reservoir A are mixing with a X-EM I combination, and hence it seems likely that EM I is derived from the lithospheric wall-rocks that are host to the trapped, Precambrian carbonatite X and, presumably, the “Q” mantle solidus ledge. Numerical modelling involving storage of a carbonatite magma with $Rb/Sr \sim 0$ and $Sm/Nd = 0.15$ for ~ 200 Ma within the lithosphere, has shown that this model is a feasible option to justify the locations of (X) and the mixing arrow in Figure 8.10c, although it must be remembered that the A-X-EM I triangle drawn in Figure 8.10c implies that all three components must have the same Sr/Nd ratios.

The second model proposed to explain the isotopic signatures of the central complex phoscorites and carbonatites which lie below the KCL is outlined in Figure 8.10d. It involves two-component mixing between a carbonatite derived from the asthenospheric reservoir (A), and a melt with an isotopic signature and trace element concentrations similar to either the olivine melteigite (OM) [sample C.AG.48A] or Terskii kimberlite (TK) [sample K77A]. The similarities between the isotopic and trace element signatures of the olivine melteigite [C.AG.48A] and the Terskii kimberlites has previously been discussed, but the nature of the relationship between the Terskii kimberlites and the enriched end-member of the KCL remains unresolved. Mixing lines calculated for A-OM and A-TK are shown in Figure 8.10d. The enriched end-member component of the KCL was assumed by Kramm (1993) to be EM I. Certainly, the trace element signature of the olivine melteigite sample [C.AG.48A] would seem to affirm this identity, but whether the true location of EM I during the Devonian lies closer to bulk

Earth - CHUR than suggested by Kramm (1993) and as drawn in Figure 8.10d, is not known. Regardless of the true “identity” or precise location of this enriched component along the KCL, the fact that most of the silicate magmas at Turiy lie along or close to the KCL, means that the Sr/Nd ratios of the two KCL end-member reservoirs must be the same. To create the curved mixing lines seen in Figure 8.10d necessitates the derivation of a carbonatite magma from reservoir A with a much higher Sr/Nd than the original reservoir signature, and as such, it is possible that this carbonatite may be derived from a silicate parent magma. It is this carbonatite that must interact and mix with a melt derived from the enriched reservoir to form the concave mixing lines shown in Figure 8.10d. It can be seen that the parts of the mixing lines closest to (A) in Figure 8.10d are in the same location as the arrow shown in Figure 8.10c, which indicates the approximate location of the central complex carbonatite and phoscorite data. The most likely place for the “A-OM” or “A-TK” interaction and mixing to take place is within the lithosphere, in a similar fashion to that suggested for the three-component mixing model outlined previously. The main difference here is that the “ancient carbonatite” component (X) is eliminated in the two-component model, following the assumption made by Wilson et al. (1995) in similar modelling to explain the isotopic signatures of the Rhine graben melilitites, that the lithosphere is emptied of volatile-enriched melts at the end of each phase of magmatism, although porosity studies suggest that this is not entirely realistic (Minarik & Watson, 1995).

Sample C.54.x, a late-stage carbonate-zeolite (natrolite) vein from the north of the central complex containing apatite, garnet and mica, and described in detail in Chapter 4, does not fit into any of the three groups of carbonatites previously described above. It is the only sample to show strong concentric zoning within calcite (Figure 4.10c). This sample shows mineral chemistry characteristics similar to those of the northern complex carbonatites with ^{vi}Al-bearing garnet and mica compositions. However, the one isotopic

analysis carried out on a mica separate from this sample places it in with the cluster of ijolite data around the bulk earth - CHUR intersection (Figures 6.6 & 8.7). Thus, its origin and relationship to the other magmas in the complex is somewhat uncertain. No trace element data is available from this sample.

An evolutionary model for the carbonatites, based on the major element chemistry of the Turiy carbonatites along with the chemical and textural characteristics of their constituent minerals, can now be outlined. The most SiO_2 -rich carbonatites are those from the northern complex (Figure 8.8). Kapustin (1980) has suggested, on the basis of modal mineralogy, that these are early carbonatites. Several of these samples contain $^{\text{vi}}\text{Al}$ -bearing garnet and mica with abundant pyroxene. Melilite and monticellite may also be present but apatite and mica are rare. Thus, the melt appears to have contained silicate phases representing a range of structural units ($\text{NBO/T} = 1$ [mica]; $\text{NBO/T} = 2$ [pyroxene,]; $\text{NBO/T} = 3$ [melilite], and $\text{NBO/T} = 4$ [garnet and monticellite], where NBO/T = non-bridging oxygens per tetrahedra), along with calcite. The minerals precipitating out of the melt have high Al/Si ratios, moderately high Mg#s, and the melt is saturated in carbonate producing calcite but no dolomite. The $f\text{O}_2$ is too low and/or Fe contents too low to produce abundant magnetite. The carbonatites present in the northern complex can also be divided into two types - those with calcite + melilite + magnetite $\pm$ monticellite $\pm$ garnet $\pm$ perovskite [N.64.19, N.64.18], and those with calcite + mica + garnet + pyroxene $\pm$ quartz [N.187.100, N.187.134].

The central complex carbonatites appear to be more evolved, according to the mineralogical model of Kapustin (1980). Ignoring the carbonatite-zeolite vein sample C.54.x, none of the central complex carbonatites contain garnet, melilite or monticellite, and pyroxene is rare. Mica compositions are all $^{\text{iv}}\text{Fe}^{3+}$ -bearing. Magnetite becomes a common component along with apatite and mica, and amphibole and dolomite are seen for the first time in the carbonatites. Thus, the four essential minerals in most of the

central complex carbonatites are calcite, apatite, magnetite and $^{iv}\text{Fe}^{3+}$ -phlogopite. Al reaches minimal concentrations in the silicate minerals and in magnetite, and the bulk-rock Si contents are also lower than in the northern carbonatites. The earliest carbonatites in the central complex, according to field evidence (Bulakh & Ivanikov, 1984), appear to contain rare pyroxene along with magnetite and apatite (e.g. sample C.TL.344). Mica is rare and the melts appear to be saturated in Fe and P. Later carbonatites contain mica instead of pyroxene, and as tetrahedrally co-ordinated Fe^{3+} becomes abundant in the micas, the Mg# increases to reach the mid-90s. The $f\text{O}_2$ of the melts appears to be high at this point, as Fe^{2+} is used up in the production of magnetite and little is available for entry into the silicate phases. The preferential combination of CO_3^{2-} anions with Ca^{2+} leaves the micas saturated in Mg, along with K and Si. It should be noted that Na cannot partition into any of the four dominant minerals phases at this time, and that the pyroxenes in the earlier carbonatites also have low Na contents. Whether the melts were already depleted in Na at this time, or whether Na remained in solution, is unknown. The saturation of the melt in Mg, and exhaustion of available Si and K, causes the subsequent precipitation of dolomite while magnetite is still precipitating. Thus, the mineral assemblage changes from calcite + pyroxene + magnetite + apatite $\pm$ amphibole $\pm$ mica, to calcite + dolomite + magnetite + apatite + mica (e.g. sample C.49.40 and sample C.6.15). The timing of saturation in different components varies in different samples. The later, carbonate-rich samples such as the C.100 series and sample C.DC become increasingly dolomite-rich and are increasingly poor in silicate minerals, although magnetite and apatite contents can remain high in the C.100 series. Samples such as C.DC are the most evolved and contain equal amounts of calcite and dolomite, along with small quantities of apatite and sulphides, and no silicate phases or magnetite.

Amphibole is listed as a typical mineral in carbonatites (Hogarth, 1989). However, in the Turiy carbonatites, amphibole is not a common component. Richterite is

found in a few samples but is thought to be secondary, or late-stage hydrothermal in origin (Bulakh & Ivanikov, 1984).

The lack of any carbonatites in the central complex similar in composition to the northern complex carbonatites, may merely reflect inadequate sampling. However, careful examination of drill core material suggests that this is probably not the case. The northern carbonatites are good candidates, in terms of isotopic composition and mineralogy, for derivation from the same source as the melilitolites on the basis of field, mineralogical and isotopic evidence. Additionally, the high content of silicate material in the northern carbonatites suggests that they might be carbonatite fractionation products of the melilitolite magmas. If these carbonatites do represent fractionation products or residual melts from the melilitolites, then the high garnet and pyroxene contents of some of these carbonatites also provide excellent sites for the Zr-Hf-HREE trace elements, which are characteristically depleted in the melilitolites.

8.7 Summary

From the above discussion, it can be seen that there are many unresolved questions regarding the petrogenesis of the Turiy Massif. However, a brief, overall model will be presented here in an initial attempt to summarize the results derived from the data presented in this study and to provide the basis for further work.

Figure 8.11 outlines the model, where source "A" represents the depleted end-member of the KCL, located in the dolomite-phlogopite-garnet stability field within the asthenosphere. "X" represents ancient lithospheric carbonatite, originally derived from the asthenosphere, but subsequently trapped within the lithosphere and remelted during a later magmatic event. EM I is the enriched end-member of the KCL (Kramm, 1993) located in the spinel stability field within the lithosphere. The sequence of magma intrusion within Turiy suggests that melting was initiated in source "A" (pyroxenites,

melilites, melilitolites), and alternated with melting in the EM I source (olivine melteigite, ijolites), as noted in Chapter 6. Crustal contamination, probably dominated by selective dissolution processes, occurred during the formation of the satellite complexes. Magma mixing between melts derived from source “A” and EM I provided the spread of isotopic signatures observed along the KCL. Almost all the partial melts derived from the mantle represented volatile-enriched, low degrees of partial melting of the source rocks. The northern complex carbonatites were derived by fractionation from the spatially related melilitolites, while the central complex carbonatites (and phoscorites) were derived from a mixture of components from the three mantle sources (A, X, EM I), although the importance of X in the Turiy carbonatites is still a matter for debate. The relationship of the central complex carbonatites and phoscorites to the silicate magmas also remains uncertain, although the possibility that they represent fractionated primary carbonatites cannot be ruled out at this time.

Cumulate silicate bodies doubtless reside beneath the current level of the intrusion, but their composition is unknown. Olivine, pyroxene and possibly garnet and phlogopite may be constituents of such cumulate bodies. However, the Turiy intrusion could not have resulted from fractionation of the contents of a single magma chamber. Any cumulates that were formed probably reside at the base of the crust, or lining the walls of the magma conduits.

Open-system crystal fractionation is believed to have played key roles in the evolution of the Turiy magmas. The relationship between the ijolites and melilitolites hinges on many factors including fO_2 , volatile content and composition, and a_{SiO_2} . The melt structure reaction: $2Si_2O_6^{4-} = Si_2O_7^{6-} + Si_2O_5^{2-}$ (2 pyroxene = melilite + mica), exemplifies the changes in mineralogy between the two rock types. Correlation of the petrology of the Turiy samples with the results of melt structure studies suggest that separate phosphate, Fe-rich, and possibly sulphide liquids may have formed, and the

Figure 8.11 Sketch (not to scale) outlining a petrogenetic model for the Turii massif. The end-members are defined as follows: Source A = depleted end-member of the KCL, located within the garnet-dolomite-phlogopite stability field in the asthenosphere; EM I = enriched end-member of the KCL (Kramm, 1993) located within the spinel stability field in the lithosphere; Source X = ancient carbonatite, originally derived from the asthenosphere but now trapped within the lithosphere, probably at the level of the ledge in the mantle solidus, and derived according to the three-component mixing model outlined in Figures 8.10 a, b and c. If the two-component mixing model shown in Figure 8.10d is used, source X is not required. The lithosphere-asthenosphere boundary is assumed to be close to the spinel-garnet transition zone.

Magmas from (A) may ascend directly from the asthenosphere to the crust if they are hot enough, otherwise they may stop at the base of the TBL prior to reheating and eruption. Melts from the base of the TBL may ascend directly to the crust if they are hot enough. EM I melts may be derived from anywhere in the spinel stability field, possibly including the base of the TBL, and it is assumed that EM I pervades throughout much of the mantle lithosphere. The base of the TBL may also be a good location for A – EM I mixing to produce the magmas lying along the KCL. If the degree of partial melting becomes extremely small, the temperature of the resultant magma relatively low, and the peridotite solidus relatively saturated in volatiles, melts which are derived from the asthenosphere or TBL may become trapped beneath the solidus ledge within the MBL as shown in the sketch to the right. The volatiles, including carbonatites derived from (A), may escape to the surface, and any ancient volatiles trapped within the lithosphere (X) may also be remelted and erupted at this time, possibly mixing with melts from the wall-rocks (EM I).

All five complexes in Turii are believed to share a central plumbing system but crustal contamination occurred during formation of the satellite complexes.

presence of immiscible carbonate is also possible. The central complex ijolites and the turjite are the products of an pyroxene-ijolite - carbonatite reaction process.

CHAPTER 9

CONCLUSIONS

A summary of the main conclusions that can be drawn from the results obtained in this study are listed in point form below. The conclusions are arranged according to a sequential examination of the ages, sources and specific characteristics of the samples examined.

1. The earliest rocks in the Turiy massif are Precambrian in age. Pyroxenite sample C.AG.347 has a Rb-Sr age of 588 ± 72 Ma. Field and isotopic evidence indicate that the majority of the silicate rocks within the complex are Devonian in age. An ijolite from the central complex provided an Rb-Sr age of 373.1 ± 5.6 Ma (Kramm et al., 1993), which is supported by a Rb-Sr age of 363 ± 3.5 Ma from phoscorite sample C.K.7. Thus, tectonic and isotopic evidence suggest that the magmatic activity in the Kandalaksha Graben has undergone at least one phase of major reactivation. The ijolite and phoscorite ages also suggest that there may have been more than one period of magmatic activity at Turiy during the Devonian.

2. The Turiy Massif has evolved under open-system conditions, both at source, during magma intrusion, and during crystallization. The spread of data on the ϵ Sr - ϵ Nd anticorrelation plot is outside the limits of analytical uncertainties and indicates that more than one source has contributed to the Turiy magmatism. The clusters of some data points on the anti-correlation plot, corresponding to distinct rock types, suggest that each magma batch has been produced independently and that closed-system fractional crystallization of a single batch of magma cannot be responsible for producing the wide variation in rock types seen within the massif. Petrographic features within many of the silicate and carbonate samples include banding, crystal accumulation and vein intrusion,

which are all indicative of open-system crystallization.

3. Disequilibrium features are seen in the petrographic textures, as well as in the isotopic, and mineralogical data. Disequilibrium textures include garnet rims around melilite grains, perovskite and garnet rims around magnetite grains, and baddelyite overgrowths within a single apatite vein from a banded carbonatite. Cathodoluminescence examination of phases that include carbonates, apatite, and fluorite show extensive concentric and mottled zonation in several carbonatite and silicate samples. SEM photographs also demonstrate chemical zonation in many of the silicate and non-silicate mineral phases. Within-sample isotopic disequilibrium, outside the bounds of experimental error, has not yet been found, although the lack of Sm-Nd isochrons for samples with well-defined Rb-Sr isochrons (e.g. pyroxenite sample C.AG.347) may indicate isotopic disequilibrium during crystallization.

4. At least two mantle sources are needed to explain the isotopic data from the phoscorites, carbonatites, and ijolite series rocks. These two sources have depleted and enriched Sr-Nd isotopic signatures respectively, and fall approximately along the Kola Carbonatite Line, based on data from several Devonian complexes from the Kola Peninsula (Kramm, 1993). The isotopic data from samples taken from rocks from the outer complexes and three coastal dykes suggest that these magmas were subjected to crustal contamination, prior to, or during magma emplacement.

5. The stable isotope data from calcite separates from the Turiy carbonatites and phoscorites indicate that these rocks are magmatic in origin, and have not been produced by metasomatic activity. Comparison with data from world-wide carbonatites shows that the $\delta^{13}\text{C}_{\text{PDB}}$ values from the central complex samples are enriched relative to their respective $\delta^{18}\text{O}_{\text{SMOW}}$ values. $\delta^{13}\text{C}_{\text{PDB}}$ enrichment may be due to re-equilibration of

magmatic carbonate with meteoric water, or may be a mantle source feature that has been tapped by several carbonatite intrusions within the Fennoscandian shield area (Alnö, Fen, Kovdor).

6. The pyroxenite-melilitolite-(nephelinite) series share depleted isotopic signatures but enriched trace element signatures, indicative of the influence of trace-element-enriched fluids, just prior to intrusion. The olivine melteigite - ijolite - turjite series share more enriched isotopic signatures but only moderately enriched trace element signatures, suggesting a decoupling, in both groups of silicate rocks, between trace element and isotopic signatures.

7. Comparisons of the trace element data from the melilitite dyke [TU 107], representative of the “depleted” group of samples, and the olivine melteigite [C.AG.48A], representative of the “enriched” group of samples, with trace element data from modern oceanic basalts, suggest that the trace element signatures of the depleted and enriched sources mantle sources at Turiy are similar to those of the HIMU and EM I mantle components, respectively. This finding is not supported by the isotope data.

8. The trace element signatures of the pyroxenites, melilitolites and nephelinites share many similarities in their mantle-normalized spidergram patterns. However, the melilitolites are distinctly depleted in Zr, Hf and HREE, and enriched in Nb, Ta and LREE with respect to mantle normalized values, and with respect to almost all other samples from Turiy. The melilitolites have the steepest REE patterns (LREE-enriched, HREE-depleted) of all the samples analyzed from the Turiy complex, barring the south coast carbonatite dyke [S.W58].

9. The olivine melteigite, ijolite series rocks and the turjite also share similar trace element spidergram patterns. The samples have moderately enriched but flat mantle-normalized spidergrams. The fenite sample NC.153.82 also shows similarities to the trace element signature of the turjite sample TU 119.

10. A combination of trace element data, mineral compositions and petrographic textures suggest that the turjite dykes on the south coast are similar to the products produced by the reaction of an ijolite with a carbonatite, in keeping with the experimental results of Verwoerd (1978). Suitable end-members include a fractionated liquid from the olivine melteigite pipe (based on mineralogical, trace element and isotopic evidence) and the south coast carbonatite dykes (based on trace element evidence). The possible role of a natrocarbonatite, used as a component in the carbonatite end-member in the experiments of Verwoerd (1978), cannot be constrained in the Turiy rocks at this time.

11. Two suites of ijolite-series rocks can be distinguished on the basis of their pyroxene : garnet ratio. The pyroxene-rich samples correspond approximately to the “first ijolite series” proposed by Bulkah & Ivanikov (1984). The garnet-rich samples are concentrated in the “second ijolite series” proposed by Ronenson et al. (1978) and Bulakh & Ivanikov (1984), found enclosing the carbonatite and phoscorite stock in the middle of the central complex. One of these “second ijolite series” samples (C.IJ.14) bears substantial mineralogical, trace element and isotopic similarities to the turjite, and may be at least partially derived from reaction of an ijolitic magma with a carbonatite. The turjite may represent an end product of this reaction.

12. The phoscorites are believed to be representative of magmatic liquids, and not cumulates from a carbonatite magma. They are coarse-grained and contain segregations of specific minerals, either as clusters (e.g. phlogopite), or as veins (e.g. apatite). They

exhibit the most depleted Sr-Nd isotopic signatures of any of the samples analyzed from Turiy, and the isotopic data from them does not overlap with any other rock type from the massif.

13. The carbonatites at Turiy can be divided into three (or possibly four) groups, based on spatial distribution, isotopic, trace element, mineralogical and petrographic variations. The first group is represented by the central complex carbonatites which have moderately depleted isotopic signatures, lying below the Kola Carbonatite Line, and depleted trace element signatures. They have low REE concentrations, low Ce/Yb ratios and $(\text{Nb/Ta})_N < 1$. The second group is formed by the northern complex carbonatites which have enriched $^{87}\text{Sr}/^{86}\text{Sr}$ signatures, lying above the KCL, and overlapping with the isotopic signatures of the melilitolites from the satellite complexes, suggesting possible crustal contamination. Their stable isotope signatures also show heavy $\delta^{18}\text{O}_{\text{SMOW}}$ values. The third type of carbonatite on the Turiy peninsula is represented by the carbonatite dyke sample S.W58. This sample exhibits a Sr-Nd isotopic signature in between those of the central complex and northern complex carbonatites, and has an extremely enriched trace element pattern, similar to that of the “average calcite carbonatite” of Woolley & Kempe (1989). Carbonatite sample C.54.x represents a late calcite-zeolite vein, relatively rich in silicate minerals, with an isotopic signature similar to that of the olivine melteigite - ijolite series, and may represent a fourth type of carbonatite.

14. The effects of fractional crystallization, including cumulus processes, the existence of residual liquids, and liquid immiscibility, on the trace element signatures of the rocks have been partially constrained in this study by using suitable partition coefficients for the elements involved with respect to the minerals present within the rocks. For example, it is believed that fractionation or accumulation of garnet and/or pyroxene have caused

the significant variations in Zr, Hf and HREE signatures between many of the silicate rocks.

15. Mineral compositions can be correlated with the host rock-type. Systematic variations in mineral compositions are seen in apatite, garnet, melilite, mica, nepheline, magnetite, perovskite and pyroxene.

Minerals from the melilitolites contain relatively high quantities of Al (magnetite), Al in octahedral co-ordination (garnet, mica), high quantities of Tschermak components (pyroxene), high REE contents (apatite, perovskite) and $\text{Si} > \text{Sr}$ in apatite compositions. Mineral compositions of pyroxene, apatite, perovskite and magnetite from the pyroxenites share many similarities with those from the melilitolites.

Petrographic and mineral composition similarities are also seen between the olivine melteigite, ijolite and turjite samples. These include a lack of $^{\text{vi}}\text{Al}$ and the presence of $^{\text{iv}}\text{Fe}^{3+}$ in mica and garnet compositions, $\text{Sr} > \text{Si}$ in apatite compositions, lower Tschermak and higher aegirine contents in pyroxene compositions relative to those seen in the melilitolites and pyroxenites. The REE contents in the apatite and perovskite compositions are lower than those of the melilitolites and pyroxenites.

The mica compositions from a suite of central complex carbonatites and phoscorites have distinct compositions, with high Mg# and enriched in $^{\text{iv}}\text{Fe}^{3+}$, while apatite compositions from the two rock-types have low Si and Sr contents.

16. No direct evidence is seen for direct immiscible relationships between any of the rocks examined in this study, based on isotopic, trace element or mineralogical compositions. Similarities in mineralogical compositions suggest that the northern complex carbonatites may be produced by the fractionation of a magma similar in composition to the melilitolites.

The origins of the phoscorites and central complex carbonatites are harder to constrain. Their distinct $^{87}\text{Sr}/^{86}\text{Sr}$ - $^{143}\text{Nd}/^{144}\text{Nd}$ initial ratios that lie beneath the KCL, their unusually depleted trace element signatures and the lack of systematic variation in mineral compositions, provide no obvious links to the silicate rocks present in the Turiy Massif.

17. The Terskii Coast kimberlites are of similar age to the Devonian rocks of the Turiy massif (Kalinkin et al., 1993; Kramm et al., 1993) and have a similar isotopic signature to the “enriched” group of Turiy rocks, falling near the bulk Earth - CHUR intersection on the $^{87}\text{Sr}/^{86}\text{Sr}$ - $^{143}\text{Nd}/^{144}\text{Nd}$ isotope ratio plot. Comparison of data from Beard et al. (1996) suggests that they may also share trace element similarities with the isotopically “enriched” group of samples (olivine melteigite - ijolite - turjite).

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APPENDIX A

ROCK TYPES, SAMPLE NUMBERS + ANALYTICAL TECHNIQUES USED

	<u>Micro- probe</u>	<u>XRF / ICPMS</u>	<u>Sr-Nd</u>	<u>C-O</u>		<u>Micro- probe</u>	<u>XRF / ICPMS</u>	<u>Sr-Nd</u>	<u>C-O</u>
<u>Pyroxenite</u>					<u>Carbonatite</u>				
C.AG.347	x		x		N.63.4	x		x	x
C.90.29	x	x	x		N.64.19	x		x	x
<u>Melteitgite - Ijolite - Urtite</u>					N.187.100	x			
C.AG.48A	x	x	x		N.187.134	x			
C.AG.122	x	x	x		S.W58		x	x	x
C.ND.500	x		x		C.DC		x	x	x
C.90.y	x		x		C.XC	x		x	
C.170.38	x				C.XC5			x	x
C.IJ.14	x	x	x		<u>Turjite</u>				
E.123.106	x		x		TU 119	x	x	x	
WKM	x		x		TU 120A	x			
<u>Melilitolite</u>					TQ	x			
C.5.118	x	x			<u>Late-stage veins</u>				
C.190.105	x	x	x		TD 1	x			
N.64.18	x		x	x	TD 2	x		x	
N.64.18c	x		x		<u>Fenites</u>				
E.133.30	x		x		ij/fen	x			
S.78.15	x				C.190.x	x			
S.162.300	x				N.153.82	x	x	x	
WKT	x	x	x		<u>Xenolith</u>				
TUR I	x		x		W523.2A	x			
TUR II	x	x	x		<u>Dykes</u>				
<u>Phoscorite</u>					TU 105		(x)		
C-K.7	x		x	x	TU 107		x		
C.3.93	x		x	x	<u>Terskii Coast</u>				
C.18.26	x				Terskii kimb.	x			
C.18.30			x		K77A			x	
C.19.163	x		x	x	K77B			x	
<u>Carbonatite</u>					K77C			x	
C.TL.344	x	x			OM 1			x	
C.6.15	x				OM 29			x	
C.18.60	x				ON 165			x	
C.23.270b	x				Sample number abbreviations based on sample location are listed in Appendix Bii. Samples from drill core cover the complete range of rock types from all of the complexes.				
C.23.300	x								
C.49.40	x	x	x	x					
C.54.x	x		x	x					
C.100.a	x		x						
C.100.b	x								
C.100.d	x								
C.100.y	x								

APPENDIX B i

ROCK TYPE DEFINITIONS

All definitions listed below have been taken from the IUGS Subcommission on the Naming of Igneous Rocks (Le Maitre, 1989) except where otherwise stated. Note that the definitions of the melilitolite group of rocks used in this thesis follow the classification scheme described in Appendix C.

pyroxenite a plutonic rock with $M > 90$ and $ol/(ol+opx+cpx) < 40$.
chiefly consisting of pyroxene, with minor amphibole, mica, oxides.

ijolite group:

melteigite a plutonic rock in which $60 < F < 100$ and $70 < M < 90$, and $Na > K$
part of the ijolite series.

ijolite a plutonic rock in which $60 < F < 100$ and $30 < M < 70$ and $Na > K$
mostly nepheline and pyroxene with minor biotite, garnet, titanite.

urtite a plutonic rock in which $60 < F < 100$ and $0 < M < 30$ and $Na > K$
rich in nepheline with minor aegirine and apatite.

melilitolite group: plutonic rocks containing more than 10% modal melilite.

uncompahgrite a plutonic rock consisting chiefly of melilite + pyroxene, oxides,
perovskite, melanite, phlogopite, apatite, calcite.

turjaite a plutonic rock containing nepheline and 60-90% mafic minerals.
Distinguished from melteigite by the presence of melilite.

okaite a plutonic rock similar to turjaite but with hauyne instead of
nepheline.

[uncompahgrite, turjaite and okaite are not recommended usage by the IUGS]

phoscorite a plutonic igneous rock principally containing magnetite and apatite
with accessory calcite, phlogopite and pyroxene.

carbonatite an igneous rock containing at least 50% carbonate minerals.

turjite an alnöite containing melanite, calcite and analcime.

kimberlite An ultramafic rock consisting of major amounts of serpentized olivine with variable amounts of phlogopite, orthopyroxene, clinopyroxene, carbonate and chromite. Characteristic accessory minerals include pyrope garnet, monticellite, rutile and perovskite. (Le Maitre, 1989).

“Group I kimberlites consist of volatile-rich (dominantly CO₂) potassic ultrabasic rocks commonly exhibiting a distinctive inequigranular texture resulting from the presence of macrocrysts (a general term for large crystals, typically 0.5-10mm in diameter) and, in some cases, megacrysts (larger crystals, typically 1-20 cm) set in a fine-grained matrix. The assemblage of macrocrysts and megacrysts, at least some of which are xenocrystic, includes anhedral crystals of olivine, magnesian ilmenite, pyrope, diopside (in some cases subcalcic), phlogopite, enstatite and Ti-poor chromite. Olivine macrocrysts are a characteristic and dominant constituent in all but fractionated kimberlites. The matrix contains a second generation of primary euhedral to subhedral olivine, which occurs together with one or more of the following primary minerals: monticellite, phlogopite, perovskite, spinel (magnesian ulvospinel - magnesiochromite - ulvospinel - magnetite solid solutions), apatite, carbonate and serpentine. Many kimberlites contain a late-stage poikilitic mica belonging to the barian phlogopite - kinoshitalite series. Nickeliferous sulfides and rutile are common accessory minerals. The replacement of earlier-formed olivine, phlogopite, monticellite and apatite by deuteritic serpentine and calcite is common. Evolved members of the group may be poor in, or devoid of, macrocrysts, and composed essentially of second-generation olivine, calcite, serpentine and magnetite, together with minor phlogopite, apatite and perovskite.”

from Woolley et al. (1996) after Mitchell (1995).

APPENDIX Bii

ABBREVIATIONS

Abbreviations used in Appendices A + B iii

<u>First letter</u>	<u>Complex</u>
N	northern
E	eastern
S	southern
W	western
C	central
NC	north-central

Example: N.187.100

1st number	drill core number
2nd number	depth
TD	Late-stage vein
TU	Turiy dyke
OM	olivine melilitite
ON	olivine nephelinite
K	Terskii Coast kimberlite

Abbreviations used in Appendix B iv

A	anhedral	m	medium
aeg	aegirine	mel	melilite
altn	alteration	mg	medium-grained
amph	amphibole	mag	magnetite
ap	apatite	micro	microcrystalline
assoc	associated	min	mineral
biref	birefringence	ne	nepheline
br	brown	nr	near
c (adjective)	coarse	nxt	next
c (noun)	core	orge	orange
cal	calcite	orient	oriented
cancr	cancrinite	overgr	overgrowth
cg	coarse-grained	overgrn	overgrown
chr	chromite	perp	perpendicular
cm	centimetre	pheno	phenocryst
conc	concentrated	phlog	phlogopite
crypto	cryptocrystalline	pl	pale
dk	dark	pleo	pleochroic
dp	deep	prods	products
E	euhedral	prv	perovskite
encl	enclosed/enclosing	pxite	pyroxenite
esp	especially	pyx	pyroxene
euh	euhedral	r	rim
f	fine	resbd	resorbed
feld	feldspar	S	subhedral
fg	fine-grained	sch	schorlomite
fr	fracture	uncompah	uncompahgrite
gb	grain boundary	v	very
gm	groundmass	wo	wollastonite
gr	green	xeno	xenocryst
grt	garnet	xtl	crystal
I	interstitial	xtn	extinction
ilm	ilmenite	xp	cross-polars
immisc	immiscible	yl	yellow
incln	inclusion	&	and
intergr	intergrown	+	with
lge	large	@	at
lwr	lower	//	parallel

Mineral abbreviations from Kretz (1983) where appropriate

APPENDIX Biii: MODAL MINERALOGY

Pyroxenite - melteigite - ijolite - urtite

complex	C	C	C	C	C	C	C	C	E	W
sample	AG.347	90.29	AG.48A	AG.122	ND.500	170.38	90.y	IJ.14	123.106	KM
rock type	pxite	pxite	ol.melt.	melteigite	ijolite	ijolite	ijolite	ijolite	ijolite	ijolite
amphibole	8	5								
analcime				<1						
apatite	<1		2		3		<1	3	1	1
banalsite				<1				<1		
calcite	1		<1	2	2	1	<1			
cancrinite				2		40				
chromite	1									
garnet	1	<1		13	5	6	25	25	6	13
ilmenite						<1		<1		<1
magnetite	10	5	3		3	7		2		3
mica			3	1	2	12	<1	5	2	10
nepheline		<1	6	18	55	10	55	52	60	47
olivine			38							
pectolite						1		1		
perovskite	8	4	1	<1		2		<1	<1	5
pyroxene	70	85	55	62	25	20	17	5	28	20
sodalite								2		
stronalsite									<1	
sulphides				1						
titanite		<1	1				<1		<1	
wollastonite								4		
zeolites			1		5		2		2	

melilitolite

complex	N	N	C	C	E	S	S	W	W	W
sample	64.18c	64.18	190.105	5.118	133.3	78.15	162.300.	KT	TUR 1	TUR II
rock type	cal-mel.	cal-mel.	turjaite	mica-turjaite	turjaite	? (contact)	mag-turjaite	turjaite	"copiamel."	
alteration	x				15	x				
apatite	x			1	3	x		4	4	<1
baddelyite							x?			
calcite	x	28	<1			x	x	1	4	1
cancrinite									16	
chlorite						x				
copiapite						x				
cuspidine							x		<1	
djerfisherite										<1
dolomite								<1		
garnet	x	1	3	2	2	x	x		4	2
hillebrandite									<1	
ilmenite							x?			
magnetite	x	10	5	3	8	x		30	3	7
melilite	x	55	40	46	35		x	20	30	65
mica	x	2	12	22	3	x	x	12	35	6
monticellite		1								<1
nepheline			35	23	27	x	x	28		10
pectolite									<1	
perovskite		2	5	3	3	x	x	5	<1	5
pyroxene	x	<1			4		x		4	<1
strontianite						x	x		<1	
sulphides		1								
vesuvianite	x									
wollastonite		?			<1		x			

all numbers = modal % based on petrographic and SEM observations.

x = mineral present. Representative modal analysis not possible.

altn = unidentified alteration products

APPENDIX Biii: MODAL MINERALOGY (cont.)

carbonatites

complex sample rock	C TL.344	C 6.15	C 18.60.	C 23.270b	C 23.300	C 49.40.	C 54.x	C 100.a	C 100.b	C 100.d	C 100.y	C XC	C DC
amphibole	5		4	4							6	8	
apatite	29	17	15	15	14	22	4	3	4	20	2	9	4
baddelyite		1											
barite				<1									
calcite	40	40	58	70	45	43	52	55	84	65	78	52	46
calzirtite					<1								
chlorite			1						<1		<1	1	
dolomite		5		5		5		34		5	3		46
garnet							15						
ilmenite	3	3	<1	<1	<1				<1		<1	<1	
magnetite	15	26	1	2	16	12			5	5	6	13	
melilite													
mica	4	8	20	3	23	18	20	5	4	2	3	13	
monticellite													
quartz													
perovskite	<1						2						
pyrochlore											<1	<1	
pyroxene	4											3	
strontianite			<1	<1			<1		<1	<1	<1		
sulphides					1			3	2	2		<1	3
titanite													
zeolites							7						
zircon			<1								<1	<1	

carbonatites

complex sample	N 64.19	N 63.4	N 187.100	N 187.134
apatite	<1	<1	1	2
calcite	70	47	55	52
garnet	1	1	7	6
ilmenite			<1	
magnetite	2	2	1	2
melilite	6			
mica	11	2	17	26
monticellite	10			
quartz			3	
perovskite	?			
pyroxene		48	15	12
titanite			<1	

phoscorites*

complex sample rock type	C K.7 phosc	C 19.163 phosc	C 3.93 phosc	C 18.26 phosc. pyxite.
amphibole	1		4	2
apatite	<1	10	1	1
calcite	27	18	58	5
chlorite		<1	1	
dolomite		2		
ilmenite	<1	1		
magnetite	28	47	8	18
mica	30	22	28	22
perovskite				2
pyroxene	10			44
strontianite		<1		
sulphides	3			2
unknown				5
zircon		<1		

* texturally heterogeneous. Apatite more abundant than indicated from thin-section observation

APPENDIX B iv: SAMPLE DESCRIPTIONS

Sample Number Rock type	Mineral	Modal %	Grain size	Habit	Pleochroism / Colour	Altn	Comments
C. AG 347 pyroxenite [banded]	cumulate pyroxenite with rare interstitial amphibole and chromite exsolution lamellae in pyx cores						
	amphibole	8	m	A	c: pink/red - orge/br r: bl/gr ey - pale gr		assoc. + mag 32° xtn
	apatite	<1	crypto				
	calcite	1	f-m	I/A			encl. prv & grt in amph-rich area
	chromite	1	f-m	E			exsolution rods in pyx cores following cleavage planes
	garnet	1	f-m	S-A	yellow-brown/pink		assoc. + calcite
	magnetite	10	m-c	A/C		prv rims	solitary xtls or mag rims
	perovskite	8	m	S			rims most abundant near mag
	pyroxene	70	m-c	A/C	faint green/br-pink rims pinker - Ti rich		no chr in rims. Rims lwr biref Non-uniform xtn core-rim gradual core-rim transition
C. 90.29 pyroxenite [banded] (f-m)	almost mono-minerallic						
	amphibole	5	f-m	A/I	blue/grey - green		inclns unknown. enlc. mag occurs in coarser bands along grain boundaries
	garnet	<1	micro	A/I			resorbed, blebby, prv rims
	magnetite	5	f-m	A/I			
	nepheline	<1	micro	I			thin irregular veins.
	perovskite	4	f	A/I			assoc + amph & grt, & mag rims
	pyroxene	85	m	A			equigranular. Other minerals more interstitial in finer parts
	titanite	<1	micro	I			
C. AG 48A olivine melteigite	from the eastern part of the Central Complex						
	apatite	2	m	E			needles, rare inclusions, usually infilling vesicles
	calcite	<1					f-m inclns. in olivine, tiny xtls around olivine edges, f-m isolated & glomeroporphyritic areas in gm
	magnetite	3	f-m	A			mottled pleo. rims/incl on/in ol or in gm. Often embayed/skeletal
	mica	3	f	A	orge-br, straw-green, br-dk br, khaki-green		Deep orge + pyx, orge + olivines
	nepheline	6	c	S		dusty	aegirine inclns. altn propnl. to degree of non-euhedrality
	olivine	38	c	A		Fe in fr.	<1 mm. xeno/pheno? rounded inclns = mag, etc.
	perovskite	1	f	A			rims around mag?
	pyroxene	55	m-c	S	c: pl gr/yl, r: pink gm: yl-gr		gm finer grained, cumulate, rare euh. aeg. (fg) in nepheline in gm (hard to distinguish)
	titanite	1	f	A			in gm.
C. AG 122 melteigite	medium grained, unbanded, relatively homogeneous ijolite of the first series						
	analcime	<1	m	I			isotropic
	banalsite	<1	crypto	I			along grain boundaries
	calcite	2	f	I			assoc + cancrinite
	cancrinite	2	f	I			surrounding calcite. high biref
	garnet	13	f	S/C	c: brown - r: yellow		overgroth/resorbed rims, mottled, banded, poikilitically encloses everything in its path
	mica	1	f	S	orge/straw - khaki green/brown		assoc + pyx/grt
	nepheline	18	f-m	A/I		dusty (c)	altn. or inclusions in cores?
	perovskite	<1	f	A			rims around magnetite or rare inclusions in pyroxene
	pyroxene	62	f-c	A	yellow-green		blotchy xtn, irregular edges, includes ne, phlog, perov, and larger mag, grt, ne.
	sulphides	1	f	A			blebby, often with grt
	unknown	<1					Na ₁₋₃ (Ca,REE) ₅ TiSi ₂ (O,F,OH) ₁₈

Sample Number Rock type	Mineral	Modal %	Grain size	Habit	Pleochroism or Colour	Altn	Other features
C.ND.500 ijolite	first? generation of ijolites, equigranular and relatively homogeneous.						
	apatite	3	m	S			elongate, rounded, unzoned in veins or parallel bands
	calcite	2	f	I			assoc. + zeolites
	garnet	5	c	S/I	c: brown - r: straw		ap. inclns. assoc + mag zoned
	magnetite	3	m	A			adjoining grt in places
	mica	2	m	S	pale green - green		assoc + lge mag xtls
	nepheline	55	m-c	A/S		dusty (fr & gb)	fractured uniform xtn
	pyroxene	25	c	S	yl-green faint pl.	dusty (fr & gb)	heavily fractured, rare inclns,
	zeolites	5		A/I			within ne & along gb. high birefringence
C.170.38 ijolite [banded]	central ijolite, 2nd series, heavily altered.						
	calcite	1	f	A			blebs
	cancr (1)	25		A		grey/dusty	invaded by grt veins. dusty in xp
	cancr (2)	15	micro	A	colourless		1st-2nd order biref. cumulate. replacing ne? Occurs between cancr. (1) and ne. - maybe natrolite?
	garnet	6	f-m	A/I	yl in ne-rich dk br in pyx-phlog with yl rims.		stringers run through (1) outlining relict grain boundaries? & around phlog. Thin rim around mag. Thicker rim around mag. in ne. yl parts often non-isotropic
	ilmenite	<1	f	A			blebs in magnetite
	magnetite	7	c-m	A/E			coarse anhedral in pyx layer
	mica	12	m-c	A/E	pl yl - green		equant med-gr in ne/prv/grt surrounded by grt, interspersed with pyx in the mafic layer
	nepheline	10	m	S/E			equant, clear triple jncts, occur in specific areas esp with prv
	pectolite	1	f	S			rare, with cancrinite?
	perovskite	2	m	A/S	brown - beige		inclns in cores, pale rims on mag. stringers of prv invade grain boundaries between ne. grains at triple junction made by 2 ne + 1 prv
	pyroxene	20	c-m	A	dk green - yl/green	grn edges & cl pl	33° xtn angle
C.90y ijolite	unbanded ijolite of the second series in the Central Complex						
	apatite	<1	f	S			thin needles
	calcite	<1	f	I			along ne grain boundaries
	garnet	25	m	S	dk br - pl br (conc. z)		intergrown with & encl by pyx encl or surrounding ne
	nepheline	55	c	S/E		dusty (fr & gb)	solitary inclns in cores
	mica	<1	f	A	pl orge - br		assoc + pyx
	pyroxene	17%	m	A	deep gr/blue - green		assoc + & encl by grt
	titanite	<1	f	A			rare
	zeolites	2					prevalent @ ne triple junctions high biref. clear

Sample Number Rock type	Mineral	Modal %	Grain size	Habit	Pleochroism or Colour	Altn	Other features
C.IJ.14 ijolite	2nd-stage ijolite-urtite, very similar to turjite TU 119. Coarse-grained and generally euh.						
	apatite	3	m	S			needles, minute inclns
	banalsite	<1	crypto				along grain boundaries
	garnet	25	m-c	S	v. dk brwn - br/yl		surrounding ne& pyx or intergrown with phlog.
							nr wo (phlog) euhedral with strong concentric zonation
	ilmenite	<1	f	A			exsolved from magnetite
	magnetite	2	f-m	A			assoc. + grt
	mica	5	m	A/S	brt orge - brown		solitary-cumulate, or intergrown with/replacing pyx.
	nepheline	52	m-c	S/E			trails of tiny inclusions
	pectolite	1	f	S			radiating swarm
	perovskite	<1	f-m	A			encl. by phlog. assoc + phlog/ pyx/ap/wo cluster
	pyroxene	5	c	S(E)	pale green		being replaced by phlogopite
	sodalite	2	m-c	A			v. irregular/resbd, assoc + ne
	wollastonite	4	m-c	S-E/I			high birefringence
E.123.106 ijolite	apatite	1	m	S			needles, cut pyx,
	garnet	6	m	A/S	mottled deep brown		few solitary inclns
	mica	2	f	S	brt orge - brown		irregular at edges, enclosing small xtls
	nepheline	60	m-c	A		dusty	assoc + pyx rims, or intergrown with garnet
	perovskite	<1	f	S			altm throughout Irreg rims & contacts esp + pyx (some orient)
	pyroxene	28	m-c	S	c: gm-yl, r: dk green		rare
	stronalsite	<1	crypto				rims irregular
	titanite	<1	f-m	S			along grain boundaries
	zeolite	2	m	A			nxt to pyx & grt. Zr-rich irreg. contacts + ne, high biref
WKM ijolite	relatively fine-grained melteigite from the early series on the shore at Kusnavalok on the west coast.						
	apatite	1	f	S			blebs/needles in ne.
	garnet	13	f	A	brown/black		fg cumulate
	ilmenite	<1	f	A			exsolution blebs in mag
	magnetite	3	f-m	A			blebs/crystals in garnet
	mica	10	m	A/S	pl orge - brown		contains pyx blebs. parallel orientation of xtls in thin bands
	nepheline	47	m	A			clear, contains inclusions,
	perovskite	5	f	A			hard to distinguish from grt
	pyroxene	20	f	A/S	pl green		
C.190.105 turjaite	irregular fine and medium-grained sections. Fresh and unaltered.						
	calcite	<1	f	I			occasional veins/stringers
	garnet	3	f-m	A/S	c: brown - r: beige (can be reversed)		concentrically zoned, (coarse) often partially/wholly encl. prv.
	magnetite	5	f	A/S			
	melilite	40	f-m	A			equant, aligned fractures,
	mica	12	f-m	A	yl - br (+ green rims)		thin, often partly encl. other xtls.
	nepheline	35	f-m	A		Fe in fr.	inclusions in fractures. Fresh
	perovskite	5	f-m	A	deep brown		lots of inclusions. Also colourless rims around mag
C.5.118 uncompahgrite	coarse-grained and medium grained parts of the section, but a lack of structure within the two halves. Generally fresh, with interlocking crystal relationships.						
	apatite	1	f	S			inclusion-free
	garnet	2	crypto	A			dusty melilite grain boundaries
	magnetite	3	f-m	S			prv (+grt?) rims
	melilite	46	c-m	A		more in cg	fractured, unidentified inclns, deep yellow birefringence,
	mica	22			straw-pale orange green rims perp. cleav.		elongate // cleavage, faintly banded
	nepheline	23	m	A			rounded, some trains of inclns.
	perovskite	3	f	S	brown		hard to distinguish

Sample Number Rock type	Mineral	Modal %	Grain size	Habit	Pleochroism or Colour	Altn	Other features
N.64.18 carb-melilitolite	overgrowth sequence around magnetite: grt + prv blebs -> monticellite' -> calcite -> melilite. see Fig. 5.9b						
	calcite	28	m-c	A/I			interstitial/poikilitic. v. broad twin planes - almost laths
	garnet	1	crypto	I			part of mag reaction sequence
	magnetite	10	m	A/S			postage stamp edges
	melilite	55	m-c	A		heavy fr & gb	postage stamp edges next to calcite some grains completely altered
	mica	2	m	S	pale yellow-green		
	monticellite	1	f	A			part of reaction sequence
	perovskite	2	f	A	brown solitary		blebs around mag & solitary
	pyroxene	<1	f	A			tiny. or monticellite??
	sulphides	1	f	A			
N.64.18c contact sample carb-mel:	carbonate-melilitolite next to cal carbonatite: Carbonatite side is 100% calcite, showing uniform xtn but movement along cleavage planes. Contact lined with fine-coarse subhedral magnetite crystals.						
	apatite		f-m	E			
	calcite		f-m	I			fresh, abundant, esp. assoc + mel.
	garnet		f	A/I	deep brown solitary		altn. prod. of mel. & as abundant grains surrounding mag & phlog
	hydrous magnetite		f m	A/I S			esp in mel altn areas finer-grained away from the contact + postage-stamp edges
	melilite		m	A		fr & gb	blue biref in non-altered parts
	mica		m	S	straw - green		occasionally has calcite inclns
	pyroxene		micro	E			pure diopside. altn product of mel
	vesuvianite		micro	S			mel altn product
E.133.30 turjaite	more altered than the other samples, with chlorite & zeolites, mostly along grain boundaries & especially concentrated in melilite and pyroxene. Banded (f & m -grained bands, each 1cm wide).						
	alteration	15	crypto				messy/clustered. unidentified
	apatite	3	f	E			clusters along gb, or isolated xtls in mag & prv. few inclusions
	garnet	2	f	I			altn around melilite
	magnetite	8	f	S			rounded xtls, evenly scattered.
	melilite	35	m	A		gb & fr	often enclosing other minerals
	mica	3	f-m	A	straw - pl br + r: green		inclns in solitary xtls. Also as green rims around prv around mag
	nepheline	27	f-m	A			clear. few inclns along fractures
	perovskite	3	m	A/S	deep brown solitary pale rims on mag		strong twinning
	pyroxene	4	m	E/A	pale green. non-pleo		irregular resorbition features
	wollastonite	<1	f	A			trace
S.78.15 ex-turjaite	large proportion of the rock is very altered - modal mineralogy not representative of original sample.						
	apatite		f	S			trace
	calcite		crypto				part of alteration products
	chlorite		f	A			part of the grand alteration scheme
	copiapite		f-m	A			highly birefringent, colourless.
	garnet		m	A	c: br/yl - r: beige		great concentric zonation. can be assoc. + mag
	magnetite		m	S			rare and rounded
	mica		f-c	S/E	green - brown		
	nepheline		m	A		dusty	lots of inclusions, fairly abundant
	perovskite		f-m	A/E	deep brown solitary beige rims on mag		c: inclusions. fairly abundant
	sodalite?						?part of alteration.
	strontianite		crypto	A			Pink in hand specimen rare/trace

Sample Number Rock type	Mineral	Modal %	Grain size	Habit	Pleochroism or Colour	Altn	Other features
S.162.300 contact	ijolite - turjaite contact sample. Mineral %s meaningless.						
	apatite		f-m	S			vein in one area. No inclusions.
	baddelyite		crypto	A			trace
	calcite		f	A			trace in melilitolite section
	cuspidine		f	A			trace
	garnet		f-m	A/S	dk br solitary; paler rims around grt		interstitial veins around everything
	ilmenite		f	A			exsolution from magnetite
	magnetite		m	A			rare, and with prv rims
	melilite		m	A/I			highly fractured, abundant inclns of mica, prv, etc,
	mica		f	S	orange - green		small xtls in mel, similar in ijolite, but few thin bright orange-brown xtls near pyx, like IJ II
	nepheline		m	A/S			clear, rounded, esp at edge of ijolite areas. One cg xtl, inclns in fractures
	perovskite		m-c	A/S	brown		overgrown by grt in mel-zone
	pyroxene		m	A/S	pale green		abundant and abundant inclusions very little difference between pyx in the two magma types. Ijolite section
	strontianite		crypto	A			pyx has a few aegirine-rich rims.
	wollastonite		m	A/S			traces along grain boundaries contains phlog, pyx, grt, ne inclns high biref at 90° cleavage
WKT turjaite	the finest-grained of all the melilitolites sectioned. Medium and fine-grained areas. Very fresh, unaltered and unfractured. Type-locality rock.						
	apatite	4	f	S			hard to identify
	calcite	1	f	A			rare
	dolomite	<1	f	A			blebs assoc + calcite
	magnetite	30	f	A/S			evenly distributed
	melilite	20	f	A/S			finer-grained than nepheline
	mica	12	f-m	S	straw - pl br/khaki		
	nepheline	28	f-m	A/S			the coarsest phase. very fresh
	perovskite	5	f	A	dark brown		hard to distinguish from mag
TUR I turjaite	late-stage, lots of hydrous minerals. Cancrinite occurs instead of nepheline.						
	apatite	4	m	S/E			stubby, cross-cuts many of the phases
	calcite	4	f-m	I			often assoc + diopside
	cancrinite	16	c	A			lots of inclns, pectolite, sulphides, etc blue pleochroism
	cuspidine	<1	f	A			trace in cancrinite
	garnet	4	f-m	S/I			brown subhed xtls, or colourless rims around prv around mag, or altn rims around melilite sch -> andradite
	hillebrandite	<1	f	A			trace
	magnetite	3	m	A/S			prv rims
	melilite	30	m-c	A		garnet gb	heavily fractured - radiating mottley assortment of inclusions
	mica	35	c	S	straw - dp br, r: green		early-crystallizing phase
	pectolite	<1	f	S			traces in cancrinite
	perovskite	<1	f	I	colourless		rims around magnetite
	pyroxene	4	f-m	A			normally assoc + calcite
	strontianite	<1	f	A			traces in calcite or around apatite

Sample Number Rock type	Mineral	Modal %	Grain size	Habit	Pleochroism or Colour	Altn	Other features
TUR II turjaite	almost the melilite-bearing equivalent of a pyroxenite. Coarse-grained and melilite-cumulate						
	apatite	<1	f	A			a couple of tiny crystals
	calcite	1	f	I			rare
	djerfisherite	<1	f	A			rare K-sulphide
	garnet	2	f	A			occasional veins cutting through mag
	magnetite	7	m-c	S			& prv. Occasional altn around mel
	melilite	65	m-c	A/S		rare, gb	overgrn. by prv & phlog. Rare fg rounded anhedral xtls at mel. triple junctions with no overgrowths almost cumulate. Few inclns. rare fracturing.
	mica	6	f-m	A	straw - green		most common as overgrowth around [magnetite + prv]. Rare solitary. one crystal at a grain boundary
	monticellite	<1	f	A			few fractures containing small incln.
	nepheline	10	f-m	A			inclusions in deep brown cores
	perovskite	5	m	A/S	deep brown solitary pale rims [+ on mag]		gradational brown - beige
	pyroxene	<1	f	A			one xtl found in a grain boundary
C.K.7 phoscorite	section split into 3rds: magnetite & mica, middle third is filled with a vein of pyroxene and calcite and rare sulphide, which seems to flow between the coarse magnetites at the top of the section.						
	amphibole	1	m	S	colourless		one radiating daisy, richterite
	apatite	<1	f	S			rare
	calcite	27	m-c	A/I			poikilitic or interstitial to mag & mica. Uniform xtn. Cleavage strain.
	ilmenite	<1	f	A			exsolution blebs in magnetite
	magnetite	28	c	S/E			
	mica	30	c-m	S	orange-straw (reverse pleo cores only)		normally kinked, mottled, pale rims
	pyroxene	10	f-m	E	pale green		stubby, one larger grain,
	sulphides	3	f-m	A			immiscible blebs in pyx-rich areas
C.3.93 phoscorite	extremely coarse-grained and inhomogeneous. Modal % only apply to the section observed.						
	amphibole	4	f	E			tiny rhombohedral xtls in a vein assoc + inclusion-rich calcite
	apatite	1	f	S			intergrown with mag in one area
	calcite	58	f-c	A/S		dusty v. dusty	1) m-c, A/S, interlocking 2) vein, v. rich in inclns. fg, between mica and (1), elongated flow texture
	chlorite	1	f	A			alteration around mica & amph.
	magnetite	8	c	S/E			mostly S/E except where intergrown with apatite
	mica	28	c	S	deep orange - straw		<2cm, concentric zoning, green rims, additional tiny green xtls in gm with amphibole-calcite vein
C.18.26 pyroxenite invded by phoscorite.	non-uniform magnetite, phlogopite, pyroxene, pyroxene/calcite-rich areas. Evidence of mixing is clear, but precise relationships are hard to define. Few changes in pyroxene colour or texture.						
	amphibole	2	f-m	S	khaki green - brown		43° xtn
	apatite	1	f	S			rare
	calcite	5	f-c	I			few inclns, mostly small, a few xtls poikilitically enclosing pyx
	magnetite	18	m	A			cumulate -> cg masses, often surrounded/cemented by prv rims
	mica	22	m	A/S	dp/pl orge - brown		mottled, inclns of pyx, mag, cal, can become skeletal
	perovskite	2	f-m	A	dp brown = isolated beige rims on mag		few inclns. looks like prv
	pyroxene	44	f-m	S	pl bl/green - green	chlorite	cumulate, fractured, often encl'd. in sieved mica
	sulphides	2	f	A			chalcopyrite blebs, scattered,
	unknown	5	f-m	I		v. dusty	veins around pyx & in pyx-rich area less altered - blue biref - melilite? pyx? grt?

Sample Number Rock type	Mineral	Modal %	Grain size	Habit	Pleochroism or Colour	Altn	Other features
C.19.163 phoscorite	banded: calcite + coarse mag + coarse mica -> apatite + finer mica & mag						
	apatite	10	f	E			vein cutting the section, solitary inclusions.
	calcite	18	f-m	A			blotchy/blebby, lots small inclns
	chlorite	<1	f	A			rare altn product of mica
	dolomite	2	f	A			blebs assoc with/in calcite
	ilmenite	1	f	A			blebs & lamellae exsolved from mag
	magnetite	47	c	A/E			anhedral adjacent to apatite, S/E elsewhere. inclns of mica, ap, cal.
	mica	22	m	E	c: pl yl-pl green r: dp orge-pl orge		picture-frame concentric zoning, one cluster of small mottled xtls
	strontianite	<1	f	A			inclns in calcite
	zircon	<1	f	S			rare
C.TL.344 carbonatite	inhomogeneous, areas of crystal segregation, magnetite conc. in one band across the centre of the section.						
	amphibole	5	f-m	S			colourless radiating richterite
	apatite	29	f-m	A/S			1) most fg, along calcite gb and in triple junctions. inclusions conc. in cores, and grains randomly oriented.
	calcite	40	f-m	A/S			2) coarser, rare, few inclns
	ilmenite	3	f	A			triple junctions. Rare strain/banding
	magnetite	15	c	E/S			blebs in magnetite
	mica	4	f-m	S	pl orange - pl green		fg blebs interstitial to cal. immisc?
	perovskite	<1	m	A	brown		kinked
	pyroxene	4	m-c	S	pale green. non-pleo.	richter.	one elongate xtl, zoned, inclusions, heavily fractured & altered
C.XC carbonatite	amphibole	8	f	S	blue/green - blue/grey		elongate needles, clusters, repl pyx
	apatite	9	f-m	A			veins along cal gb. rounded, few coarser xtls contain inclns
	calcite	52	f-c	A/S			interlocking, few inclns, clear,
	chlorite	1	f	S			trace altn of phlog/amph
	ilmenite	<1	f	A			exsolution features in magnetite
	magnetite	13	f-c	A/E			irregular, sometimes skeletal
	mica	13	f-c	A/E	orange - pl orge/green		virtually unzoned, large solitary xtls or clusters of small ones
	pyrochlore	<1	f	A			rare
	pyroxene	3	m	A	v. pl green, non-pleo	extensive amphibole	heavily fractured
	sulphides	<1	f	A			rare (more common than trace)
	zircon	<1	f	A			trace
	zirsinalite	<1	f	A			trace
C.6.15 carbonatite	banded, not in equilibrium, different generations of the apatite, magnetite & calcite bands.						
	apatite	17	f-m	S			some v. fg bands, others f-m
	baddelyite	1	f	I			rare inclns
	calcite	40	f-m	A/S			thin rims around apatites in 1 band
	dolomite	5	f	A			fg banded/oriented, coarser parts
	ilmenite	3	f	A			have more triple junctions & rare subhedral xtls. Some intergrown
	magnetite	26	f-g	A			and embayed textures.
	mica	8	f-m	S	c: orange - green r: pale yellow - green		blebs assoc. with calcite exsolution blebs in magnetite interstitial to apatite. Coarseness reflects coarseness of the host band not cumulate. mostly in band with coarsest mag mostly fg, few mg xtls

Sample Number Rock type	Mineral	Modal %	Grain size	Habit	Pleochroism or Colour	Altn	Other features
C.18.60 carbonatite	equigranular and unbanded.						
	amphibole	4	f	S/E	pl yl - bl/green	rare	stubby radiating clusters or rhombs
	apatite	15	f	A/S			along gb, rare large solitary inclns
	calcite	58	f	A/S			undulating gb, little strain
	chlorite	1	f	A			altn prods of mics & amphibole
	ilmenite	<1	f	A			exsolution blebs in mag
	magnetite	1	f-m	A			1 mg xtl, & 1 cluster of fg near phlog
	mica	20	f-m	A	c: pl orange - dp green r: deep orge - pl green		blotchy, 1 cluster + solitary xtls
	strontianite	<1	crypto	A			stubby.
	zircon	<1	f	A			inclusions with calcite trace
C.23.270b carbonatite	very fresh sample. Richterite late-stage.						
	amphibole	4	m	S			radiating masses, normally in veins.
	apatite	15	f-m	A			one vein cutting apatite colourless
	barite	<1	f	A			tightly clustered inclns in core
	calcite	70	m	A			inclusion-free rims
	dolomite	5	m	A			trace
	ilmenite	<1	f	A			interlocking, enclosing other mins
	magnetite	2	m	S/E			along gb, slight strain thick twins
	mica	3	m-c	S	c: pl orge-pl green r: deep orge-pl orge		abundant minute inclns exsolution blebs in magnetite single xtls thin rims
	strontianite	<1	f	A			trace amounts. assoc with calcite
C.23.300 carbonatite	faint banding emphasized by magnetite and apatite. fresh.						
	apatite	14	f	S/A			veins & bands, equigranular, free of inclns; few rare phenocrysts anhedral with inclusions.
	calcite	45	f-m	A/S			interlocking gb, few small inclns clustered in a few xtls cal-rich parts more subhedral, fewer inclns.
	calzirtite	<1	f	A			trace
	ilmenite	<1	f	A			exsolution blebs in mag
	magnetite	16	f-m	S/A			skeletal, encl ap & cal. Gb most irregular next to ap.
	mica	23	m	S	c: pl yl - green r: orange - pl green		random distribution, but some preferential orientation in bands. rare
C.49.40 carbonatite	sulphides	1	f	A			
	more homogeneous than most carbonatites and finer-grained. Very fresh.						
	apatite	22	f	S		(dusty)	veins, randomly oriented. Inclns // long axis & scattered through grains
	calcite	43	f-m	A			embayed, cumulate, slightly oriented. Some cleavage strain.
	dolomite	5	f	A			assoc + calcite
	magnetite	12	f-m	E/S			fine-euhedral and cumulate, or medium-subhedral, embayed next to apatite (+ calcite) Skeletal.
C.54 x carbonatite	mica	18	m-f	S/E	deep orange - straw straw - green		picture-frame concentric zoning Rare inter-grain zoning consistency
	hydrothermal vein, not in equilibrium, concentric zoning in calcites (cathodoluminescence) banded.						
	apatite	4	micro	A			"fried egg" effect where fg cores are cemented by a later stage of apatite
	calcite	52	m	E			occurs in swirls and veins
	garnet	15	f-m	A/S	brown - yellow (c-r)		conc zoning + strontianite inclns rhombohedral.
	mica	20	f-m	A/S	colourless - pl green	cp	irregular & concentric zoning sometimes skeletal
	perovskite	2	f-m	A	brown		kinked, assoc + grt & prv bands assoc + or enclosed by grt
	zeolites	7	f-m	I			no inclns unzoned. natrolite interstitial to cal rhombs radiating, assoc + ap veins

Sample Number Rock type	Mineral	Modal %	Grain size	Habit	Pleochroism or Colour	Altn	Other features
C.100.a carbonatite	cal carbonatite cut by dolomite band						
	apatite	3	f-m	S			rounded in coarse carbonatite micro. in dol band, + thin vein of ap along the boundary. mg aps contain large inclns of calcite strained. interlocking fg band with oriented xtls
	calcite	55	f-c	A			
	dolomite	34	f	S			
	mica	5	m-c	S	deep orge - deep brown		
	sulphides	3	f-m	S			one vein cutting across the section
C.100.b carbonatite	almost pure carbonate, veined by mag-ap-phlog mixtures.						
	apatite	4	f	S			no inclns, veins, cumulate,
	calcite	84	f-c	A			interlocking gb, unaltered, few inclns
	chlorite	<1	f	S			altn prods of micas
	ilmenite	<1	f	A			exsolution blebs in magnetite
	magnetite	5	m-f	A			often skeletal, encl. variety of mins
	mica	4	f-C	S	c: bright orange - green r: dp red - bright orgage		1 cg xtl, rest fg. Fg in mag-rich area
	strontianite	<1	f	A			trace assoc + calcite
C.100.d carbonatite	texturally inhomogeneous						
	apatite	20	f-m	A/S			veins & clusters, coarser xtls contain abundant inclns in cores, fg/oriented xtls have few inclns
	calcite	65	f-m	A			interlocking, fg oriented/elongate
	dolomite	5	f	A			blebs assoc + calcite
	magnetite	5	m	A			one radiating rosette, skeletal,
	mica	2	m	S	pl orange - v. pl green		little zonation
	strontianite	<1	f	A			inclns assoc with calcite
	sulphides	2	f	A			stringers cutting across section
C.100.y carbonatite	banded						
	amphibole	6	f	S	grey-green		clustered, banded, elongate
	apatite	2	f-m	A			isolated grains in phlog/mag-rich areas. Few inclns, but coarser xtls contain calcite blebs
	calcite	78	f-m	A		(SrCO ₃)	interlocking gb, incln-rich calcite occurs in mag/phlog/amph bands cal bands between are clearer with larger aps with calc inclns.
	chlorite	<1	f	A	green		altn prods of phlog/amph.
	dolomite	3	f	A			blebs with cal and in triple junctions
	ilmenite	<1	f	A			exsolution features in mag
	magnetite	6	f-m	A			irregular, rounded, solitary, most fg.
	mica	3	f	S	c: dp orange to straw r: green - brown		kinked, clustered, occurring in specific bands, very varied pleo. Assoc + mag
	pyrochlore	<1	f	A			trace
	strontianite	<1	f	A			trace with calcite & apatite
	zircon	<1	f	A			incln in apatite + other Nb-Ta mins
N 63.4 pyx carb	coarse-grained, very crumbly in hand-specimen, and texturally inhomogeneous						
	apatite	<1	f	S			trace
	calcite	47	c	E/S			some strain,
	garnet	1	f	A/I	brown - yellow		altn rims in calcite & around mag
	magnetite	2	c-m	A			grt (prv?) rims
	mica	2	f-m	A	pale straw - green		
	perovskite	<1	f	A/S	brown		tiny grains & rims around mag?
	pyroxene	48	m	S/E	pl yl-green pleo	fr & gb	fractured

Sample Number Rock type	Mineral	Modal %	Grain size	Habit	Pleochroism or Colour	Altn	Other features
N 64 19 mel carb.	out of textural and chemical equilibrium with altered melilite.						
	apatite	<1	f	S			one cluster of rounded equigranular and inclusion-free xtls
	calcite	70	m-c	A/S			rounded, rare strain indicators
	garnet	1	f	A/I	yellow		rare rims around mag
	magnetite	2	f-m	A			one cluster showing postage-stamp edges
	melilite	6	m	A		fr & gb	fractured, blue biref, many inclns
	mica	11	f-m	S	pale straw - green		grains occur in clusters when not assoc with other silicate minerals
	monticellite	10	m	A/I		Fe-stain fr & gb	rounded, high biref,
N 187 100 carbonatite	triple banded, centre band carbonate-dominated, other two bands grt-pyx-mica (& quartz) dominated						
	one end of the section: mag strings, pl-dp br grt, v. pl kinked phlog, quartz veins.						
	other end: small yl rounded grt xtls, fg S/E pyx (+ conc. zoning), orge/gr phlog, interstitial calc.						
	apatite	1	f	S			tiny veins in one silicate-rich end & central cal.
	calcite	55	f-c	A			at least 2 generations - 1) coarse rounded phenos. some strained,
	garnet	7	f-m	A/E	brown-yellow		2) fg, interlocking, some oriented concentrically zoned, extensively fractured, euhedral yl/paler
	ilmenite	<1	f	A			one end rounded yellow xtls
	magnetite	1	f-m	A			trace
	mica	17	f-c	S/E	v. pl straw - green pale orange - green		resorbed, in one band, kinked, banded, varied compositions at the two ends orge mica more euh
	pyroxene	15	f	A/E			clustered, finer grained at one end
N 187 134 carbonatite	quartz	3	f	A			cumulate veins of tiny xtls.
	titanite	<1	f	A			trace
	banded, coarse pyx-phlog-grt-mag-rich ends of the section, finer-grained cal-phlog rich centre						
	apatite	2	micro	A			tiny clusters & veins
	calcite	52	f-c	A			interlocking, outlined in the centre by tiny elongate phlogpites along gb
	garnet	6	f-m	E/S			clusters of f-m deep brown zoned xtls with skeletal rims, some encl mag
	magnetite	2	f-m	A/S			Tiny euh grts in cal band too often enclosed by garnet. Small xtls in cal band more euhedral, solitary
	mica	26	f-c	A/S	c: pl orge - green gm & r: dp orge - brown		fg elongate xtls along gb in cal band
	pyroxene	12	f-c	S/E	c: pl gr; r: dk gr	calcite & dusty	coarse subhedral xtls one end with an invading mass of small orge xtls
							concentric & sector zoning altn strong in places. Lots inclns aeg rims
TU 119 turjite							Some v. fg pyx in cal band.
	type-locality turjite from the south coast dykes. relatively homogeneous and coarse-grained.						
	analcime	16	m-c	I			replacing nepheline?
	apatite	1	f	S			rare in gm
	calcite	8	m-c	I			rarely contains small euh. grt
	cancrinite	<1	f	A			trace amongst analcime
	garnet	55	f-c	A/E	dk br - pl yl		concentrically & sector-zoned anhedral dk br sch veins
	magnetite	2	f	A			rare, with grt
	mica	12		S/E	c: pl yl - brown r: deep orge - green		rare aeg inclns & annite rims
	perovskite	1	f	A			hard to distinguish from grt
	pyroxene	<1	f	A	green		one aeg xtl in phlog
	strontianite	<1	f	I			trace in gm & around cal xtls

Sample Number Rock type	Mineral	Modal %	Grain size	Habit	Pleochroism or Colour	Altn	Other features
C.190.x ijolite/fenite contact	ijolite/contact:	fine-grained green pyroxene and interstitial nepheline, with deep brown microcrystalline garnet dominating the contact zone, and rare but coarser-grained phlogopite (pale orange - deep brown). Banded wrt garnet layers in pyroxene.					
	fenite:						
	barite	2	f-m	A			moderate high relief
	garnet	-	f	A			dominating contact zone
	hydrous	25	f-m	A			natrolite? moderate biref, mottled in xp, relatively unaltered
	orthoclase	35	c	E		dusty	varying degrees of altn
	pectolite	<1	f	S			trace
	pyroxene	27	m	E/S	deep green-brown		commonly in small clusters, 2 types 1) lge euh., and 2) v. fine long needles radiating away from contact and 5cm distant from it. Zone 1cm next to contact smaller more euh and much darker crystals.
	sodalite	10	m-c	A			nearer the contact
	strontianite	<1	f	A			trace
	tausonite	<1	f	A			trace with titanite
	titanite	<1	f	S			along contact
	wollastonite	<1	f	A			inclusions in aegerine
NC.153.82 fenite	fenitized gneiss from the north of the central complex.						
	amphibole	8	f	S/E	blue-green		needles, radiating around micas
	apatite	<1	f	E			trace
	barite	1	f?	S			+ celestine & BaCO ₃
	calcite	3	f	I			
	"hydrous" ?	10	m	S			Ca-rich, unknown
	mica	2	f	S/E	orange-brown		elongate, often surrounded by tiny amphibole needles
	orthoclase	55	m-c	A/S			mottled/perthitic. Wollastonite in lamellae. Strings of radiating orthoclase in albite
	plagioclase	4	f-m	A			around mottled orthoclase. 15° xtn
	pyroxene	15	m	S	yellow-green pleo		often elongate & in clusters
	sulphides	<1	f	E			assoc. + pyroxene
	titanite	<1	f	E			rare needles in amphibole
	unknown	1	f	S	colourless		2nd order biref & night-sky xtn
	wollastonite	<1	f	S			parallel xtn, phlogopite? inclns. in feld. along perthite lam
W523-2A xenolith	xenolith from coastal pipe, relatively fresh and equigranular.						
	amphibole	5	c	A/S	pleo yl-orge		rarely intergrown with mica
	garnet	3	m	A	pale pink	edges	few unidentified inclusions
	olivine	5	m-c	A		edges	fractured, isotropic
	mica	28	m-c	A/S	straw-deep orange		high relief/biref
	clino-pyx	23	m-c	A/S	v. pale green		heavily kinked
	ortho-pyx	20	m-c	A/S	orange pleo		10-15 degree xtn, high biref
	titanite	2	m	A/S	v. deep green-brown		parallel xtn, low biref
							90° cleavage, low 2V, high relief
Terskii kimberlite	from kimberlite on the Terskii coast. Heavily altered.						
	alteration	30	crypto	I			dusty alteration groundmass
	apatite	1	micro	S			small radiating clusters + phlog
	olivine	30	c	A		serp.	rounded, seriate between gm and large xenocrysts
	mica	15	f-c	S/E			elongate needles, colourless, don't cut through anything
	magnetite	20	micro	E			evenly distributed in groundmass
	"red min "	3	f	A			around mag. bright red. non-pleo

APPENDIX C

MELILITOLITE CLASSIFICATION

Root names:

1. Melilitolite is affirmed as a *general root-name* for all plutonic rocks containing > 10% modal melilite and < 50% modal primary carbonate (Woolley *et al.* 1996), as shown in Figure C.1. Any sample containing < 10% melilite is not a melilitolite and should be defined on the basis of the rest of its modal mineralogy with “melilite-bearing-” used as a prefix (Le Maitre 1989). Samples of this nature fall into the shaded area in Figure C.1.
2. If more detailed nomenclature is required, and in order to define terms used in previous studies with respect to each other, the following use of *specific root-names* is also suggested:

Any sample containing > 10% modal abundance of any of five “defining minerals” olivine, pyroxene, perovskite, nepheline, hyne, may use the specific root name associated with that mineral, as shown in Figure C.2. If any sample contains more than one of the “defining minerals” with an abundance > 10%, the specific root name is derived from the most abundant “defining” mineral, and the less abundant “defining” mineral(s) is used as a modifier(s).

3. Given the need for a specific root-name that can be used to describe samples dominantly composed of melilite, we propose that any plutonic sample with melilite > 65% should be given the specific root-name “copiamelilite”.

Modifiers :

Modifiers are normally used to denote high quantities of a particular mineral in a sample. It is suggested that any mineral with > 10% absolute modal abundance in a melilitolite should have the mineral name added as a modifier to the root name. If more

than one modifier is required, the modifiers are listed in order of increasing mineral abundance. However, the “defining” mineral used to define a specific root-name, as described above, should not be used as an additional modifier. For example, the scheme precludes the use of “nepheline turjaite”.

It is also suggested that mica and magnetite should be used as modifiers only when they reach >15% absolute modal abundance, in order to denote atypical abundances of each mineral. Many rocks which contain > 10% melilite also contain > 10% mica and/or magnetite which would lead to almost all samples using at least one of these modifiers.

Note:

It should be emphasized that the terms of rock nomenclature derived by using general and specific root-names are equivalent. For example, “nepheline pyroxene melilitolite” $\equiv$ “nepheline uncompahgrite”.

Figure C.1 Classification of rocks containing primary carbonate, melilite and other minerals, based on modal mineralogy. The division between melilitolite and carbonatite occurs at 50% carbonate, following the flow chart classification for all igneous rocks outlined in Woolley et al. (1996). The creation of the specific root name “copiamelilite” within the melilitolite field occurs at 65% melilite. The shaded area represents rocks with < 10% melilite which are not considered to be melilitolites.

Figure C.2 Flow chart to illustrate melilitolite classification (Dunworth & Bell, submitted)

APPENDIX D

ANALYTICAL METHODS: PETROLOGY

D.1 Petrography

Polished thin-sections 30 μm thick were produced from 1" x 1.75" slabs for the samples listed in Appendix A. Sections were photographed using ASA 100 Elite Ektachrome slide film with ND 10 blue filter. Slides were subsequently reprocessed to produce photographic prints.

D.2 Cathodoluminescence

A Nikon microscope with Nuclide Luminescope® ELM Model 2B was used to obtain cathodoluminescence images of the massif samples. Samples were analyzed at 12kV, 0.2 millitorr and 0.5-0.6 mA. Photographs were taken with ASA 1600 print film.

D.3 Scanning electron microscope

A Jeol 3600 scanning electron-microscope (SEM) at Carleton University was used to obtain back-scattered-electron (BSE) photographs using a slide-maker with Ektachrome 100 slide film. Additional EDS mineral composition analyses were also obtained for fine-grained dykes, and selected massif samples using an energy-dispersive system (EDS) detection operating at 15 kV with a beam current of 0.8 nA and a beam diameter of $\sim 2\text{-}5\ \mu\text{m}$. A selection of natural and synthetic standards were used.

D.4 Electron microprobe

A JEOL 733 electron microprobe at the Canadian Museum of Nature using 4 wave-dispersive system (WDS) spectrometers operating at 15 kV with a beam current of 20 nA and a beam diameter of $\sim 10\text{-}30\ \mu\text{m}$ was used to obtain the majority of the mineral analyses. Data for microprobe standards were measured for 25 s or 0.25% precision per

element, whichever was obtained first; data for the samples were measured for 25 s or to 0.5% precision, whichever was attained first. Cross-calibration checks were within analytical error. A selection of natural and synthetic standards were used.

D.5 Calculation of standard deviation values on microprobe data

D.5a Calculation of standard deviation values on oxide data

Standard deviation values on oxide values from microprobe analyses were calculated for a range of analyses taken from the majority of the major silicate and oxide phases analyzed within this study. Standard deviation values on each oxide wt.% value were calculated by taking the precision value from the raw measurement (0.5-25%) of a particular element, multiplying it by the wt.% oxide, and dividing by 100. This produces a linear relationship between standard deviation and weight % oxide, which is demonstrated graphically in Figure D.1. The minerals represented within Figure D.1 include amphibole, ankerite, apatite, dolomite, feldspar, garnet, ilmenite, magnetite, melilite, mica, perovskite and pyroxene.

D.5b Calculation of standard deviation values on cation data

The following methodology, including equations and co-existing descriptions and definitions, is taken from Numerical Methods in Petrology (Le Maitre, 1982). It should be noted that the calculated standard deviation value for each individual cation within an analysis is dependent on the errors associated with all the other cations present within the analysis, and hence a linear relationship such as that seen for weight % oxides, dependent solely on the individual weight % oxide value for a particular element, is no longer valid here. Table D.A provides an example of a garnet analysis from Turiy for which a standard deviation value has been calculated for each element present within the analysis, as well as the cation total, summarized at the bottom of the table. The full calculations for the cation total and one element (Mg) have been provided as examples of how all the

error values were calculated. It should be noted that the error associated with each individual element is greater than the value associated with the cation total.

Obviously, such calculations are long and time-consuming, and to provide the error values for every element within every microprobe analysis obtained in this study (~2000 analyses x (8 oxides + 8 cations) = 32,000 values) is impractical. The values listed in Table D.A. are provided as a guide.

Equations:

S_c = sum of the “raw” cations

x_i = weight percent of the i th oxide

a_i = number of cations in the i th oxide divided by the molecular weight of the i th oxide

$$S_c = \sum x_i a_i$$

S_o = sum of the “raw” oxygens

x_i = weight percent of the i th oxide

b_i = number of oxygens in the i th oxide divided by the molecular weight of the i th oxide

$$S_o = \sum x_i b_i$$

N_o = number of oxygens to which the analysis is normalized (e.g. 24 in garnet)

k_i = number of i th cations calculated on the basis of N_o oxygens

$$k_i = \frac{N_o x_i a_i}{S_o} = \frac{N_o x_i a_i}{\sum x_i b_i}$$

W_k = weighted sum of the cations

g_i = weighting factor for each cation

$$W_k = \sum k_i g_i = \frac{N_o \sum x_i a_i g_i}{S_o}$$

var = variance

s_i = standard deviation of the i th oxide

$$\text{var } W_k = \frac{\sum (N_o a_i g_i - b_i W_k)^2 s_i^2}{S_o}$$

In Table D.A:

$$A = N_o a_i g_i \quad B = b_i W_k \quad \text{Std.Dev.} = \frac{\sqrt{\text{Sum}(A - B)^2 s_i^2}}{S_o}$$

D.6 Data reduction

Data reduction for both SEM and electron microprobe data was carried out using a ZAF correction program, and the FORMULA program was used to normalise the analyses. Mineral formulae (listed in Appendix I) were taken from Nickel and Nichols (1991). Mica analyses where $(\text{Si}+\text{Al}) < 8$ for 22 oxygen equivalents were recalculated using the MICA program, whereby Fe^{2+} was converted to Fe^{3+} to fill the tetrahedral site, hence normalising to 8 tetrahedral cations and 22 oxygen equivalents. This calculation is iterative, and allows for vacancies in the octahedral and inter-layer sites. Both FORMULA and MICA programs were written and provided by T.S. Ercit.

Figure D.1 Standard deviation values on weight % oxides calculated by the method outlined in Table D.1. Note that several minerals, including amphibole, ankerite, apatite, dolomite, feldspar, garnet, ilmenite, magnetite, melilite, mica, perovskite and pyroxene have been used to construct this figure, in order to get a wide range of values for each of the five oxides chosen. The trend seen is essentially linear, with Na and Ca occasionally showing higher errors, presumably due to volatilization during analysis. The slope of the trend is also steeper for all elements at low absolute concentrations.

Table D.A Microprobe data error calculation: Example (garnet)

Whole rock	x_i	s_i	a_i	b_i	$x_i a_i$	$x_i b_i$	k_i	A	B	$(A-B)^2 s_i^2$
CaO	33.81	0.17	0.017832	0.017832	0.602900	0.602900	6.750736	0.427968	0.308866	0.00040996
MgO	0.54	0.017	0.024611	0.024611	0.013290	0.013290	0.148809	0.590664	0.426284	7.809E-06
MnO	0.25	0.016	0.014097	0.014097	0.003524	0.003524	0.039461	0.338328	0.244172	2.2695E-06
FeO	22.68	0.11	0.013919	0.013919	0.315683	0.315683	3.534736	0.334056	0.241089	0.00010458
Al ₂ O ₃	1.53	0.022	0.019615	0.029423	0.030011	0.045017	0.336036	0.470760	0.509632	7.3134E-07
SiO ₂	33.27	0.17	0.016643	0.033286	0.553713	1.107425	6.199980	0.399432	0.576543	0.00090654
TiO ₂	2.22	0.041	0.012516	0.025032	0.027786	0.055571	0.311117	0.300384	0.433576	2.9821E-05
				Total	1.546906	2.143410				0.00146171

total cations Std. Dev. 0.01783711

Mg only	x_i	s_i	a_i	b_i	$x_i a_i$	$x_i b_i$	k_i	A	B	$(A-B)^2 s_i^2$
CaO	33.81	0.17	0.017832	0.017832	0.602900	0.602900	6.750736	0	0.117936	0.00040197
MgO	0.54	0.017	0.024611	0.024611	0.013290	0.013290	0.148809	0.590664	0.162770	5.2914E-05
MnO	0.25	0.016	0.014097	0.014097	0.003524	0.003524	0.039461	0	0.093234	2.2253E-06
FeO	22.68	0.11	0.013919	0.013919	0.315683	0.315683	3.534736	0	0.092056	0.00010254
Al ₂ O ₃	1.53	0.022	0.019615	0.029423	0.030011	0.045017	0.336036	0	0.194596	1.8328E-05
SiO ₂	33.27	0.17	0.016643	0.033286	0.553713	1.107425	6.199980	0	0.220145	0.0014006
TiO ₂	2.22	0.041	0.012516	0.025032	0.027786	0.055571	0.311117	0	0.165555	4.6074E-05
				Total	1.546906	2.143410				0.00202465

Mg Std. Dev. 0.02099275SUMMARY

Oxide	wt. %	Std Dev	Element	cations	Std Dev
CaO	33.81	0.17	Ca	6.751	0.030
MgO	0.54	0.02	Mg	0.149	0.021
MnO	0.25	0.02	Mn	0.039	0.012
FeO	22.68	0.11	Fe	3.535	0.018
Al ₂ O ₃	1.53	0.02	Al	0.336	0.017
SiO ₂	33.27	0.17	Si	6.200	0.021
TiO ₂	2.22	0.04	Ti	0.311	0.011
Total	94.30		Total	17.321	0.018

APPENDIX E

ANALYTICAL METHODS: SR - ND ISOTOPE GEOCHEMISTRY

Table E.A Radioactive decay schemes

E.1 Sample preparation

Bulk rock analyses were carried out on powders crushed for XRF analysis, as explained in Appendix H. An average of 70 mg of sample was used for each analysis.

To obtain mineral separates from the majority of the texturally inhomogeneous samples (i.e. those for which bulk rock analyses would be meaningless), clean material was split from the drill cores using a hydraulic splitter. The material was broken into fragments using a steel crusher, and manually crushed to a powder using an agate pestle and mortar. The powder was sieved into $>250\mu\text{m}$, $[-250+149]\mu\text{m}$ and $<149\mu\text{m}$ fractions and washed alternately in ethanol and distilled water in a plastic petri dish, and dried under an infrared lamp. The $[-250+149]\mu\text{m}$ fraction was put through a Franz magnetic separator to initiate separation of the required minerals. Mineral purity was enhanced by hand-picking under a binocular microscope, and is estimated at $> 99.8\%$ for most samples. The samples were subsequently rewashed twice in ethanol and water and leached in dilute HCl. Calcite was not leached, apatite was leached in cold dilute HCl for 10 minutes, whilst silicates were leached in warm dilute HCl for up to 45 minutes.

E.2 Dissolution

Samples were accurately weighed in Sevelex beakers and dissolved according to the table below. Concentrated HCl was used to dissolve calcite, concentrated HNO_3 for apatite, and an HNO_3 -HF mixture for silicate minerals. The relative proportion of HF was decreased for minerals with high Ca contents. The dissolution mixture was refluxed

at 125°C for 48 hrs before being redissolved in conc. HNO_3 and dried down, followed by conc. HCl , then dilute HCl . The samples were then placed in an ultrasonic bath for 20 minutes to ensure complete dissolution. Any precipitate that remained was treated to a repeat dissolution procedure after most of the solution had been decanted off. The two solutions were recombined after 100% dissolution had been obtained.

Table E.B Mineral separate cleaning & leaching procedures

<u>Mineral</u>	<u>calcite</u>	<u>apatite</u>	<u>pyrox./amph.</u>	<u>phlogopite</u>
wash H_2O x2	✓	✓	✓	✓
wash $\text{C}_2\text{H}_5\text{OH}$	✓	✓	✓	✓
wash H_2O x2	✓	✓	✓	✓
leach	✗	0.5M HCl cold 10 mins	2.5M HCl warm 10-40 m.	2.5M HCl warm 10-40 m.
dry & weigh	✓	✓	✓	✓
dissolve [reflux 48 hrs]	6M HCl	6M HNO_3	1 HF : 1 HNO_3	5 HF : 1 HNO_3
6M HCl	✗	✓	✓	✓
2.5M HCl	✓	✓	✓	✓
Redissolve [if necessary]	6M HNO_3	5 HF : 1 HNO_3	5 HF : 1 HNO_3	5 HF : 1 HNO_3
6M HCl	✓	✓	✓	✓
2.5M HCl [rarely 0.5M]	✓	✓	✓	✓

E.3 Spiking

Once clear solutions in 2.5M HCl had been obtained, they were each transferred to a large disposable beaker. All samples were spiked with ^{87}Rb , ^{84}Sr , ^{145}Nd and ^{152}Sm to determine parent-daughter concentrations in the samples in order to enable isochron calculation and age-correction procedures to be carried out. Each sample solution was

accurately weighed and divided into an unspiked aliquot (Sr and Nd isotopic compositions determined only) and a spiked aliquot (Rb, Sr, Nd and Sm isotopic compositions measured). A precisely weighed aliquot of each of the four laboratory spike solutions were added to each spiked-sample solution. All solutions were subsequently dried down and redissolved in 2.5M HCl ready for ion-exchange chromatography.

The precision of the concentration values obtained by this method is $\pm 0.5\%$ for Sr, Sm and Nd, and 1% for Rb (Morisset, 1992; Simonetti, 1994).

Table E.C Composition of laboratory spike solutions

Element	Isotopic ratios			Concentration
Rb	$^{85}\text{Rb}/^{87}\text{Rb}$	=	0.008	Rb: $0.03782 \mu\text{molg}^{-1}$
Sr	$^{88}\text{Sr}/^{86}\text{Sr}$	=	3.03418	Sr : $0.01353 \mu\text{molg}^{-1}$
	$^{87}\text{Sr}/^{86}\text{Sr}$	=	0.39076	
	$^{84}\text{Sr}/^{86}\text{Sr}$	=	17.991	
Sm	$^{149}\text{Sm}/^{147}\text{Sm}$	=	266.26	Sm: $0.01190 \mu\text{molg}^{-1}$
	$^{149}\text{Sm}/^{152}\text{Sm}$	=	251.85	
Nd	$^{146}\text{Nd}/^{144}\text{Nd}$	=	1.4252	Nd: $0.09808 \mu\text{molg}^{-1}$
	$^{145}\text{Nd}/^{144}\text{Nd}$	=	35.864	
	$^{143}\text{Nd}/^{144}\text{Nd}$	=	0.26182	

E.4 Ion-exchange chromatography

Two sizes of ion-exchange columns filled with Bio-Rad AG50W-X8 cation exchange resin (200-400 mesh) were used to separate Rb, Sr, and rare earth element (REE) fractions using 2.5M and 6M HCl respectively. Sample solutions with <3mg sample were put through 10ml polypropylene columns containing 3ml resin. Larger sample sizes were loaded onto 20cm Teflon columns of 1.1 cm internal diameter

containing 13ml resin. The unspiked solution for each sample was processed prior to the spiked solution, using the same column. Columns were cleaned between runs with 6M HCl and reconditioned with 2.5M HCl. An intermediate 1.75M HCl reconditioning stage was used for the smaller columns. The full elution procedure is outlined in Table E.D.

Given the high concentrations of Ca in many of the samples, it was necessary to use an additional exchange procedure to separate Ca from both the Rb and Sr fractions. It was found that the smaller of the two sets of AG50W-X8 resin columns provided excellent Rb-Ca separation, and thus the Rb fraction from the larger columns was subsequently processed through the smaller columns where necessary, and the Rb fraction collected. A set of the smaller 10cm polypropylene columns containing 0.8ml of teflon resin coated with di-hexyl orthophosphoric acid was used to separate Ca from the Sr fraction collected from the AG50W-X8 resin columns, where necessary. The columns were cleaned between runs with 6M HCl and reconditioned with 0.01M HCl. The full elution procedure is outlined in Table F.D.

The REE fraction collected from the first set of AG50W-X8 resin columns was dried down and redissolved in 0.15M HCl. Nd and Sm separation was carried out using 20 cm Bio-Rad Vycor (quartz) columns of 6mm internal diameter containing teflon powder coated with di-hexyl orthophosphoric acid to a height of 7cm. Nd was collected in 0.3M HCl, Sm was collected in 0.5M HCl. Columns were cleaned between runs with 6M HCl and reconditioned with 0.15M HCl. The full elution procedure is outlined in Table E.D.

All final element fractions were collected in small disposable beakers, partially dried down, and then transferred to teflon microbeakers for final evaporation, ready for mass spectrometer sample loading.

Table E.D Column elution procedures

Rb-Sr (large)			Rb-Sr (small)			Ca-Sr			REE		
	<u>M</u>	<u>X</u>		<u>M</u>	<u>X</u>		<u>M</u>	<u>X</u>		<u>M</u>	<u>X</u>
Sample	2.5	2	Sample	2.5	0.5	Sample	0.01	0.25	Sample	0.15	0.2
Wash	2.5	10	Wash	2.5	4.5	Wash	0.01	1.0	Wash	0.15	0.2
Wash	2.5	30	Rb	2.5	2.0	Sr	0.01	3.0	Wash	0.15	0.2
Rb	2.5	10	Wash	2.5	4.0	Clean	6.2	30	Wash	0.3	3.0
Wash	2.5	25	Sr	2.5	4.0	Cond.	0.01	20	Nd	0.3	6.0
Sr	2.5	15	Wash	2.5	1.0				Wash	0.3	4.0
Wash	2.5	20	Wash	6.2	1.0				Wash	0.5	3.0
Wash	6.2	15	REE	6.2	5.0				Sm	0.5	8.0
REE	6.2	25	Wash	6.2	20				Clean	6.2	25
Clean	6.2	60	Clean	1.8	10				Cond.	0.15	20
Cond.	2.5	60	Cond.	2.5	10						

Key: “Sample” = sample solution loading; **Sr** = collect Sr fraction; Wash = add solution and discard eluted fraction; Cond. = condition.

M = HCl molarity; X = quantity of solution added to the column.

E.5 Blank values:

Ion-exchange column blank values were determined as follows:

Rb:	2.8 ng	[1.01x10 ⁻⁴ µmolg ⁻¹]	Nd:	0.26 ng	[4.42x10 ⁻⁷ µmolg ⁻¹]
Sr:	89 ng	[9.11x10 ⁻⁶ µmolg ⁻¹]	Sm:	0.16 ng	[1.55x10 ⁻⁷ µmolg ⁻¹]

Total procedure blanks were determined by Morisset (1992) as 0 ng (Rb), 240 ng (Sr - one measurement), 6 ng Nd and 0.2 ng Sm. Simonetti (1994) estimated blank values as ~ 0.5 ng per element. Blank corrections were not applied to the Turiy isotope results as the blank values were insignificant in comparison with the absolute trace element contents of almost all the samples analyzed.

E.6 Mass spectrometry

All samples were analyzed using a Finnigan-MAT 261 multi-collector solid source mass spectrometer operated in the static mode at Carleton University.

Table E.E Loading procedures

	<u>Rb</u>	<u>Sr</u>	<u>Nd</u>	<u>Sm</u>
Filament:	double Re	single Ta	double Re	double Re
Loading acid	dil. HCl	dil. HCl on H ₂ PO ₄	dil HCl on HNO ₃ *	dil HCl
Drying current	1.0 - 1.2 A	1.8 - 2.5 A	1.0 - 1.2 A	1.0 - 1.2 A

* nitric acid filament-etching after March 1996

E.7 Analysis

Essential analytical parameters for all runs are shown in Table F.F. Unspiked Sr and Nd samples were aborted once the fractionation factors exceeded the maxima stated below. Unspiked analyses were corrected to natural fractionation values and normalized to the internationally accepted standard values for NBS 987 and La Jolla. One standard was run in each magazine of samples. The raw data from runs of all standards and unspiked samples was recalculated to produce standard deviation values and eliminate any particularly extraneous ratio measurements. Precision is estimated as $\pm 0.003\%$ for $^{87}\text{Sr}/^{86}\text{Sr}$ and $\pm 0.004\%$ for $^{143}\text{Nd}/^{144}\text{Nd}$ (Morisset, 1992).

Table E.F Essential mass spectrometer parameters

	<u>Sr</u>	<u>Nd</u>
max. fractionation value	$^{88}\text{Sr}/^{86}\text{Sr} = 8.42$	$^{146}\text{Nd}/^{144}\text{Nd} = 0.742$
fractionation correction value	$^{88}\text{Sr}/^{86}\text{Sr} = 8.19$	$^{146}\text{Nd}/^{144}\text{Nd} = 0.719$
international standards	NBS 987 = 0.710250	La Jolla = 0.511860
Carleton standard values [average 1994-1997]	NBS 987 = 0.710267	La Jolla = 0.511849

All unspiked isotope ratio measurements were normalised to a $^{86}\text{Sr}/^{88}\text{Sr}$ value of 0.1194, and $^{146}\text{Nd}/^{144}\text{Nd}$ value of 0.719.

Element:	<u>Rb</u>	<u>Sr</u>	<u>Nd</u>	<u>Sm</u>
Analysis temp (°C)	1270 - 1320	1350 - 1600	1730 - 1830	1830 - 1900
No. blocks data	1-3	5 - 20	4 - 20	2 - 5
Isobaric interference	^{87}Sr	^{87}Rb	^{144}Sm	
Mass monitored	none	^{85}Rb	^{147}Sm	
Element interference	Ca, Sr	Rb, Ca	Ba, Ce,	Nd, Eu,

ANALYTICAL METHODS: Ar ISOTOPE GEOCHEMISTRY**F.1 Outline**

Phlogopite grains > 1mm diameter were separated from a suite of silicate rocks, phoscorites and carbonatites for ^{40}Ar - ^{39}Ar isotopic analysis. The separates were washed using nanopure water and dried under infrared light.

Neutron irradiation of the samples took place at the McMaster University research reactor, followed by a three month cooling period. The samples were due to be analyzed using single-crystal laser ablation at the University of Toronto in the spring or summer of 1997. Unfortunately, due to technical problems, the data was not available prior to the date of submission of this thesis.

ANALYTICAL METHODS: STABLE ISOTOPE GEOCHEMISTRY

G.1 Sample Preparation

10-20 mg of calcite from mineral separates with grain size $149\mu\text{m} < x < 250\mu\text{m}$ was used for C-O stable isotope determination. Whole-rock samples from a dolomite-calcite carbonatite (CDC), and the carbonatite dyke (S.W58) were also analyzed. Appendices H and E outline rock-crushing and mineral separation procedures respectively.

G.2 CO₂ extraction and analysis

Calcites were reacted at 25°C with 100% H₃PO₄ (McCrea 1950) and the dolomite at 50°C. The CO₂ produced was analyzed on a triple collector VG SIRA 12 mass spectrometer to determine the isotopic composition of the C and O. Routine precision (2σ) on a pure carbonate analysis is 0.1‰.

G.3 Standards and fractionation factors

Reference standards used:

oxygen: Standard Mean Ocean Water (SMOW)

carbon: Pee-Dee Belemnite (PDB)

$$\text{Thus } \delta^{18}\text{O}_{\text{sample}} = \left[\frac{{}^{18}\text{O}/{}^{16}\text{O}_{\text{sample}}}{{}^{18}\text{O}/{}^{16}\text{O}_{\text{reference}}} - 1 \right] \times 1000 \text{ ‰}$$

Table G.A Fractionation factors

Mineral	temp.	α	reference
calcite	25°C	1.01025	Friedman and O'Neil (1977), Sharma and Clayton (1965)
dolomite	50°C	1.01065	Rosenbaum and Sheppard (1986)
where	$\alpha = \frac{1000 + \delta X}{1000 + \delta Y}$		
	X and Y are the phases which are exchanging the isotope in question (carbonate-CO ₂).		

ANALYTICAL METHODS: MAJOR AND TRACE ELEMENT GEOCHEMISTRY

H.1 XRF

Only samples which were texturally and mineralogically homogeneous were used for bulk rock analysis. All samples were split using a hydraulic jack to obtain approximately 40 g of clean, fresh material. A steel Braun Chipmunk jaw-crusher was used to pulverise the samples to < 1cm diameter fragments, which were subsequently crushed to a fine powder in an agate shatter box. Duplicate analyses (Cousens, pers comm) indicate that contamination effects from these facilities are minimal. The use of the agate shatter box indicates that the SiO₂ values are a maximum for all samples.

Major (and trace) element data were obtained by XRF at the University of Ottawa (Table 7.A). A mixture of 64g Li₂B₄O₇ and 17.5g LiBO₂ was used a flux. 4.075g of this mixture were added to 1.300g sample and heated in a Pt crucible until molten. The fused material was poured into a Pt disc mold and analyzed on a Philips PW2400 sequential XRF spectrometer. Numerous duplicate runs and standard analyses ensure that a high degree of accuracy is obtained at all times. Precision has been determined for all elements analyzed, and is calculated from the following equation:

$$s = \underline{a}C^{\underline{b}}$$

where s = estimate of standard deviation, C = concentration, and $\underline{a}$ and $\underline{b}$ are constants specific to a certain group of elements, as listed in Table H.A. Such precision estimates are considered good for concentrations ranging from 10ppm to 70 wt%. Further information is contained in: Kretz R, Campbell JL, Joffman EL, Hartree R, Teesdale WJ: *Distribution of twenty-four trace elements among mineral phases in high-grade Grenville gneiss, amphibolite, and granulite, as determined by X-ray fluorescence, neutron activation, and proton microprobe analyses.* (unpubld. ms.)

Table H.A XRF precision calculation parameters

<u>Elements</u>	<u>a</u>	<u>b</u>	<u>C units</u>
Si, Ti, Al, Fe, Mg, Mn, Ca, K	0.010	0.664	wt %
Na	0.0046	0.664	wt %
Sr	0.15	0.67	ppm
V, Zn, Y, Zr, Nb, Th	0.086	0.67	ppm
Cr, Co, Ni, Rb, Ba	0.15	0.67	ppm

Semi-quantitative range:

< 10 ppm	Co, Ni, Zn, Rb, Y, Zr, Th, U
<20 ppm	V, Pb
<50 ppm	Ba, La, Ce, Nd

H.2 ICP-MS

Additional trace element data were obtained from the powders produced for XRF analysis (as described in a) above) by ICP-MS. Analyses were carried out by Activation Laboratories Ltd. in Ancaster, Ontario. The following description of sample preparation and analysis is taken directly from the Activation Laboratories manual:

“Samples are prepared and analyzed in a batch system. Each batch contains a method reagent blank, certified reference material and 17% replicates. Samples are mixed with a flux of lithium metaborate and lithium tetraborate and fused in an induction furnace. The molten melt is immediately poured into a solution of 5% nitric acid containing an internal standard, and mixed continuously until completely dissolved (30 minutes). For the ICP analysis, reagent blanks with and without the lithium borate flux are analyzed, as well as the method reagent blank. Interference correction verification standards are analyzed. Calibration is performed using ten prepared USGS and Canmet certified reference materials. One of the ten standards is used during the analysis for every group of ten samples. This standard brackets the group of samples. The same sample solution is also

spiked with an additional internal standard and is further diluted and introduced into a Perkin-Elmer-SCIEX ELAN 6000 ICP-MS using a proprietary sample introduction methodology. Calibration is performed using ten prepared USGS and Canmet certified reference materials.”

The accuracy and precision values for repeated runs of standard MRG 1, during the period when the Turiy samples were analyzed by Actlabs, are reported in Table H.B. below.

Table H.B Standard MRG 1: average values Sept 96 - April 97 (Actlabs)

Element	detection limit (ppm)	average value (ppm)	standard dev.	precision (RSD%)	accuracy %	certified value (ppm)
V	5	521	31.13	5.97	-0.93	526
Cr	10	443	18.26	4.12	3.15	430
Co	0.5	86.3	4.48	5.19	-0.76	87
Ni	10	185	19.89	10.73	-3.97	193
Cu	10	123	19.53	15.77	-7.58	134
Zn	10	176	22.06	12.48	-7.43	191
Ga	1	18.6	0.87	4.66	9.43	17
Ge	1	1.5	0.46	313.37	45.35	1
As	5	0	0	0	0	0.73
Rb	0.5	8.29	0.82	10.06	-3.67	8.5
Sr	0.1	269.06	10.74	3.99	1.15	266
Y	1	12.6	0.67	5.33	-10.2	14
Zr	0.5	98.18	11.96	12.18	-9.08	108
Nb	1	20.1	2.03	10.10	0.45	20
Mo	0.5	1.22	0.57	46.5	40.71	0.87
Ag	0.5	0	0	0	0	0.11
In	0.2	0	0	0	0	
Sn	1	3.3	0.62	18.99	-8.91	3.6
Sb	0.1	0.77	0.29	37.41	-10.85	0.86
Cs	0.5	0.75	0.32	42.27	31.84	0.57
Ba	1	49.2	7.64	15.52	-19.31	61
La	0.1	9.90	0.49	4.9	1.01	9.8
Ce	0.1	26.49	1.10	4.15	1.88	26
Pr	0.05	3.23	0.22	6.87	-4.99	3.4
Nd	0.1	19.31	0.94	4.87	0.57	19.2
Sm	0.1	4.48	0.21	4.78	-0.52	4.5

Table H.B (continued)

Element	detection limit (ppm)	average value (ppm)	standard dev.	precision (RSD%)	accuracy %	certified value (ppm)
Eu	0.05	1.41	0.07	4.85	4.59	1.39
Gd	0.1	3.98	0.26	6.42	-0.55	4
Tb	0.1	0.59	0.06	7.78	15.83	0.51
Dy	0.1	2.94	0.11	3.73	1.4	2.9
Ho	0.1	0.49	0.02	4.34	0.94	0.49
Er	0.1	1.25	0.06	5.09	11.67	1.12
Tm	0.05	0.13	0.01	9.75	21.22	0.11
Yb	0.1	0.78	0.20	25.38	29.61	0.6
Lu	0.04	0.12	0.01	7.76	0.63	0.12
Hf	0.2	3.75	0.21	5.52	-0.27	3.76
Ta	0.05	0.85	0.07	8.64	6.48	0.8
W	0.5	0	0	0	0	0.3
Tl	0.1	0	0	0	0	0.055
Pb	5	5.26	1.36	25.87	-47.4	10
Bi	0.2	0	0	0	0	0.13
Th	0.1	0.92	0.10	10.37	-1.36	0.93
U	0.1	0.27	0.02	6.97	13.27	0.24

APPENDIX I

MINERAL FORMULAE

List of minerals found in the Turiy rocks in this study. Mineral formulae, taken from Nickel & Nichols (1991), were used as the basis for the calculation of the mineral analyses presented in Appendix J.

amphibole:

hastingsite	$\text{NaCa}_2(\text{Fe,Mg})_5(\text{Si}_6\text{Al}_2)\text{O}_{22}(\text{OH})_2$
pargasite	$\text{NaCa}_2(\text{Mg,Fe})_4\text{Al}(\text{Si}_6\text{Al}_2)\text{O}_{22}(\text{OH})_2$
richterite	$\text{Na}_2\text{Ca}(\text{Mg,Fe})_5\text{Si}_8\text{O}_{22}(\text{OH})_2$
tremolite	$\text{Ca}_2(\text{Mg,Fe})_5\text{Si}_8\text{O}_{22}(\text{OH})_2$
analcime	$\text{Na}(\text{AlSi}_2)\text{O}_6 \cdot \text{H}_2\text{O}$
ankerite	$\text{Ca}(\text{Fe,Mg,Mn})(\text{CO}_3)_2$
apatite	$\text{Ca}_5(\text{PO}_4)_3(\text{F,OH,Cl})$
baddelyite	ZrO_2
banalsite	$\text{Na}_2\text{BaAl}_4\text{Si}_4\text{O}_{16}$
barite	BaSO_4
calcite	CaCO_3
calzirtite	$\text{Ca}_2\text{Zr}_5\text{Ti}_2\text{O}_{16}$
cancrinite	$(\text{Na,Ca})_8(\text{Si}_6\text{Al}_6)\text{O}_{24}(\text{CO}_3)_2$
celsian	$\text{BaAl}_2\text{Si}_2\text{O}_8$
chlorite	$(\text{Mg,Al,Fe,Li,Mn,Ni})_{4-6}(\text{Si,Al,B,Fe})_4\text{O}_{10}(\text{OH,O})_8$
chromite	FeCr_2O_4
copiapite	$\text{Fe}_5(\text{SO}_4)_6(\text{OH})_2 \cdot 20\text{H}_2\text{O}$
cuspidine	$\text{Ca}_4\text{Si}_2\text{O}_7(\text{F,OH})_2$
djerfisherite	$\text{K}_6\text{Na}(\text{Fe,Cu})_{24}\text{S}_{26}\text{Cl}$
dolomite	$\text{CaMg}(\text{CO}_3)_2$

feldspar:

orthoclase $(\text{K,Na})\text{AlSi}_3\text{O}_8$

plagioclase $(\text{Na,Ca})(\text{Al,Si})_4\text{O}_8$

fluorite CaF_2

garnet:

andradite $\text{Ca}_3\text{Fe}_2(\text{SiO}_4)_3$

schorlomite $\text{Ca}_3(\text{Ti,Fe})_2[(\text{Si,Fe})\text{O}_4]_3$

hillebrandite $\text{Ca}_2\text{SiO}_4 \cdot \text{H}_2\text{O}$

hibschite $\text{Ca}_3\text{Al}_2[(\text{SiO}_4)(\text{OH})_4]_3$

ilmenite FeTiO_3

katoite $\text{Ca}_3\text{Al}_2[(\text{SiO}_4)(\text{OH})_4]_3$

magnetite Fe_3O_4

melilite:

akermanite $\text{Ca}_2\text{MgSi}_2\text{O}_7$

gehlenite $\text{Ca}_2\text{Al}_2\text{SiO}_7$

soda-melilite $\text{CaNaAlSi}_2\text{O}_7$

mica :

annite $\text{K}_2\text{Fe}_6(\text{Al}_2\text{Si}_6)\text{O}_{20}(\text{OH,F})_4$

phlogopite $\text{K}_2\text{Mg}_6(\text{Al}_2\text{Si}_6)\text{O}_{20}(\text{F,OH})_4$

monticellite $\text{Ca}(\text{Mg,Fe})\text{SiO}_4$

mosandrite $(\text{Ca,Na,Ce})_{12}(\text{Ti,Zr})_2\text{Si}_7\text{O}_{25}(\text{OH})_6\text{F}_4$

natrolite $\text{Na}_2(\text{Al}_2\text{Si}_3)\text{O}_{10} \cdot 2\text{H}_2\text{O}$

nepheline $(\text{Na,K})\text{AlSiO}_4$

olivine $(\text{Mg,Fe})_2\text{SiO}_4$

pectolite $\text{NaCa}_2\text{Si}_3\text{O}_8(\text{OH})$

perovskite CaTiO_3

pyrochlore $(\text{Na,Ca})_2\text{Nb}_2\text{O}_6(\text{OH,F})$

pyroxene:

aegirine $\text{NaFeSi}_2\text{O}_6$ diopside: $\text{CaMgSi}_2\text{O}_6$ orthopyroxene $(\text{Mg,Fe})\text{SiO}_3$ quartz SiO_2 rinkite $(\text{Na,Ca})_3(\text{Ca,Ce})_4\text{Ti}_4(\text{Si}_2\text{O}_7)_2(\text{O,F})_4$ rutile TiO_2 sodalite $\text{Na}_4(\text{Si}_3\text{Al}_3)\text{O}_{12}\text{Cl}$ stronalsite $\text{Na}_2\text{SrAl}_4\text{Si}_4\text{O}_{16}$ strontianite SrCO_3

sulphides:

chalcopyrite CuFeS_2 pyrite FeS_2 tausonite SrTiO_3 titanite CaTiSiO_5 unknown $\text{Na}_{(1-x)}(\text{Ca,REE})_5\text{TiSi}_4(\text{O,F,OH})_{18}$ wollastonite CaSiO_3 zircon ZrSiO_4 zirsinalite $\text{Na}_6(\text{Ca,Mn,Fe})\text{ZrSi}_6\text{O}_{18}$

APPENDIX J MICROPROBE ANALYSES

Gaps in the data = below detection. "n.a." = not analysed.

AN SUM = anion sum, O = oxygens (or oxygen equivalents)

APPENDIX Ji ADDITIONAL MINERALS

C.AG.122				C.AG.122				C.170.38				C.IJ.14							
ijolite				ijolite				ijolite				urtite							
PT 3		PT 4		PT 5		PT 13		PT 14		PT 19		PT 40		PT 41		PT 44		PT 49	
SiO ₂	33.74	32.81	32.85	SiO ₂	25.69	22.69	32.86	SiO ₂	36.87	36.88	36.78	SiO ₂	33.49					SiO ₂	33.49
TiO ₂	11.74	9.53	10.65	TiO ₂			0.20	Al ₂ O ₃	16.01	15.24	14.74	TiO ₂	19.61					TiO ₂	19.61
Al ₂ O ₃	0.07	0.58	0.33	Al ₂ O ₃	0.22	0.18	0.88	FeO	4.10	4.85	3.01	Al ₂ O ₃	0.31					Al ₂ O ₃	0.31
Ce ₂ O ₃	2.87	2.14	2.50	FeO	4.40	4.13	4.75	MnO	0.06		0.14	BaO	15.27					BaO	15.27
MgO	0.11	1.34	0.29	MnO	0.17	0.24	0.21	MgO	2.90	2.83	4.60	SrO	2.76					SrO	2.76
FeO	0.24	1.11	0.91	MgO	6.34	5.64	8.74	SrO	0.59	0.77	0.47	FeO	7.64					FeO	7.64
SrO	0.64	0.55	0.59	SrO	1.95	1.63	1.31	CaO	35.44	35.42	36.13	MnO	1.43					MnO	1.43
CaO	36.39	38.96	38.29	CaO	37.44	40.02	32.56	Na ₂ O	0.04			MgO	0.37					MgO	0.37
Na ₂ O	3.91	4.65	4.42	Na ₂ O	0.89	0.76	0.67	H ₂ O*	2.85	3.03	3.04	CaO	6.83					CaO	6.83
K ₂ O	0.00	0.31	0.22	TOTAL	77.10	75.29	82.18	F	0.40			Na ₂ O	8.71					Na ₂ O	8.71
F	9.10	8.55	8.01					O=F	-0.17			K ₂ O	1.24					K ₂ O	1.24
O=F	-3.83	-3.60	-3.37	Si	1.439	1.330	1.629	TOTAL	99.1	99.02	98.91	TOTAL	98.11					TOTAL	98.11
TOTAL	94.98	96.93	95.69	Ti			0.007	Si	1.3971	1.4052	1.3946	Si	1.539					Si	1.539
				Al	0.015	0.012	0.051	Al	0.715	0.6844	0.6587	Ti	0.678					Ti	0.678
Si	1.399	1.354	1.375	Fe ²⁺	0.206	0.203	0.197	Mg	0.1638	0.1608	0.26	Al	0.017					Al	0.017
Ti	0.366	0.296	0.335	Mn	0.008	0.012	0.009	Fe ²⁺	0.1299	0.1545	0.0955	Ba	0.275					Ba	0.275
Al	0.003	0.028	0.016	Mg	0.529	0.493	0.646	Mn	0.0019		0.0045	Sr	0.074					Sr	0.074
Ce	0.044	0.032	0.038	Sr	0.063	0.055	0.038	Sr	0.013	0.017	0.0103	Fe ²⁺	0.294					Fe ²⁺	0.294
Mg	0.007	0.082	0.018	Ca	2.246	2.514	1.729	Ca	1.4388	1.446	1.4678	Mn	0.056					Mn	0.056
Fe ²⁺	0.008	0.038	0.032	Na	0.097	0.086	0.064	Na	0.0029			Mg	0.025					Mg	0.025
Sr	0.015	0.013	0.014	TOTAL	4.603	4.706	4.370	F	0.0479			Ca	0.336					Ca	0.336
Ca	1.616	1.723	1.717	O	6	6	6	H	0.7213	0.7692	0.7692	K	0.073					K	0.073
Na	0.314	0.372	0.359					O	5.9521	6	6	Na	0.776					Na	0.776
K		0.016	0.012					TOTAL	3.8623	3.8679	3.8915	TOTAL	4.142					TOTAL	4.142
F	1.193	1.116	1.061					AN SUM	6	6	6	O	6					O	6
TOTAL	3.773	3.956	3.917																
O	6	6	6																

N.63.4		N.64.18		N.64.18		N.64.18		N.64.19		N.64.19		
carb		mel		mel		PT 88		carb		carb		
PT 80		PT 38		PT 6		PT 12		PT 77		PT 64		
Na ₂ O	0.19	SiO ₂	35.98	SiO ₂	46.46	48.00	CaO	7.23	SiO ₂	32.46	Na ₂ O	0.24
CaO	42.21	TiO ₂	0.40	Al ₂ O ₃	5.46	4.33	MgO		TiO ₂	2.87	CaO	8.53
SrO	0.31	Al ₂ O ₃	11.06	FeO	2.04	0.56	FeO	0.22	Al ₂ O ₃	7.46	SrO	3.98
Ce ₂ O ₃	0.68	FeO	9.30	MnO	0.63	0.27	SrO	11.40	Fe ₂ O ₃	8.45	La ₂ O ₃	15.30
Nd ₂ O ₃	0.27	MgO	2.88	MgO	9.74	7.61	La ₂ O ₃	14.68	MgO	4.09	Ce ₂ O ₃	27.80
SiO ₂	19.29	SrO	0.21	CaO	23.20	26.21	Ce ₂ O ₃	25.52	SrO	0.17	Nd ₂ O ₃	8.95
P ₂ O ₅	22.81	CaO	34.73	Na ₂ O	4.55	7.29	Nd ₂ O ₃	8.36	CaO	30.18	SiO ₂	9.31
F	1.42	Na ₂ O	0.05	K ₂ O	3.98	0.84	CO ₂ *	30.20	F	0.40	P ₂ O ₅	11.64
Cl	1.81	TOTAL	94.61	TOTAL	96.06	95.11	TOTAL	97.61	O=F	-0.17	F	1.47
H ₂ O*	0.46							TOTAL	85.91	H ₂ O*	0.30	
O=F	-0.60	Si	1.685	Si	1.844	1.902	Ca	0.188	Si	1.485	O=F	-0.62
O=Cl	-0.41	Ti	0.014	Al	0.256	0.202	Mg		Ti	0.099	TOTAL	86.90
TOTAL	88.45	Al	0.538	Fe ²⁺	0.068	0.019	Fe	0.004	Al	0.402	Na	0.070
		Fe ²⁺	0.364	Mn	0.021	0.009	Sr	0.160	Fe ³⁺	0.291	Ca	1.369
Na	0.035	Mg	0.201	Mg	0.576	0.450	La	0.131	Mg	0.279	Sr	0.346
Ca	4.252	Sr	0.006	Ca	0.987	1.113	Ce	0.227	Sr	0.005	La	0.845
Sr	0.017	Ca	1.743	Na	0.351	0.560	Nd	0.072	Ca	1.480	Ce	1.524
Ce	0.023	Na	0.005	K	0.202	0.043	C	1	F	0.058	Nd	0.479
Nd	0.009	TOTAL	4.208	TOTAL	4.305	4.298	TOTAL	1.787	O ²⁻	5.942	Si	1.394
Si	1.813	O	6	O	6	6	O	3	TOTAL	4.041	P	1.476
P	1.815							AN SUM	6	F	0.696	
F	0.422									H	0.304	
Cl	0.288									O	12.304	
H	0.289									TOTAL	7.503	
O	12.289									AN SUM	13	
TOTAL	7.964											
AN SUM	13											

APPENDIX Ji ADDITIONAL MINERALS (cont)

TUR I <u>turjaite</u>					TUR II <u>turjaite</u>				C.19.163	
	PT 13	PT 14	PT 60	PT 61		PT 36	PT 37	PT 39		PT 12
SiO ₂	32.84	38.43	33.00	33.15	K	9.12	9.30	9.07	ZrO ₂	64.60
Al ₂ O ₃	0.14	1.39	4.10	0.15	Mg				HfO ₂	1.03
FeO	0.06	0.50	0.09	0.07	Fe	53.61	53.97	53.22	SiO ₂	33.03
MgO	0.05	2.39		0.11	Cu	1.28	1.09	0.92	Fe ₂ O ₃	0.11
CaO	58.63	47.50	51.14	58.62	S	31.91	31.83	31.75	CaO	0.08
SrO	0.51	0.33	0.43	0.44	Cl	1.27	1.32	1.26	TOTAL	98.85
Na ₂ O	0.15	2.62	1.16	0.15	TOTAL	97.19	97.51	96.22		
K ₂ O	0.05	0.11	0.28	0.06					Zr	0.970
TOTAL	92.43	93.27	90.20	92.75	K	6.029	6.125	6.048	Hf	0.009
					Mg				Si	1.017
Si	2.040	2.298	2.072	2.051	Fe	24.806	24.895	24.840	Fe ³⁺	0.003
Al	0.010	0.098	0.303	0.011	Cu	0.518	0.440	0.375	Ca	0.003
Fe ²⁺	0.003	0.025	0.005	0.004	S	25.718	25.582	25.812	TOTAL	2.002
Mg	0.005	0.213		0.010	Cl	0.928	0.959	0.925	O	4
Ca	3.902	3.043	3.440	3.886	atoms	58.000	58.000	58.000		
Sr	0.018	0.011	0.016	0.016						
Na	0.018	0.304	0.141	0.018	djerfisherite					
K	0.004	0.008	0.022	0.005						
TOTAL	6.000	6.000	6.000	6.000						
O	8.034	8.191	8.142	8.045						

zircon

C.23.300	<u>carbonatite</u>		
	PT 84	PT 85	PT 86
ZrO ₂	66.42	67.09	66.29
TiO ₂	15.76	15.41	15.70
CaO	11.91	11.71	11.71
TOTAL	94.09	94.21	93.70
Zr	5.119	5.175	5.130
Ti	1.873	1.833	1.874
Ca	2.017	1.985	1.991
TOTAL	9.008	8.992	8.996
O	16	16	16

calzirtite

C.100.y	<u>carbonatite</u>			
	PT 87	PT 88	PT 89	PT 90
UO ₂	5.63			
Nb ₂ O ₅	43.34	37.84	31.32	36.54
Ta ₂ O ₅	5.94	13.10	10.66	15.61
TiO ₂	7.83	4.85	6.59	5.60
SiO ₂	4.13	6.85	7.29	6.24
Ce ₂ O ₃	2.04	2.10	2.82	1.74
FeO	1.89	2.65	2.98	2.16
CaO	4.16	5.60	4.95	4.99
K ₂ O	0.71	0.28	0.36	0.47
Na ₂ O	0.42	0.33	0.33	0.35
H ₂ O *	1.87	1.90	1.78	1.88
TOTAL	77.96	75.50	69.08	75.58
U	0.098			
Nb	1.571	1.347	1.195	1.317
Ta	0.130	0.281	0.245	0.338
Ti	0.472	0.287	0.418	0.336
Si	0.331	0.539	0.615	0.497
Ce	0.060	0.061	0.087	0.051
Fe ²⁺	0.127	0.175	0.210	0.144
Ca	0.357	0.473	0.448	0.426
K	0.073	0.028	0.039	0.048
Na	0.065	0.050	0.054	0.054
H	1.000	1.000	1.000	1.000
TOTAL	3.283	3.241	3.310	3.210
AN SUM	7	7	7	7

TQ	<u>turjite</u>
	PT 86
SiO ₂	43.94
Al ₂ O ₃	38.76
FeO	0.62
MgO	0.38
BaO	0.17
CaO	0.12
Na ₂ O	0.17
K ₂ O	11.33
TOTAL	95.49
Si	1.597
Al	1.660
Fe ²⁺	0.019
Mg	0.021
Ba	0.002
Ca	0.005
Na	0.012
K	0.525
TOTAL	3.841
O	6

APPENDIX Ji ADDITIONAL MINERALS (cont)

TD 2 Melt inclns in apatite			TD2			ijolite-fenite				
	PT 60	PT 61		PT 41		PT 51	PT 52	PT 53	PT 82	PT 62
SiO ₂	47.91	46.1	SiO ₂	50.68	SiO ₂	31.18	31.64	31.25	32.23	47.62
TiO ₂	2.16	2.12	TiO ₂	0.32	TiO ₂	9.93	10.83	10.28	10.57	0.66
Al ₂ O ₃	1.69	1.47	Al ₂ O ₃	0.62	Al ₂ O ₃		0.11	0.11	0.05	
FeO	29.97	29.48	FeO	0.20	FeO	0.18	0.38	0.29	0.18	5.18
MnO	0.28	0.25	MgO	7.55	MnO				0.07	1.40
MgO	3.12	2.95	CaO	17.37	BaO		0.40	0.24		0.13
SrO	0.57	0.57	Na ₂ O	3.85	SrO	1.45	1.46	1.46	1.42	5.44
CaO	3.14	2.42	TOTAL	80.59	CaO	39.23	35.42	37.57	42.86	11.88
Na ₂ O	0.29	0.49			Na ₂ O	5.50	4.11	4.87	5.15	11.60
K ₂ O	0.64	0.98	Si	2.224	K ₂ O					0.71
P ₂ O ₅	0.18	0.09	Ti	0.011	H ₂ O *	4.01	4.07	4.02	4.14	6.12
TOTAL	89.95	86.93	Al	0.032	F	12.08	9.15	10.76	10.03	0.53
			Fe ²⁺	0.007	O=F	-5.09	-3.85	-4.53	-4.22	-0.22
Si	2.102	2.104	Mg	0.494	TOTAL	98.47	93.71	96.31	102.48	91.05
Ti	0.071	0.073	Ca	0.817						
Al	0.088	0.079	Na	0.328	Si	7.000	7.000	7.000	7.000	7.000
Fe ²⁺	1.100	1.125	TOTAL	3.913	Ti	1.676	1.802	1.732	1.726	0.073
Mn	0.010	0.010	O	6	Al		0.029	0.029	0.013	
Mg	0.204	0.201			Fe ²⁺	0.034	0.070	0.054	0.033	0.637
Sr	0.015	0.015			Mn				0.013	0.174
Ca	0.148	0.118			Ba		0.035	0.021		0.007
Na	0.025	0.043			Sr	0.189	0.187	0.190	0.179	0.464
K	0.036	0.057			Ca	9.436	8.396	9.017	9.974	1.871
P	0.007	0.004			Na	2.394	1.763	2.115	2.169	3.306
TOTAL	3.804	3.829			K					0.133
O	6	6			F	8.577	6.402	7.623	6.890	0.246
					H	6.000	6.000	6.000	6.000	6.000
					O	26.920	27.015	27.035	28.309	21.896
					TOTAL	20.729	19.282	20.158	21.106	13.666
					AN SUM	35.497	33.418	34.658	35.199	22.142

APPENDIX Jii AMPHIBOLE

C.AG.347 pyroxenite

	pt58	pt62	pt66	pt73	pt75	pt79	pt86	pt89
SiO ₂	39.98	39.47	41.29	41.28	40.24	41.08	39.75	39.72
TiO ₂	2.18	1.88	0.31	1.33	2.22	0.62	2.72	2.65
Al ₂ O ₃	13.62	14.24	13.28	12.87	13.66	12.76	13.64	13.65
Fe ₂ O ₃	3.09	4.77	6.32	4.46	3.63	6.26	3.19	3.15
FeO	4.64	3.2	2.25	3.23	4.31	2.74	4.52	4.43
MnO	0.08	0.08	0.10		0.14	0.09		0.06
MgO	16.31	16.33	17.08	17.02	16.38	16.79	16.22	16.30
SrO								
BaO								
CaO	12.68	12.74	13.00	12.69	12.63	12.80	12.65	12.56
Na ₂ O	2.88	3.02	3.42	3.29	3.00	3.37	2.69	2.69
K ₂ O	0.85	0.80	0.27	0.61	0.89	0.41	1.35	1.31
H ₂ O*	2.02	2.02	2.04	2.03	2.03	2.02	2.02	2.02
F								
O=F								
TOTAL	98.33	98.55	99.36	98.81	99.13	98.94	98.75	98.54
Si	5.895	5.800	5.992	6.025	5.887	6.004	5.850	5.854
^{iv} Al	2.105	2.200	2.008	1.975	2.113	1.996	2.150	2.146
^{vi} Al	0.262	0.266	0.263	0.239	0.242	0.202	0.216	0.225
Ti	0.343	0.528	0.690	0.490	0.400	0.689	0.354	0.349
Fe ³⁺	0.242	0.208	0.034	0.146	0.244	0.068	0.301	0.294
Mg	3.585	3.578	3.695	3.703	3.572	3.658	3.558	3.581
Fe ²⁺	0.572	0.394	0.273	0.394	0.528	0.336	0.556	0.546
Mn	0.010	0.010	0.012		0.017	0.011		0.007
Sr								
Ba								
Ca	2.003	2.006	2.021	1.984	1.980	2.004	1.995	1.983
Na	0.823	0.860	0.962	0.931	0.851	0.955	0.768	0.769
K	0.160	0.150	0.050	0.114	0.166	0.076	0.253	0.246
F								
H	2	2	2	2	2	2	2	2
O	22	22	22	22	22	22	22	22
TOTAL	16.000	16.000	16.000	16.001	16.000	15.999	16.001	16.000
AN SUM	23	23	23	23	23	23	23	23

C.AG.48a ol melt

incln in ol

SiO ₂	41.56
TiO ₂	2.54
Al ₂ O ₃	11.36
Fe ₂ O ₃	
FeO	6.88
MnO	0.08
MgO	16.22
SrO	
BaO	
CaO	12.04
Na ₂ O	3.05
K ₂ O	1.19
H ₂ O*	2.00
F	
O=F	
TOTAL	96.92
Si	6.222
Al ^{iv}	1.778
Al ^{vi}	0.226
Ti	
Fe ³⁺	0.286
Mg	3.620
Fe ²⁺	0.861
Mn	0.010
Sr	
Ba	
Ca	1.931
Na	0.885
K	0.227
F	
H	2
O	22
TOTAL	16.046
AN SUM	23

C.K.7 phoscorite

	pt33	pt36	pt45	pt47	pt48	pt51
SiO ₂	55.86	55.23	54.16	55.62	55.93	55.99
TiO ₂	0.60	0.58	1.03	0.51	0.65	0.63
Al ₂ O ₃	1.38	1.55	1.55	1.12	0.88	0.78
Fe ₂ O ₃						
FeO	2.51	2.59	3.18	2.35	3.39	2.90
MnO	0.08	0.08	0.12	0.05	11.00	0.09
MgO	22.43	22.27	21.75	22.54	21.83	22.12
SrO						
BaO						
CaO	7.85	7.98	7.96	7.56	7.24	7.04
Na ₂ O	6.38	6.27	5.72	6.42	6.38	6.41
K ₂ O	0.55	0.55	1.20	0.53	0.69	0.77
H ₂ O*	2.16	2.14	2.12	2.14	2.14	2.14
F						
O=F						
TOTAL	99.80	99.24	98.79	98.84	99.24	98.87
Si	7.758	7.723	7.659	7.788	7.832	7.851
^{iv} Al	0.226	0.255	0.258	0.185	0.148	0.129
^{vi} Al						
Ti	0.062	0.061	0.109	0.054	0.069	0.066
Fe ³⁺						
Mg	4.644	4.642	4.585	4.705	4.557	4.624
Fe ²⁺	0.292	0.303	0.376	0.275	0.397	0.430
Mn	0.009	0.009	0.014		0.013	0.011
Sr						
Ba						
Ca	1.168	1.196	1.206	1.134	1.086	1.058
Na	1.718	1.699	1.569	1.643	1.732	1.743
K	0.097	0.098	0.216	0.095	0.123	0.138
F						
H	2	2	2	2	2	2
O	22	22	22	22	22	22
TOTAL	15.974	15.986	15.992	15.879	15.957	16.050
AN SUM	23	23	23	23	23	23

C.90.29 pyroxenite

	pt 38	pt 39	pt 41	pt 42	pt 44
SiO ₂	39.41	40.21	39.15	39.38	38.97
TiO ₂	1.78	1.78	2.20	2.06	2.12
Al ₂ O ₃	12.75	12.03	12.85	12.72	12.72
Fe ₂ O ₃					
FeO	9.49	9.14	9.24	9.42	9.54
MnO	0.13	0.13	0.08	0.10	0.13
MgO	14.85	15.30	14.77	14.61	14.38
SrO	0.21	0.16	0.21	0.21	0.22
BaO	0.13			0.18	
CaO	12.45	12.41	12.27	12.14	12.20
Na ₂ O	2.33	2.39	2.35	2.38	2.34
K ₂ O	1.76	1.89	1.87	1.87	1.89
H ₂ O*	1.44	1.41	1.39	1.36	1.51
F	1.12	1.19	1.22	1.27	0.94
O=F	-0.47	-0.50	-0.51	-0.53	-0.40
TOTAL	97.38	97.54	97.09	97.17	96.56
Si	6.005	6.095	5.973	6.014	5.988
^{iv} Al	1.995	1.905	2.027	1.986	2.012
^{vi} Al	0.295	0.244	0.283	0.304	0.291
Ti	0.204	0.203	0.252	0.237	0.245
Fe ³⁺					
Mg	3.373	3.457	3.359	3.326	3.294
Fe ²⁺	1.209	1.158	1.179	1.203	1.226
Mn	0.017	0.017	0.010	0.000	0.017
Sr	0.018	0.014	0.018	0.018	0.019
Ba	0.008			0.011	
Ca	2.033	2.017	2.006	1.987	2.009
Na	0.688	0.702	0.695	0.705	0.697
K	0.342	0.365	0.364	0.364	0.370
F					
H	2	2	2	2	2
O	22	22	22	22	22
TOTAL	16.187	16.177	16.166	16.155	16.168
AN SUM	23	23	23	23	23

C.3.93 phoscorite

	pt16	pt19
SiO ₂	55.82	55.53
TiO ₂	0.36	0.58
Al ₂ O ₃	0.77	0.18
Fe ₂ O ₃		
FeO	5.05	3.37
MnO	0.10	0.08
MgO	20.88	22.01
SrO		
BaO		
CaO	5.51	6.17
Na ₂ O	7.53	6.67
K ₂ O	0.76	1.55
H ₂ O*	1.46	1.48
F	1.40	1.34
O=F	-0.59	-0.56
TOTAL	99.05	98.39
Si	7.893	7.883
^{iv} Al	0.107	0.030
^{vi} Al	0.021	
Ti	0.038	0.062
Fe ³⁺		
Mg	4.401	4.658
Fe ²⁺	0.598	0.400
Mn	0.012	0.010
Sr		
Ba		
Ca	0.835	0.938
Na	2.065	1.836
K	0.137	0.281
F		
H	2	2
O	22	22
TOTAL	16.107	16.098
AN SUM	23	23

APPENDIX Jii AMPHIBOLE (cont.)

C.TL.344 <u>carbonatite</u>					
	pt57	pt60	pt67	pt70	pt82
SiO ₂	55.04	54.79	55.26	56.33	55.97
TiO ₂	0.73	0.63	0.76	0.48	0.61
Al ₂ O ₃	1.24	1.41	1.16	0.88	1.01
Fe ₂ O ₃					
FeO	2.32	2.32	2.28	1.96	2.11
MnO	0.06	0.06		0.06	
MgO	22.62	22.29	22.64	22.76	22.75
SrO					
BaO					
CaO	7.79	7.78	7.46	7.28	7.53
Na ₂ O	6.50	6.41	6.53	6.73	6.75
K ₂ O	0.41	0.39	0.43	0.42	0.38
H ₂ O*	2.14	2.13	2.14	2.15	2.15
F					
O=F					
TOTAL	98.85	98.31	98.66	99.05	99.26
Si	7.719	7.725	7.751	7.845	7.794
²⁷ Al	0.205	0.234	0.192	0.144	0.166
²⁹ Al					
Ti	0.077	0.067	0.080	0.051	0.064
Fe ³⁺					
Mg	4.729	4.685	4.734	4.725	4.722
Fe ²⁺	0.272	0.274	0.267	0.228	0.246
Mn		0.007			
Sr					
Ba					
Ca	1.171	1.175	1.121	1.086	1.123
Na	1.767	1.780	1.776	1.817	1.823
K	0.073	0.070	0.077	0.075	0.068
F					
H	2	2	2	2	2
O	22	22	22	22	22
TOTAL	16.013	16.017	15.998	15.971	16.006
AN SUM	23	23	23	23	23

C.18.60 <u>carbonatite</u>						
	PT 4	PT 5	PT 7	PT 11	PT 12	PT 14
SiO ₂	55.44	52.04	52.30	53.90	53.73	51.26
TiO ₂	0.37	0.47	0.55	0.48	0.41	0.63
Al ₂ O ₃	1.12	3.15	2.45	1.83	1.59	3.97
Fe ₂ O ₃						
FeO	6.90	8.86	10.87	7.60	8.78	9.64
MnO	0.07	0.15	0.16	0.13	0.16	0.12
MgO	19.57	17.82	16.66	18.58	18.23	17.32
SrO						
BaO						
CaO	4.53	6.66	5.10	5.40	5.18	5.65
Na ₂ O	7.85	6.60	7.40	7.15	7.42	7.20
K ₂ O	0.76	0.72	0.70	0.84	0.68	0.71
H ₂ O*	1.50	1.64	1.61	1.59	1.57	1.65
F	1.28	0.89	0.92	1.02	1.06	0.87
O=F	-0.54	-0.37	-0.39	-0.43	-0.45	-0.37
TOTAL	98.85	98.63	98.33	98.09	98.36	98.65
Si	7.902	7.554	7.666	7.788	7.786	7.464
²⁷ Al	0.098	0.446	0.334	0.212	0.214	0.536
²⁹ Al	0.090	0.093	0.089	0.100	0.058	0.145
Ti	0.040	0.051	0.061	0.052	0.045	0.069
Fe ³⁺						
Mg	4.158	3.856	3.641	4.002	3.938	3.760
Fe ²⁺	0.822	1.076	1.333	0.918	1.064	1.174
Mn	0.008	0.018	0.020		0.020	0.015
Sr						
Ba						
Ca	0.692	1.036	0.801	0.836	0.804	0.881
Na	2.169	1.857	2.103	2.003	2.085	2.033
K	0.138	0.133	0.131	0.155	0.126	0.132
F	0.602	0.427	0.446	0.486	0.507	0.418
H	1.423	1.591	1.573	1.534	1.514	1.599
O	23.423	23.591	23.573	23.534	23.514	23.599
TOTAL	16.117	16.120	16.179	16.066	16.140	16.209
AN SUM	23	23	23	23	23	23

C.XC <u>carbonatite</u>					
	PT 16	PT 17	PT 23	PT 28	PT 3
SiO ₂	54.15	52.8	52.43	51.77	52.09
TiO ₂	0.49	0.8	0.37	0.34	0.38
Al ₂ O ₃	1.15	2.03	1.12	1.19	0.88
Fe ₂ O ₃					
FeO	4.47	5.18	4.33	10.58	10.86
MnO	0.11	0.13	0.11	0.11	0.07
MgO	20.84	20.19	20.77	16.97	16.41
SrO		0.08			
BaO					
CaO	6.76	7.41	6.84	2.39	2.22
Na ₂ O	6.47	6.03	6.48	8.76	9.09
K ₂ O	0.58	0.64	0.44	0.54	0.41
H ₂ O*	1.8	1.76	1.75	1.71	1.77
F	0.59	0.66	0.59	0.57	0.42
O=F	-0.25	-0.28	-0.25	-0.24	-0.18
TOTAL	97.17	97.43	94.99	94.69	94.43
Si	7.791	7.629	7.732	7.844	7.915
²⁷ Al	0.195	0.346	0.195	0.156	0.085
²⁹ Al				0.057	0.073
Ti	0.053	0.087	0.041	0.039	0.043
Fe ³⁺					
Mg	4.470	4.349	4.566	3.833	3.717
Fe ²⁺	0.538	0.626	0.534	1.341	1.380
Mn	0.013	0.016		0.014	
Sr		0.007			
Ba					
Ca	1.042	1.147	1.081	0.388	0.361
Na	1.805	1.689	1.853	2.574	2.678
K	0.106	0.118	0.083	0.104	0.079
F	0.2797	0.3151	0.287	0.2849	0.2108
H	1.732	1.698	1.725	1.727	1.798
O	23.732	23.698	23.725	23.727	23.798
TOTAL	16.013	16.014	16.085	16.350	16.331
AN SUM	23	23	23	23	23

C.100.Y <u>carbonatite</u>					
	PT 24	PT 30	PT 31	PT 37	PT 38
SiO ₂	55.49	52.60	55.70	52.36	56.04
TiO ₂	0.36	0.51	0.28	0.77	0.25
Al ₂ O ₃	0.77	2.59	0.52	2.81	0.53
Fe ₂ O ₃					
FeO	8.50	12.42	9.00	10.03	8.13
MnO			0.05	0.09	
MgO	18.57	15.59	18.23	17.07	18.74
SrO					
BaO					
CaO	3.58	2.18	2.68	4.73	3.05
Na ₂ O	7.91	9.08	8.44	7.65	8.53
K ₂ O	1.26	0.41	1.32	0.53	1.07
H ₂ O*	1.46	1.52	1.48	1.68	1.64
F	1.31	1.07	1.27	0.79	0.94
O=F	-0.55	-0.45	-0.53	-0.33	-0.40
TOTAL	98.66	97.52	98.43	98.18	98.53
Si	7.978	7.776	8.038	7.642	8.038
²⁷ Al	0.022	0.224		0.358	
²⁹ Al	0.108	0.227	0.088	0.125	0.090
Ti	0.039	0.057	0.030	0.085	0.027
Fe ³⁺					
Mg	3.980	3.436	3.922	3.714	4.007
Fe ²⁺	1.022	1.536	1.086	1.224	0.975
Mn			0.006	0.011	
Sr					
Ba					
Ca	0.551	0.345	0.414	0.740	0.469
Na	2.205	2.603	2.361	2.165	2.372
K	0.231	0.077	0.243	0.099	0.196
F	0.622	0.522	0.605	0.381	0.445
H	1.404	1.500	1.420	1.635	1.574
O	23.404	23.500	23.420	23.635	23.574
TOTAL	16.136	16.281	16.188	16.163	16.174
AN SUM	23	23	23	23	23

NC.153.82 <u>fenite</u>		PT 54
SiO ₂		41.36
TiO ₂		2.43
Al ₂ O ₃		10.35
Fe ₂ O ₃		
FeO		15.69
MnO		0.26
MgO		10.70
SrO		
BaO		
CaO		10.75
Na ₂ O		3.33
K ₂ O		1.76
H ₂ O*		1.39
F		1.18
O=F		-0.50
TOTAL		98.71
Si		6.349
²⁷ Al		1.651
²⁹ Al		0.221
Ti		0.281
Fe ³⁺		
Mg		2.449
Fe ²⁺		2.014
Mn		
Sr		
Ba		
Ca		1.768
Na		0.991
K		0.345
F		0.598
H		1.427
O		23.427
TOTAL		16.069
AN SUM		23

APPENDIX Jiii   ANALCIME + NATROLITE

analcime

TU 120A	<u>turjite</u>		NC.153.82	<u>fenite</u>		
	PT 10	PT 16		PT 56	PT 57	PT 73
SiO ₂	52.86	53.63	SiO ₂	59.64	58.49	56.83
Al ₂ O ₃	23.45	22.64	Al ₂ O ₃	23.57	24.05	25.03
FeO	0.13	0.19	FeO	0.11	0.16	0.19
SrO			SrO	0.68	0.63	0.23
CaO	0.14	0.15	CaO	0.78	0.52	0.28
Na ₂ O	13.04	12.69	Na ₂ O	11.89	12.52	13.40
K ₂ O	0.11		K ₂ O	2.09	2.72	2.74
TOTAL	89.73	89.30	TOTAL	98.76	99.09	98.70
Si	1.980	2.012	Si	2.043	2.012	1.968
Al	1.036	1.001	Al	0.952	0.975	1.022
Fe ²⁺	0.004	0.006	Fe ²⁺	0.003	0.005	0.006
Sr			Sr	0.014	0.013	0.005
Ca	0.006	0.006	Ca	0.029	0.019	0.010
Na	0.947	0.923	Na	0.790	0.835	0.900
K	0.005		K	0.091	0.119	0.121
TOTAL	3.973	3.949	TOTAL	3.921	3.978	4.031
O	6	6	O	6	6	6

natrolite

C.54.x

carbonatite

	PT 84	PT 85	PT 1	PT 2	PT 3	PT 4	PT 5
SiO ₂	46.32	46.91	46.14	46.13	46.43	46.61	46.99
Al ₂ O ₃	26.32	26.72	26.88	26.72	26.46	26.77	26.91
SrO	0.09	0.09					
CaO						0.14	
Na ₂ O	10.02	9.88	15.75	15.70	16.02	16.41	16.27
TOTAL	82.81	83.60	88.77	88.55	88.91	89.93	90.17
Si	3.108	3.113	2.975	2.982	2.992	2.977	2.987
Al	2.087	2.090	2.043	2.035	2.010	2.015	2.017
Sr	0.003	0.003					
Ca						0.0067	
Na	1.303	1.272	1.970	1.967	2.002	2.032	2.005
TOTAL	6.502	6.478	6.988	6.983	7.003	7.033	7.008
O	10	10	10	10	10	10	10

ijolite-fenite + C.190.x

	PT 9	PT 13	PT 81	PT 59	PT 69	PT 1	PT 2
SiO ₂	46.05	45.76	45.74	45.96	46.51	46.03	46.06
Al ₂ O ₃	26.38	26.34	26.02	26.18	25.89	25.58	25.33
Fe ₂ O ₃	0.10						0.23
SrO							
Na ₂ O	13.38	12.89	14.84	14.02	14.42	15.91	16.02
TOTAL	85.91	84.99	86.60	86.16	86.82	87.52	87.62
Si	3.033	2.997	2.900	2.947	2.953	3.013	3.016
Al	2.048	2.033	1.943	1.978	1.938	1.973	1.955
Fe ³⁺	0.005						0.010
Sr							
Na	1.708	1.637	1.823	1.743	1.775	2.018	2.035
TOTAL	6.795	6.667	6.667	6.668	6.667	7.005	7.017
O	10	10	10	10	10	10	10

APPENDIX J iv   APATITE

C.AG.347				pyroxenite			C.AG.48A					olivine melteigite			C.AG.122			ijolite					
	PT 55	PT 64	PT 77		PT 42	PT 51	PT 52	PT 68	PT 87		PT 47	PT 51											
Na ₂ O	0.17	0.06	0.12									0.05	Na ₂ O										
CaO	56.23	53.84	55.33		53.83	53.04	53.46	53.12	54.17		53.86	53.52	CaO										
SrO	0.32	0.44	0.30		1.85	1.14	1.62	1.54	1.01		2.10	2.39	SrO										
La ₂ O ₃					0.23	0.51	0.24	0.55	0.30		0.32	0.26	La ₂ O ₃										
Ce ₂ O ₃					0.16	0.95	0.38	1.01	0.69		0.63	0.37	Ce ₂ O ₃										
Nd ₂ O ₃						0.31							Nd ₂ O ₃										
SiO ₂	1.20	2.87	1.21		0.55	0.97	0.70	0.86	0.72		0.47	0.39	SiO ₂										
P ₂ O ₅	40.51	39.78	40.61		41.48	40.74	40.87	40.62	41.38		41.01	41.80	P ₂ O ₅										
F	1.30	1.92	2.13		3.19	2.41	4.80	3.00	3.49		3.82	2.97	F										
Cl													Cl										
H ₂ O*	1.17	0.86	0.76		0.26	0.62	-0.52	0.33	0.12		-0.05	0.37	H ₂ O*										
O=F	-0.55	-0.81	-0.90		-1.34	-1.01	-2.02	-1.26	-1.47		-1.61	-1.25	O=F										
O=Cl													O=Cl										
TOTAL	100.35	98.97	99.57		100.21	99.67	99.53	99.77	100.41		100.56	100.87	TOTAL										
Na	0.028	0.010	0.020									0.008	Na										
Ca	5.063	4.874	5.009		4.876	4.845	4.889	4.861	4.896		4.901	4.831	Ca										
Sr	0.016	0.022	0.015		0.091	0.056	0.080	0.076	0.049		0.103	0.117	Sr										
La					0.007	0.016	0.008	0.017	0.009		0.010	0.008	La										
Ce					0.005	0.030	0.012	0.032	0.021		0.020	0.011	Ce										
Nd						0.009							Nd										
Si	0.101	0.243	0.102		0.046	0.083	0.060	0.073	0.061		0.040	0.033	Si										
P	2.882	2.846	2.905		2.969	2.940	2.953	2.937	2.955		2.949	2.981	P										
F	0.346	0.513	0.569		0.853	0.650	1.296	0.810	0.931		1.026	0.791	F										
Cl													Cl										
H	0.654	0.487	0.431		0.147	0.350	-0.296	0.190	0.069		-0.026	0.209	H										
O	12.654	12.487	12.431		12.147	12.350	11.704	12.190	12.069		11.974	12.209	O										
TOTAL	8.090	7.994	8.050		7.994	7.979	8.001	7.997	7.991		8.022	7.990	TOTAL										
AN SUM	13	13	13		13	13	13	13	13		13	13	AN SUM										
C.ND.500				ijolite				C.90.y				ijolite				C.IJ.14				urtite			
	PT 1	PT 7	PT 13	PT 22		PT 1	PT 2	PT 26			PT 3	PT 6	PT 26	PT 38	PT 39	PT 46	PT 47						
Na ₂ O	0.07	0.06	0.08	0.05				0.15		Na ₂ O	0.08		0.06		0.07	0.05	0.07						
CaO	55.83	55.47	54.37	55.90		54.53	54.28	55.47		CaO	52.38	51.72	51.81	53.47	53.53	53.17	55.08						
SrO	0.69	0.65	0.58	0.79		1.47	1.45	0.88		SrO	4.10	5.28	4.96	2.40	1.82	2.95	1.30						
La ₂ O ₃		0.19				0.30	0.30	0.19		La ₂ O ₃	0.33	0.16	0.45	0.59	0.75	0.50							
Ce ₂ O ₃		0.22	0.21			0.53	0.74	0.20		Ce ₂ O ₃	0.49	0.24	0.57	0.97	1.17	0.71							
Nd ₂ O ₃										Nd ₂ O ₃					0.29								
SiO ₂	0.79	0.74	1.12	0.65		0.53	0.64	1.29		SiO ₂	0.43		0.34	0.82	1.09	0.76	0.69						
P ₂ O ₅	40.77	41.40	40.73	41.49		41.40	40.85	40.17		P ₂ O ₅	41.42	41.47	40.82	40.85	40.24	41.24	41.54						
F	1.85	2.64	1.49	2.40		4.59	4.20	2.86		F	1.23	1.84	1.04	1.22	1.54	2.08	1.51						
Cl										Cl													
H ₂ O*	0.90	0.54	1.06	0.66		-0.40	-0.22	0.42		H ₂ O*	1.19	0.88	1.26	1.19	1.03	0.79	1.07						
O=F	-0.78	-1.11	-0.63	-1.01		-1.93	-1.77	-1.20		O=F	-0.52	-0.77	-0.44	-0.51	-0.65	-0.88	-0.64						
O=Cl										O=Cl													
TOTAL	100.12	100.80	99.01	100.92		101.02	100.47	100.42		TOTAL	101.13	100.82	100.87	101.00	100.88	101.38	100.63						
Na	0.011	0.010	0.013	0.008				0.025		Na	0.013		0.010		0.012	0.008	0.011						
Ca	5.047	4.979	4.946	5.007		4.921	4.935	5.023		Ca	4.754	4.729	4.750	4.850	4.876	4.805	4.946						
Sr	0.034	0.032	0.029	0.038		0.072	0.072	0.043		Sr	0.201	0.261	0.246	0.118	0.090	0.144	0.063						
La		0.006				0.010	0.010	0.006		La	0.010	0.005	0.014	0.018	0.024	0.016							
Ce		0.007	0.007			0.017	0.023	0.006		Ce	0.015	0.007	0.018	0.030	0.036	0.022							
Nd										Nd					0.009								
Si	0.067	0.062	0.095	0.054		0.045	0.055	0.109		Si	0.036	0.000	0.029	0.069	0.093	0.064	0.058						
P	2.912	2.936	2.928	2.937		2.952	2.935	2.874		P	2.971	2.996	2.957	2.928	2.896	2.945	2.948						
F	0.494	0.700	0.400	0.635		1.223	1.127	0.765		F	0.330	0.497	0.281	0.327	0.414	0.555	0.400						
Cl										Cl													
H	0.506	0.300	0.600	0.365		-0.223	-0.127	0.236		H	0.670	0.503	0.719	0.673	0.586	0.445	0.600						
O	12.506	12.300	12.600	12.365		11.778	11.873	12.236		O	12.670	12.503	12.719	12.673	12.586	12.445	12.600						
TOTAL	8.071	8.032	8.017	8.045		8.015	8.028	8.086		TOTAL	8.001	7.999	8.024	8.014	8.035	8.004	8.026						
AN SUM	13	13	13	13		13	13	13		AN SUM	13	13	13	13	13	13	13						

APPENDIX J iv   APATITE (cont.)

E.123.106 "ijolite"					WKM melteigite					C.5.118 melilitolite				
	PT 13	PT 17	PT 21	PT 26		PT 60	PT 72	PT 73	PT 74		PT 1	PT 19	PT 24	
Na ₂ O	0.05			0.08	Na ₂ O					Na ₂ O	0.05			
CaO	51.55	52.15	52.14	51.38	CaO	55.45	54.72	54.53	53.80	CaO	55.47	56.12	55.78	
SrO	4.68	2.99	2.49	4.33	SrO	0.72	0.82	1.27	0.62	SrO	0.58	0.61	0.56	
La ₂ O ₃	0.49	0.63	0.98	0.56	La ₂ O ₃	0.27	0.44	0.37	0.67	La ₂ O ₃	1.28	0.24		
Ce ₂ O ₃	0.71	0.97	1.64	0.65	Ce ₂ O ₃	0.57	0.79	0.77	0.95	Ce ₂ O ₃			0.18	
Nd ₂ O ₃			0.21		Nd ₂ O ₃					Nd ₂ O ₃				
SiO ₂	0.45	1.12	1.33	0.27	SiO ₂	0.70	0.71	0.70	1.11	SiO ₂	0.99	1.05	1.15	
P ₂ O ₅	39.48	39.56	38.91	39.83	P ₂ O ₅	40.64	39.95	39.19	38.65	P ₂ O ₅	40.85	40.75	40.58	
F	3.65	3.01	3.40	3.32	F	2.89	2.97	3.29	3.12	F	3.01	3.35	2.97	
Cl					Cl					Cl	3.30			
H ₂ O*	-0.01	0.31	0.11	0.14	H ₂ O*	0.40	0.34	0.17	0.23	H ₂ O*	0.28	0.20	0.37	
O=F	-1.54	-1.27	-1.43	-1.40	O=F	-1.22	-1.25	-1.39	-1.31	O=F	-1.27	-1.41	-1.25	
O=Cl					O=Cl					O=Cl	-0.74			
TOTAL	99.51	99.47	99.78	99.16	TOTAL	100.42	99.49	98.90	97.84	TOTAL	101.48	100.91	100.34	
Na	0.009			0.014	Na					Na	0.008			
Ca	4.824	4.835	4.854	4.807	Ca	5.028	5.028	5.067	5.042	Ca	4.987	5.046	5.035	
Sr	0.237	0.150	0.126	0.220	Sr	0.036	0.041	0.064	0.032	Sr	0.028	0.030	0.027	
La	0.016	0.020	0.032	0.018	La	0.009	0.014	0.012	0.022	La	0.040	0.007		
Ce	0.023	0.031	0.052	0.021	Ce	0.018	0.025	0.025	0.031	Ce			0.006	
Nd			0.007		Nd					Nd				
Si	0.040	0.097	0.116	0.024	Si	0.059	0.061	0.061	0.097	Si	0.083	0.088	0.097	
P	2.920	2.898	2.862	2.945	P	2.912	2.901	2.878	2.862	P	2.902	2.895	2.894	
F	1.009	0.824	0.934	0.917	F	0.774	0.806	0.903	0.863	F	0.799	0.889	0.791	
Cl					Cl					Cl	0.043			
H	-0.009	0.177	0.066	0.083	H	0.227	0.195	0.098	0.137	H	0.159	0.111	0.209	
O	11.992	12.177	12.066	12.083	O	12.227	12.195	12.098	12.137	O	12.159	12.111	12.209	
TOTAL	8.067	8.031	8.047	8.047	TOTAL	8.060	8.069	8.105	8.084	TOTAL	8.048	8.066	8.059	
AN SUM	13	13	13	13	AN SUM	13	13	13	13	AN SUM	13	13	13	

E.133.30 melilitolite					S.78.15 melilitolite					WKT turjaite					TUR 1 turjaite				
	PT 30	PT 36	PT 55		PT 37	PT 44		PT 8	PT 26	PT 29		PT 16	PT 28	PT 30	PT 36	PT 43			
Na ₂ O									0.06		Na ₂ O								
CaO	53.54	54.66	54.21		54.77	55.43		55.89	55.86	55.38	CaO	53.15	53.90	53.69	54.17	53.39			
SrO	1.00	1.02	0.98		0.69	0.66		0.56	0.52	0.58	SrO	1.13	0.69	0.84	0.87	0.96			
La ₂ O ₃	0.40	0.52	0.40								La ₂ O ₃	0.53	0.23	0.28	0.35	0.70			
Ce ₂ O ₃	0.93	0.81	0.80			0.15					Ce ₂ O ₃	0.92	0.43	0.64	0.65	1.25			
Nd ₂ O ₃											Nd ₂ O ₃		0.25	0.30		0.36			
SiO ₂	1.24	1.23	1.20		0.64	0.70		0.70	0.83	0.96	SiO ₂	1.30	1.06	1.12	1.35	1.73			
P ₂ O ₅	40.18	40.26	40.32		39.26	40.66		40.15	40.39	40.18	P ₂ O ₅	40.50	40.48	40.14	39.51	38.99			
F	2.50	2.06	2.35		2.38	2.09		2.84	3.03	2.58	F	3.49	3.51	2.13	2.35	2.17			
Cl											Cl								
H ₂ O*	0.57	0.79	0.65		0.59	0.78		0.41	0.33	0.54	H ₂ O*	0.10	0.09	0.74	0.63	0.71			
O=F	-1.05	-0.87	-0.99		-1.00	-0.88		-1.20	-1.28	-1.09	O=F	-1.47	-1.48	-0.90	-0.99	-0.91			
O=Cl											O=Cl								
TOTAL	99.31	100.49	99.92		97.33	99.59		99.36	99.74	99.13	TOTAL	99.65	99.16	98.98	98.89	99.35			
Na									0.010		Na								
Ca	4.905	4.961	4.938		5.114	5.040		5.107	5.077	5.058	Ca	4.852	4.929	4.929	4.989	4.931			
Sr	0.050	0.050	0.049		0.035	0.033		0.028	0.026	0.029	Sr	0.056	0.034	0.042	0.043	0.048			
La	0.013	0.016	0.013								La	0.017	0.007	0.009	0.011	0.022			
Ce	0.029	0.025	0.025			0.005					Ce	0.029	0.013	0.020	0.020	0.039			
Nd											Nd		0.008	0.009		0.011			
Si	0.106	0.104	0.102		0.056	0.060		0.060	0.071	0.082	Si	0.111	0.090	0.096	0.116	0.149			
P	2.909	2.888	2.902		2.896	2.921		2.899	2.901	2.900	P	2.921	2.925	2.912	2.875	2.845			
F	0.676	0.552	0.632		0.656	0.561		0.766	0.813	0.696	F	0.940	0.948	0.577	0.639	0.592			
Cl											Cl								
H	0.324	0.448	0.368		0.344	0.439		0.234	0.187	0.305	H	0.060	0.052	0.423	0.361	0.408			
O	12.324	12.448	12.368		12.344	12.439		12.234	12.187	12.305	O	12.060	12.052	12.423	12.361	12.408			
TOTAL	8.011	8.044	8.027		8.100	8.057		8.093	8.084	8.069	TOTAL	7.986	8.006	8.017	8.054	8.045			
AN SUM	13	13	13		13	13		13	13	13	AN SUM	13	13	13	13	13			

APPENDIX J iv   APATITE (cont.)

C.K.7 <u>phoscorite</u>			C.19.163 <u>phoscorite</u>						C.18.26 <u>phoscorite</u>			
	PT 49	PT 52		PT 6	PT 11	PT 19	PT 22	PT 25		PT 25	PT 30	PT 37
Na ₂ O		0.28	Na ₂ O	0.11	0.09	0.15	0.07	0.10	Na ₂ O	0.17	0.13	0.17
CaO	54.68	54.61	CaO	55.39	55.02	54.97	55.64	55.32	CaO	55.73	56.37	56.74
SrO	0.28	0.34	SrO	0.37	0.62	0.39	0.42	0.32	SrO	0.41	0.31	0.34
La ₂ O ₃			La ₂ O ₃						La ₂ O ₃			
Ce ₂ O ₃			Ce ₂ O ₃			0.14			Ce ₂ O ₃	0.24		0.16
Nd ₂ O ₃			Nd ₂ O ₃						Nd ₂ O ₃			
SiO ₂	0.30	0.75	SiO ₂	0.59	0.52	0.29	0.24	0.65	SiO ₂	0.61	0.38	
P ₂ O ₅	42.35	42.10	P ₂ O ₅	41.70	42.26	42.58	42.28	42.24	P ₂ O ₅	40.41	41.31	42.77
F	2.56	2.07	F	3.25	3.45	2.81	0.00	2.67	F	1.70	1.77	1.83
Cl			Cl				0.23		Cl			
H ₂ O*	0.57	0.81	H ₂ O*	0.25	0.16	0.47	1.74	0.54	H ₂ O*	0.96	0.95	0.95
O=F	-1.08	-0.87	O=F	-1.37	-1.45	-1.18		-1.12	O=F	-0.72	-0.75	-0.77
O=Cl			O=Cl				-0.05		O=Cl			
TOTAL	99.67	100.09	TOTAL	100.29	100.67	100.61	100.57	100.71	TOTAL	99.51	100.47	102.19
Na		0.045	Na	0.018	0.015	0.024	0.011	0.016	Na	0.028	0.021	0.027
Ca	4.916	4.889	Ca	4.975	4.916	4.907	4.971	4.931	Ca	5.078	5.071	5.007
Sr	0.014	0.016	Sr	0.018	0.030	0.019	0.020	0.015	Sr	0.020	0.015	0.016
La			La						La			
Ce			Ce			0.004			Ce	0.008		0.005
Nd			Nd						Nd			
Si	0.025	0.063	Si	0.049	0.043	0.024	0.020	0.054	Si	0.052	0.032	
P	3.008	2.978	P	2.960	2.984	3.003	2.985	2.975	P	2.909	2.936	2.983
F	0.679	0.547	F	0.862	0.910	0.740		0.703	F	0.457	0.470	0.477
Cl			Cl				0.033		Cl			
H	0.321	0.453	H	0.138	0.090	0.260	0.967	0.297	H	0.543	0.530	0.524
O	12.321	12.453	O	12.138	12.090	12.260	12.967	12.297	O	12.543	12.530	12.524
TOTAL	7.963	7.992	TOTAL	8.020	7.988	7.981	8.008	7.992	TOTAL	8.095	8.075	8.038
AN SUM	13	13	AN SUM	13	13	13	13	13	AN SUM	13	13	13

C.TL.344 <u>carbonatite</u>					C.6.15 <u>carbonatite</u>				C.18.60 <u>carbonatite</u>			C.23.270b <u>carbonatite</u>		
	PT 59	PT 61	PT 62	PT 69		PT 3	PT 8	PT 21		PT 1	PT 6		PT 51	PT 52
Na ₂ O	0.13	0.13	0.12	0.13	Na ₂ O	0.16	0.10	0.09	Na ₂ O	0.12	0.07	Na ₂ O	0.12	0.10
CaO	55.11	54.51	54.88	55.31	CaO	56.40	56.63	55.97	CaO	55.96	56.20	CaO	55.50	55.97
SrO	0.27	0.21	0.25	0.24	SrO	0.32	0.43	0.42	SrO	0.67	0.49	SrO	0.25	0.31
La ₂ O ₃					La ₂ O ₃				La ₂ O ₃			La ₂ O ₃		
Ce ₂ O ₃					Ce ₂ O ₃				Ce ₂ O ₃	0.20		Ce ₂ O ₃		
Nd ₂ O ₃					Nd ₂ O ₃				Nd ₂ O ₃			Nd ₂ O ₃		
SiO ₂	0.61	0.60	0.56	0.25	SiO ₂				SiO ₂	0.10		SiO ₂	0.09	
P ₂ O ₅	42.09	42.04	42.05	42.62	P ₂ O ₅	41.37	41.80	42.50	P ₂ O ₅	42.64	42.85	P ₂ O ₅	41.48	41.15
F	2.58	2.35	1.95	2.77	F	1.89	2.30	2.65	F	3.07	2.33	F	1.31	1.01
Cl					Cl				Cl			Cl		
H ₂ O*	0.57	0.67	0.86	0.49	H ₂ O*	0.88	0.70	0.55	H ₂ O*	0.36	0.71	H ₂ O*	1.15	1.29
O=F	-1.09	-0.99	-0.82	-1.17	O=F	-0.80	-0.97	-1.12	O=F	-1.29	-0.98	O=F	-0.55	-0.43
O=Cl					O=Cl				O=Cl			O=Cl		
TOTAL	100.27	99.52	99.85	100.64	TOTAL	100.23	100.99	101.06	TOTAL	101.82	101.67	TOTAL	99.35	99.40
Na	0.021	0.021	0.019	0.021	Na	0.026	0.016	0.015	Na	0.020	0.011	Na	0.020	0.017
Ca	4.933	4.905	4.927	4.931	Ca	5.093	5.074	4.989	Ca	4.963	4.976	Ca	5.033	5.088
Sr	0.013	0.010	0.012	0.012	Sr	0.016	0.021	0.021	Sr	0.032	0.024	Sr	0.013	0.016
La					La				La			La		
Ce					Ce				Ce	0.006		Ce		
Nd					Nd				Nd			Nd		
Si	0.051	0.050	0.047	0.021	Si				Si	0.009		Si	0.008	
P	2.977	2.989	2.983	3.002	P	2.952	2.959	2.994	P	2.988	2.998	P	2.972	2.956
F	0.682	0.624	0.517	0.729	F	0.504	0.609	0.698	F	0.804	0.609	F	0.351	0.271
Cl					Cl				Cl			Cl		
H	0.318	0.376	0.483	0.271	H	0.497	0.392	0.303	H	0.197	0.391	H	0.650	0.729
O	12.318	12.376	12.483	12.271	O	12.497	12.392	12.303	O	12.197	12.391	O	12.650	12.729
TOTAL	7.994	7.976	7.988	7.986	TOTAL	8.086	8.070	8.017	TOTAL	8.017	8.009	TOTAL	8.044	8.075
AN SUM	13	13	13	13	AN SUM	13	13	13	AN SUM	13	13	AN SUM	13	13

APPENDIX J iv   APATITE (cont.)

C.23.300 <u>carbonatite</u>			C.49.40 <u>carbonatite</u>					C.54.x <u>carbonatite</u>					
	PT 52	PT 55		PT 35	PT 47	PT 50	PT 51		PT 62	PT 63	PT 65	PT 66	PT 74
Na ₂ O	0.13	0.14	Na ₂ O	0.14	0.24	0.18	0.18	Na ₂ O		2.49	2.33		
CaO	56.16	56.23	CaO	56.20	56.19	56.45	55.28	CaO	53.00	50.91	50.87	54.27	52.74
SrO	0.33	0.32	SrO	0.46	0.42	0.47	0.51	SrO	0.73	2.22	2.37	1.09	1.03
La ₂ O ₃			La ₂ O ₃					La ₂ O ₃	1.23			0.49	1.28
Ce ₂ O ₃	0.14		Ce ₂ O ₃					Ce ₂ O ₃	2.07		0.21	0.92	1.82
Nd ₂ O ₃			Nd ₂ O ₃					Nd ₂ O ₃	0.45				0.43
SiO ₂	0.34	0.28	SiO ₂		0.06	0.09	0.05	SiO ₂	1.98			1.02	1.79
P ₂ O ₅	42.07	41.93	P ₂ O ₅	42.53	42.43	42.45	42.40	P ₂ O ₅	38.58	34.69	35.01	40.67	39.37
F	1.23	1.40	F	2.22	1.82	2.36	2.02	F	1.44	1.72	2.30	2.17	1.30
Cl			Cl					Cl					
H ₂ O*	1.22	1.13	H ₂ O*	0.75	0.94	0.69	0.84	H ₂ O*	1.05	0.76	0.50	0.74	1.13
O=F	-0.52	-0.59	O=F	-0.93	-0.77	-0.99	-0.85	O=F	-0.61	-0.72	-0.97	-0.91	-0.55
O=Cl			O=Cl					O=Cl					
TOTAL	101.10	100.84	TOTAL	101.37	101.34	101.70	100.43	TOTAL	99.93	92.07	92.62	100.46	100.35
Na	0.021	0.023	Na	0.023	0.039	0.029	0.029	Na		0.458	0.427		
Ca	5.005	5.025	Ca	4.997	4.997	5.008	4.951	Ca	4.899	5.178	5.148	4.922	4.840
Sr	0.016	0.016	Sr	0.022	0.020	0.023	0.025	Sr	0.037	0.122	0.130	0.054	0.051
La			La					La	0.039			0.015	0.040
Ce	0.005		Ce					Ce	0.065		0.007	0.029	0.057
Nd			Nd					Nd	0.014				0.013
Si	0.029	0.024	Si	0.000	0.005	0.007	0.004	Si	0.171	0.000	0.000	0.086	0.153
P	2.963	2.961	P	2.988	2.981	2.976	3.001	P	2.818	2.788	2.799	2.914	2.855
F	0.324	0.370	F	0.583	0.478	0.618	0.534	F	0.393	0.516	0.687	0.581	0.352
Cl			Cl					Cl					
H	0.677	0.631	H	0.417	0.522	0.382	0.466	H	0.607	0.484	0.313	0.419	0.648
O	12.677	12.631	O	12.417	12.522	12.382	12.466	O	12.607	12.484	12.313	12.419	12.648
TOTAL	8.037	8.047	TOTAL	8.029	8.042	8.043	8.010	TOTAL	8.043	8.547	8.511	8.020	8.009
AN SUM	13	13	AN SUM	13	13	13	13	AN SUM	13	13	13	13	13

C.100.a <u>carbonatite</u>					C.100.b <u>carbonatite</u>				C.100.d <u>carbonatite</u>				
	PT 59	PT 60	PT 64	PT 68		PT 7	PT 8	PT 10		PT 17	PT 22	PT 28	PT 30
Na ₂ O	0.15	0.11	0.09	0.10	Na ₂ O	0.10	0.09	0.13	Na ₂ O	0.19	0.38	0.15	0.17
CaO	55.41	55.99	55.90	55.85	CaO	55.43	56.45	56.18	CaO	56.00	56.26	56.42	56.09
SrO	0.46	0.41	0.42	0.44	SrO	0.39	0.51	0.35	SrO	0.31	0.32	0.35	0.32
La ₂ O ₃					La ₂ O ₃				La ₂ O ₃				
Ce ₂ O ₃					Ce ₂ O ₃		0.19		Ce ₂ O ₃				
Nd ₂ O ₃					Nd ₂ O ₃				Nd ₂ O ₃				
SiO ₂		0.12	0.08	0.04	SiO ₂				SiO ₂				
P ₂ O ₅	42.19	42.09	42.67	42.88	P ₂ O ₅	41.44	41.36	42.01	P ₂ O ₅	42.30	41.98	42.28	42.39
F	2.77	2.91	2.84	2.99	F	3.06	2.77	2.87	F	2.03	1.97	1.81	1.59
Cl					Cl				Cl				
H ₂ O*	0.47	0.42	0.46	0.39	H ₂ O*	0.32	0.47	0.43	H ₂ O*	0.84	0.86	0.94	1.05
O=F	-1.17	-1.23	-1.20	-1.26	O=F	-1.29	-1.17	-1.21	O=F	-0.85	-0.83	-0.76	-0.67
O=Cl					O=Cl				O=Cl				
TOTAL	100.29	100.82	101.26	101.43	TOTAL	99.45	100.67	100.76	TOTAL	100.81	100.94	101.19	100.94
Na	0.025	0.018	0.015	0.016	Na	0.017	0.015	0.021	Na	0.031	0.062	0.024	0.028
Ca	4.978	5.011	4.968	4.953	Ca	5.036	5.091	5.035	Ca	5.004	5.034	5.028	5.003
Sr	0.023	0.020	0.020	0.021	Sr	0.019	0.025	0.017	Sr	0.015	0.016	0.017	0.016
La					La				La				
Ce					Ce		0.006		Ce				
Nd					Nd				Nd				
Si		0.010	0.007	0.004	Si				Si				
P	2.995	2.976	2.997	3.005	P	2.975	2.948	2.975	P	2.987	2.968	2.977	2.988
F	0.735	0.769	0.745	0.783	F	0.821	0.738	0.760	F	0.536	0.521	0.476	0.419
Cl					Cl				Cl				
H	0.266	0.232	0.255	0.218	H	0.180	0.263	0.241	H	0.465	0.480	0.524	0.582
O	12.266	12.232	12.255	12.218	O	12.180	12.263	12.241	O	12.465	12.480	12.524	12.582
TOTAL	8.020	8.035	8.006	7.998	TOTAL	8.046	8.084	8.048	TOTAL	8.036	8.079	8.047	8.033
AN SUM	13	13	13	13	AN SUM	13	13	13	AN SUM	13	13	13	13

APPENDIX J iv   APATITE (cont.)

C.100.y				C.XC				N.64.19					
carbonatite				carbonatite				carbonatite					
	PT 20	PT 27	PT 39		PT 6	PT 19	PT 25	PT 27		PT 59	PT 65	PT 67	PT 73
Na ₂ O	0.13	0.11	0.08	Na ₂ O	0.09	0.23	0.14	0.20	Na ₂ O	0.10	0.09	0.10	0.09
CaO	55.42	56.47	56.40	CaO	55.29	55.26	55.37	55.65	CaO	53.33	51.13	53.02	53.45
SrO	0.52	0.52	0.63	SrO	0.32	0.31	0.32	0.33	SrO	0.34	0.42	0.36	0.40
La ₂ O ₃				La ₂ O ₃					La ₂ O ₃	0.46	1.13	0.44	0.42
Ce ₂ O ₃	0.25	0.16		Ce ₂ O ₃					Ce ₂ O ₃	1.06	2.64	1.10	1.08
Nd ₂ O ₃				Nd ₂ O ₃					Nd ₂ O ₃	0.44	1.17	0.47	0.47
SiO ₂				SiO ₂		0.18	0.18	0.26	SiO ₂	2.22	2.49	1.92	1.93
P ₂ O ₅	42.52	42.68	43.22	P ₂ O ₅	41.88	41.12	41.55	41.40	P ₂ O ₅	38.85	38.34	38.76	38.63
F	2.71	2.65	2.10	F	1.74	2.13	1.83	1.93	F	1.59	2.15	1.58	2.01
Cl				Cl					Cl				
H ₂ O*	0.51	0.56	0.83	H ₂ O*	0.95	0.75	0.91	0.86	H ₂ O*	0.99	0.71	0.98	0.78
O=F	-1.14	-1.12	-0.88	O=F	-0.73	-0.90	-0.77	-0.81	O=F	-0.67	-0.91	-0.67	-0.85
O=Cl				O=Cl					O=Cl				
TOTAL	100.92	102.03	102.38	TOTAL	99.54	99.09	99.53	99.82	TOTAL	98.71	99.36	98.06	98.41
Na	0.021	0.018	0.013	Na	0.015	0.038	0.023	0.033	Na	0.017	0.015	0.017	0.015
Ca	4.951	4.998	4.958	Ca	4.998	5.035	5.013	5.031	Ca	4.920	4.757	4.928	4.961
Sr	0.025	0.025	0.030	Sr	0.016	0.016	0.016	0.016	Sr	0.017	0.021	0.018	0.020
La				La					La	0.015	0.036	0.014	0.013
Ce	0.008	0.005		Ce					Ce	0.033	0.084	0.035	0.034
Nd				Nd					Nd	0.014	0.036	0.015	0.015
Si				Si		0.016	0.015	0.022	Si	0.191	0.216	0.167	0.167
P	3.001	2.985	3.002	P	2.992	2.960	2.972	2.957	P	2.832	2.819	2.847	2.833
F	0.715	0.693	0.545	F	0.465	0.573	0.489	0.515	F	0.433	0.591	0.434	0.551
Cl				Cl					Cl				
H	0.286	0.308	0.455	H	0.536	0.427	0.511	0.485	H	0.567	0.409	0.566	0.449
O	12.286	12.308	12.455	O	12.536	12.427	12.511	12.485	O	12.567	12.409	12.566	12.449
TOTAL	8.005	8.030	8.003	TOTAL	8.020	8.064	8.038	8.059	TOTAL	8.038	7.985	8.040	8.059
AN SUM	13	13	13	AN SUM	13	13	13	13	AN SUM	13	13	13	13

N.63.4			N.187.100				TU 120A			TQ			
carbonatite			carbonatite				turjite			turjite			
	PT 75	PT 78		PT 10	PT 16	PT 18	PT 22		PT 13	PT 18		PT 32	PT 50
Na ₂ O	0.07	0.07	Na ₂ O			0.07	0.06	Na ₂ O				0.25	0.11
CaO	53.81	51.25	CaO	52.50	53.60	52.52	51.97	CaO	52.62	53.71		54.76	54.73
SrO	0.36	0.41	SrO	0.64	0.46	0.50	0.33	SrO	2.95	1.23		0.89	0.81
La ₂ O ₃	0.38	5.24	La ₂ O ₃	0.51	0.37	0.47	0.30	La ₂ O ₃	0.35			0.31	0.34
Ce ₂ O ₃	0.98		Ce ₂ O ₃	0.90	0.79	0.88	0.56	Ce ₂ O ₃	0.36			0.30	0.39
Nd ₂ O ₃	0.42	0.38	Nd ₂ O ₃					Nd ₂ O ₃					
SiO ₂	1.30	1.41	SiO ₂	1.22	1.11	1.18	1.26	SiO ₂	0.36	0.44		2.15	1.56
P ₂ O ₅	34.82	33.52	P ₂ O ₅	35.94	36.54	35.34	35.59	P ₂ O ₅	41.12	41.33		37.90	39.80
F	1.92	2.33	F	2.62	2.59	2.36	2.30	F				2.46	2.90
Cl		0.63	Cl					Cl					
H ₂ O*	0.71	0.32	H ₂ O*	0.39	0.43	0.49	0.52	H ₂ O*				0.56	0.39
O=F	-0.81	-0.98	O=F	-1.10	-1.09	-0.99	-0.97	O=F				-1.04	-1.22
O=Cl		-0.14	O=Cl					O=Cl					
TOTAL	93.96	94.44	TOTAL	93.61	94.80	92.82	91.92	TOTAL	97.76	96.93		98.55	99.81
Na	0.013	0.013	Na			0.013	0.011	Na				0.042	0.018
Ca	5.330	5.197	Ca	5.174	5.206	5.229	5.185	Ca	4.820	4.900		5.083	4.987
Sr	0.019	0.022	Sr	0.034	0.024	0.027	0.018	Sr	0.146	0.061		0.045	0.040
La	0.013	0.183	La	0.018	0.013	0.016	0.011	La	0.011			0.010	0.011
Ce	0.033		Ce	0.031	0.026	0.030	0.019	Ce	0.011			0.010	0.012
Nd	0.014	0.013	Nd					Nd					
Si	0.120	0.133	Si	0.112	0.101	0.110	0.118	Si	0.031	0.038		0.187	0.133
P	2.725	2.686	P	2.799	2.805	2.780	2.806	P	2.976	2.979		2.780	2.866
F	0.561	0.697	F	0.762	0.743	0.694	0.678	F				0.674	0.780
Cl		0.101	Cl					Cl					
H	0.439	0.202	H	0.238	0.258	0.307	0.323	H				0.326	0.220
O	12.439	12.202	O	12.238	12.258	12.307	12.323	O	13.000	13.000		12.326	12.220
TOTAL	8.268	8.247	TOTAL	8.166	8.174	8.204	8.166	TOTAL	7.994	7.978		8.156	8.067
AN SUM	13	13	AN SUM	13	13	13	13	AN SUM	13	13		13	13

APPENDIX J iv   APATITE (cont.)

TD1	carbonate-fluorite vein			TD2	aegirine-apatite vein					ij/fen	ijolite-fenite contact			
	PT 51	PT 52	PT 54		PT 33	PT 38	PT 39	PT 42	PT 46		PT 42	PT 70	PT 72	PT 83
Na ₂ O	0.05			Na ₂ O	0.04				0.05	Na ₂ O		0.05		
CaO	55.80	56.34	56.12	CaO	55.59	55.29	55.04	54.57	55.51	CaO	50.17	48.95	50.59	50.69
SrO	1.06	1.02	1.01	SrO	1.09	1.10	1.47	1.63	0.99	SrO	6.62	8.08	6.66	7.19
La ₂ O ₃			0.20	La ₂ O ₃						La ₂ O ₃	0.65	0.54	0.36	0.19
Ce ₂ O ₃			0.27	Ce ₂ O ₃	0.15			0.14	0.17	Ce ₂ O ₃	1.06	0.71	0.69	0.39
Nd ₂ O ₃				Nd ₂ O ₃						Nd ₂ O ₃				
SiO ₂			0.26	SiO ₂	0.26	0.12			0.21	SiO ₂	0.20			
P ₂ O ₅	41.62	41.82	41.41	P ₂ O ₅	40.80	40.48	41.22	41.38	41.22	P ₂ O ₅	39.77	39.69	39.65	39.89
F	4.67	4.19	4.21	F	3.41	4.08	3.94	4.07	3.80	F	3.94	4.03	3.99	3.94
Cl				Cl						Cl				
H ₂ O*	-0.43	-0.19	-0.21	H ₂ O*	0.15	-0.19	-0.10	-0.17	-0.03	H ₂ O*	-0.15	-0.21	-0.18	-0.15
O=F	-1.97	-1.76	-1.77	O=F	-1.44	-1.72	-1.66	-1.71	-1.60	O=F	-1.66	-1.70	-1.68	-1.66
O=Cl				O=Cl						O=Cl				
TOTAL	100.80	101.41	101.50	TOTAL	100.05	99.17	99.91	99.91	100.32	TOTAL	100.60	100.14	100.08	100.48
Na	0.008			Na	0.007				0.008	Na		0.009		
Ca	5.031	5.049	5.040	Ca	5.059	5.079	5.013	4.970	5.028	Ca	4.695	4.621	4.753	4.740
Sr	0.052	0.050	0.049	Sr	0.054	0.055	0.073	0.081	0.049	Sr	0.336	0.413	0.339	0.364
La			0.006	La						La	0.021	0.018	0.012	0.006
Ce			0.009	Ce	0.005			0.005	0.006	Ce	0.034	0.023	0.022	0.013
Nd				Nd						Nd				
Si			0.022	Si	0.022	0.011			0.018	Si	0.018			
P	2.965	2.961	2.939	P	2.934	2.939	2.966	2.978	2.951	P	2.941	2.961	2.943	2.948
F	1.243	1.108	1.116	F	0.916	1.107	1.059	1.094	1.016	F	1.089	1.123	1.107	1.088
Cl				Cl						Cl				
H	-0.243	-0.108	-0.116	H	0.084	-0.107	-0.059	-0.094	-0.016	H	-0.089	-0.123	-0.107	-0.088
O	11.757	11.892	11.884	O	12.084	11.894	11.941	11.906	11.984	O	11.912	11.877	11.894	11.913
TOTAL	8.057	8.059	8.064	TOTAL	8.079	8.083	8.051	8.032	8.059	TOTAL	8.044	8.043	8.069	8.070
AN SUM	13	13	13	AN SUM	13	13	13	13	13	AN SUM	13	13	13	13

APPENDIX Jv**BANALSITE + STRONALSITE****C.AG.122****ijolite**

	PT 6	PT 9	PT 49
SiO ₂	35.05	36.01	35.55
TiO ₂	0.08	0.09	0.09
Al ₂ O ₃	29.23	30.01	29.57
FeO	0.58	1.33	0.48
BaO	20.83	10.53	17.60
SrO	0.73	1.31	1.82
CaO		1.37	
Na ₂ O	9.15	11.18	9.44
K ₂ O	0.12	2.67	0.47
TOTAL	95.77	94.50	95.02
Si	4.012	3.968	4.025
Ti	0.007	0.007	0.008
Al	3.943	3.897	3.946
Fe ²⁺	0.056	0.123	0.045
Ba	0.934	0.455	0.781
Sr	0.048	0.084	0.119
Ca		0.162	
Na	2.031	2.388	2.072
K	0.018	0.375	0.068
TOTAL	11.045	11.458	11.064
O	16	16	16

C.IJ.14**urtite**

	PT 51	PT 55	PT 56
SiO ₂	37.02	37.41	37.15
TiO ₂	0.14	0.10	0.11
Al ₂ O ₃	30.15	30.31	30.59
FeO	0.46	0.23	0.33
BaO	22.86	22.27	22.02
SrO	0.14	0.48	0.56
CaO		0.16	0.07
Na ₂ O	9.35	9.26	9.29
K ₂ O	0.10		0.12
TOTAL	100.22	100.22	100.24
Si	4.053	4.075	4.045
Ti	0.011	0.008	0.008
Al	3.891	3.891	3.925
Fe ²⁺	0.043	0.021	0.029
Ba	0.981	0.949	0.939
Sr	0.008	0.029	0.035
Ca		0.019	0.008
Na	1.984	1.955	1.963
K	0.013		0.016
TOTAL	10.989	10.949	10.971
O	16	16	16

E.133.30 turjaite

	PT 60
SiO ₂	38.66
TiO ₂	
Al ₂ O ₃	32.02
Fe ₂ O ₃	0.37
MgO	0.05
SrO	16.49
CaO	0.51
Na ₂ O	10.00
K ₂ O	
TOTAL	98.10
Si	4.011
Ti	
Al	3.915
Fe ³⁺	0.029
Mg	0.008
Sr	0.992
Ca	0.056
Na	2.011
K	
TOTAL	11.021
O	16

APPENDIX Jvi CANCRINITE

C.AG.48A			C.AG.122			C.ND.500				C.170.38			
<u>ol.melt</u>			<u>ijolite</u>			<u>ijolite</u>				<u>ijolite</u>			
	PT 26	PT 67		PT 37	PT 38	PT 40		PT 2	PT 3	PT 4	PT 26		PT 47
SiO ₂	34.95	41.87	SiO ₂	35.02	33.70	33.57	SiO ₂	36.86	37.20	36.63	34.95	SiO ₂	34.07
Al ₂ O ₃	28.27	30.51	Al ₂ O ₃	29.29	28.30	28.86	Al ₂ O ₃	29.72	31.57	31.22	28.27	Al ₂ O ₃	30.87
FeO	0.79	0.50	FeO	0.51	0.50	0.12	FeO	0.60	0.41	0.48	0.79	FeO	0.12
MgO	0.06	0.05	MgO	0.04	0.07		MgO		0.21	0.23	0.06	MgO	
SrO	0.74	0.69	SrO	0.68	0.79	0.19	SrO		0.44	0.44	0.74	SrO	0.37
CaO	6.64	6.54	CaO	4.47	5.27	1.49	CaO	6.20	6.93	6.66	6.64	CaO	5.29
Na ₂ O	8.32	8.75	Na ₂ O	15.00	17.88	17.81	K ₂ O					Na ₂ O	9.73
K ₂ O			K ₂ O	1.19	0.05	3.21	Na ₂ O	14.54	3.53	3.82	8.32	K ₂ O	0.05
TOTAL	79.77	79.37	TOTAL	86.20	86.56	85.25	TOTAL	87.92	80.29	79.48	79.77	TOTAL	80.50
Si	6.144	6.111	Si	6.043	6.032	5.961	Si	6.152	6.000	5.987	6.144	Si	5.803
Al	5.856	5.889	Al	5.957	5.968	6.039	Al	5.848	6.000	6.013	5.856	Al	6.197
Fe ²⁺	0.116	0.069	Fe ²⁺	0.074	0.076	0.016	Fe ²⁺	0.083	0.053	0.065	0.116	Fe ²⁺	0.015
Mg	0.016	0.011	Mg	0.012	0.017		Mg		0.050	0.057	0.016	Mg	
Sr	0.076	0.065	Sr	0.070	0.080	0.021	Sr		0.042	0.042	0.076	Sr	0.035
Ca	1.250	1.147	Ca	0.825	1.012	0.284	Ca	1.110	1.198	0.783	1.250	Ca	0.965
Na	2.834	1.189	Na	5.017	6.205	6.134	K					Na	3.212
K			K	0.263	0.013	0.726	Na	4.706	1.103	1.211	2.834	K	0.012
TOTAL	16.295	14.486	TOTAL	18.252	19.398	19.181	TOTAL	17.898	14.450	14.537	16.295	TOTAL	16.238
O	23.952	22.945	O	24.632	25.308	25	O	24.622	22.901	22.923	23.952	O	23.529
Si+Al = 12			Si+Al = 12				Si+Al = 12				Si+Al = 12		

C.90.y	ijolite		C.IJ.14	urtite		C.190.105	mel		E.133.3 ¹	mel	S.78.15	mel	
	PT 87	PT 88		PT 4	PT 96		PT 82	PT 83		PT 63		PT 36	PT 40
SiO ₂	36.60	35.94	SiO ₂	37.22	39.90	SiO ₂	33.79	35.12	SiO ₂	34.75	SiO ₂	36.76	34.77
Al ₂ O ₃	30.54	31.75	Al ₂ O ₃	31.41	30.75	Al ₂ O ₃	29.68	31.30	Al ₂ O ₃	27.29	Al ₂ O ₃	32.22	31.83
FeO	0.70	0.43	FeO	0.08	0.64	FeO	0.38	0.38	Fe ₂ O ₃	0.59	FeO		0.43
MgO			MgO			MgO	0.23	0.28	MgO		MgO	0.06	0.24
SrO	0.48	0.50	SrO	0.27	0.70	SrO	0.50	0.43	SrO	0.79	SrO		0.18
CaO	5.39	5.79	CaO	4.84	2.69	CaO	6.92	7.23	CaO	5.23	CaO	0.05	5.98
Na ₂ O	3.32	3.76	Na ₂ O	5.45	4.80	Na ₂ O	7.07	3.82	Na ₂ O	9.32	Na ₂ O	21.81	4.69
K ₂ O			K ₂ O	0.06	0.11	K ₂ O			K ₂ O		K ₂ O		
TOTAL	77.03	78.17	TOTAL	79.33	79.59	TOTAL	78.57	78.56	TOTAL	77.97	TOTAL	90.90	78.12
Si	6.050	5.879	Si	6.015	6.289	Si	5.896	5.853	Si	6.233	Si	5.903	5.773
Al	5.950	6.121	Al	5.985	5.711	Al	6.104	6.147	Al	5.767	Al	6.097	6.227
Fe ²⁺	0.097	0.059	Fe ²⁺	0.011	0.086	Fe ²⁺	0.056	0.053	Fe ³⁺	0.080	Fe ²⁺		0.060
Mg			Mg			Mg	0.059	0.069	Mg		Mg	0.015	0.060
Sr	0.046	0.047	Sr	0.027	0.064	Sr	0.051	0.041	Sr	0.084	Sr	0.000	0.017
Ca	0.954	1.015	Ca	0.838	0.453	Ca	1.293	1.291	Ca	1.006	Ca	0.009	1.065
Na	1.064	1.192	Na	1.709	1.465	Na	2.393	1.234	Na	3.240	Na	6.789	1.510
K			K	0.011	0.022	K			K		K		
TOTAL	14.161	14.314	TOTAL	14.593	14.087	TOTAL	15.853	14.689	TOTAL	16.410	TOTAL	18.810	14.709
O	22.655	22.658	O	22.742	22.486	O	23.605	22.998	O	23.944	O	24.365	22.841
Si+Al = 12			Si+Al = 12			Si+Al = 12			Si+Al = 12		Si+Al = 12		

APPENDIX Jvi CANCRINITE (cont.)

TUR I	<u>melilitolite</u>							TUR II	<u>"copiamel"</u>	C.18.26	<u>phosc</u>	TU 119	<u>turite</u>
	PT 1	PT 2	PT 3	PT 4	PT 7	PT 5	PT 6		PT 41		PT 28		PT 3
SiO ₂	33.78	34.48	34.24	34.30	34.26	34.16	34.21	SiO ₂	33.83	SiO ₂	38.18	SiO ₂	34.97
Al ₂ O ₃	28.76	29.67	29.12	29.41	29.56	29.03	29.10	Al ₂ O ₃	27.70	Al ₂ O ₃	31.13	Al ₂ O ₃	30.79
CaO	0.30	0.20	0.22	0.27	0.21	0.49	0.32	Fe ₂ O ₃	0.34	FeO		FeO	0.56
Na ₂ O	22.50	23.08	22.46	16.61	17.04	23.35	24.35	MgO		MgO		MgO	
K ₂ O	0.30	0.19	0.23	0.36	0.23	0.49	0.29	SrO		SrO	0.22	SrO	0.46
Cl	0.67	1.98	0.10	0.18	0.20	0.17	0.20	CaO	6.19	CaO	5.01	CaO	5.52
SO ₃	6.79	5.30	7.45					Na ₂ O	19.55	Na ₂ O	5.94	Na ₂ O	5.31
O=Cl	-0.15	-0.45	-0.02	-0.04	-0.05	-0.04	-0.05	K ₂ O		K ₂ O		K ₂ O	
TOTAL	92.95	94.45	93.80	81.09	81.45	87.82	88.42	TOTAL	87.61	TOTAL	80.48	TOTAL	77.61
Si	5.990	5.958	5.993	5.967	5.950	5.996	5.992	Si	6.107	Si	6.119	Si	5.889
Al	6.010	6.042	6.007	6.033	6.050	6.004	6.008	Al	5.893	Al	5.881	Al	6.111
Ca	0.056	0.377	0.042	0.050	0.039	0.091	0.060	Fe ³⁺	0.047	Fe ²⁺		Fe ²⁺	0.080
Na	7.736	7.732	7.621	5.604	5.737	7.945	8.271	Mg		Mg		Mg	
K	0.068	0.421	0.050	0.081	0.050	0.109	0.065	Sr		Sr	0.020	Sr	0.046
Cl	0.203	0.580	0.030	0.052	0.058	0.049	0.059	Ca	1.198	Ca	0.860	Ca	0.997
S	0.904	0.687	0.979					Na	6.844	Na	1.846	Na	1.735
								K		K		K	
TOTAL	20.967	21.798	20.723	17.786	17.883	20.195	20.455	TOTAL	20.088	TOTAL	14.726	TOTAL	14.851
O	27.665	26.964	26.699	23.904	23.936	26.215	26.307	O	25.74416734	O	22.863	O	22.930
Si+Al = 12								Si+Al = 12		Si+Al = 12		Si+Al = 12	

APPENDIX Jvii CARBONATES (CO₂* = calculated CO₂ content)

C.A.347 <u>pyroxenite</u>				C.90.29 <u>pyroxenite</u>				C.AG.48A <u>ol.melteigite</u>				C.AG.122 <u>ijolite</u>			
calcite	PT 92	PT 93		calcite	PT 96	PT 98		calcite	PT 95	PT 96		calcite	PT 96	PT 99	
Na ₂ O				Na ₂ O				Na ₂ O	0.01	0.03		Na ₂ O			
CaO	55.73	55.34		CaO	53.45	54.19		CaO	53.00	53.05		CaO	52.29	52.25	
MgO		0.06		MgO	0.04			MgO	0.18	0.15		MgO			
MnO				MnO	0.07	0.07		MnO	0.05	0.03		MnO		0.05	
FeO	0.13	0.07		FeO	0.11	0.07		FeO	0.08	0.13		FeO	0.06		
SrO	0.73	0.53		SrO	2.40	2.13		SrO	3.20	3.29		SrO	3.58	3.71	
BaO				Ce ₂ O ₃				Ce ₂ O ₃	0.01			BaO	0.16	0.18	
CO ₂ *	44.13	43.76		CO ₂ *	43.12	43.52		CO ₂ *	43.24	43.31		CO ₂ *	42.64	42.66	
TOTAL	100.72	99.76		TOTAL	99.19	99.98		TOTAL	99.77	99.99		TOTAL	98.73	98.85	
Na				Na				Na		0.001		Na			
Ca	0.991	0.992		Ca	0.973	0.977		Ca	0.962	0.961		Ca	0.962	0.961	
Mg		0.001		Mg	0.001			Mg	0.005	0.004		Mg			
Mn				Mn	0.001	0.001		Mn	0.001			Mn		0.001	
Fe	0.002	0.001		Fe	0.002	0.001		Fe	0.001	0.002		Fe	0.001		
Sr	0.007	0.005		Sr	0.024	0.021		Sr	0.031	0.032		Sr	0.036	0.037	
Ba				Ce				Ce				Ba	0.001	0.001	
C	1	1		C	1	1		C	1	1		C	1	1	
TOTAL	2.000	2.000		TOTAL	2.000	2.000		TOTAL	2.000	2.000		TOTAL	2.000	2.000	
O	3	3		O	3	3		O	3	3		O	3	3	

C.ND.500 <u>ijolite</u>				C.190.105 <u>turjaite</u>				S.78.15 <u>melilitolite</u>				WK T <u>turjaite</u>				
calcite	PT 97	PT 98	PT 99	calcite	PT 84	PT 85		cal/dol	PT 96	PT 95		cal/dol	PT 82	PT 84	PT 86	PT 91
Na ₂ O	0.02		0.01	Na ₂ O				Na ₂ O				Na ₂ O				
CaO	56.35	54.04	55.71	CaO	54.27	55.29		CaO	56.80	7.12		CaO	53.83	53.31	29.81	30.01
MgO	0.02	0.08	0.02	MgO				MgO				MgO	0.63	0.81	19.85	20.03
MnO		0.10		MnO				MnO				MnO	0.15	0.13	0.22	0.25
FeO	0.04	0.06	0.03	FeO	0.08			FeO				FeO	0.23	0.33	1.28	0.96
SrO	0.07	1.59	0.29	SrO	2.02	0.92		SrO		60.61		SrO	0.63	0.71	0.30	0.32
Ce ₂ O ₃				Ce ₂ O ₃				Ce ₂ O ₃				Ce ₂ O ₃				
CO ₂ *	44.31	43.27	43.89	CO ₂ *	43.50	43.78		CO ₂ *	44.58	31.33		CO ₂ *	43.43	43.31	46.12	46.30
TOTAL	100.81	99.14	99.95	TOTAL	99.87	99.99		TOTAL	101.38	99.06		TOTAL	98.90	98.60	97.58	97.87
Na	0.001			Na				Na				Na				
Ca	0.998	0.980	0.996	Ca	0.979	0.991		Ca	1.000	0.178		Ca	0.973	0.966	0.508	0.509
Mg		0.002		Mg				Mg				Mg	0.016	0.021	0.470	0.473
Mn		0.001		Mn				Mn				Mn	0.002	0.002	0.003	0.004
Fe	0.001	0.001		Fe	0.001			Fe				Fe	0.003	0.005	0.017	0.013
Sr	0.001	0.016	0.003	Sr	0.020	0.009		Sr		0.822		Sr	0.006	0.007	0.003	0.003
Ce				Ce				Ce				Ce				
C	1	1	1	C	1	1		C	1	1		C	1	1	1	1
TOTAL	2.000	2.000	2.000	TOTAL	2.000	2.000		TOTAL	2.000	2.000		TOTAL	2.000	2.000	2.000	2.000
O	3	3	3	O	3	3		O	3	3		O	3	3	3	3

APPENDIX Jvii CARBONATES (cont.)

TUR I <u>turjaite</u>			C.K.7 <u>phoscorite</u>			C.3.93 <u>phoscorite</u>			C.19.163 <u>phoscorite</u>				
calcite	tur c1	tu3 c3	calcite	PT 90	PT 91	calcite	PT 96	PT 97	cal/dol	PT 94	PT 98	PT 99	PT 95
Na ₂ O			Na ₂ O			Na ₂ O			Na ₂ O				
CaO	55.07	54.78	CaO	55.30	55.22	CaO	51.68	53.83	CaO	52.83	55.19	29.34	3.53
MgO			MgO	0.30	0.28	MgO	0.84	0.25	MgO	0.61	0.14	20.70	
MnO			MnO			MnO	0.15	0.17	MnO	0.25	0.24	0.42	
FeO		0.23	FeO	0.09	0.07	FeO	0.27	0.09	FeO	0.26	0.17	2.61	
SrO	1.74	2.14	SrO	0.50	0.46	SrO	0.75	0.24	SrO	0.75	0.23	0.27	64.77
Ce ₂ O ₃			Ce ₂ O ₃			Ce ₂ O ₃			Ce ₂ O ₃				0.17
CO ₂ *	43.56	43.47	CO ₂ *	43.99	43.88	CO ₂ *	42.05	42.78	CO ₂ *	42.76	43.82	47.60	30.35
TOTAL	100.37	100.62	TOTAL	100.18	99.91	TOTAL	95.74	97.36	TOTAL	97.46	99.79	100.94	98.82
Na			Na			Na			Na				
Ca	0.983	0.976	Ca	0.986	0.988	Ca	0.964	0.987	Ca	0.970	0.989	0.484	0.091
Mg			Mg	0.007	0.007	Mg	0.022	0.006	Mg	0.016	0.003	0.475	
Mn			Mn			Mn	0.002	0.002	Mn	0.004	0.003	0.005	
Fe		0.003	Fe	0.001	0.001	Fe	0.004	0.001	Fe	0.004	0.002	0.034	
Sr	0.017	0.021	Sr	0.005	0.004	Sr	0.008	0.002	Sr	0.007	0.002	0.002	0.906
Ce			Ce			Ce			Ce				0.002
C	1	1	C	1	1	C	1	1	C	1	1	1	1
TOTAL	2.000	2.000	TOTAL	2.000	2.000	TOTAL	2.000	2.000	TOTAL	2.000	2.000	2.000	2.000
O	3	3	O	3	3	O	3	3	O	3	3	3	3

C.18.26 <u>phoscorite</u>			C.TL.344 <u>carbonatite</u>			C.XC <u>carbonatite</u>			C.6.15 <u>carbonatite</u>			
calcite	PT 91	PT 92	calcite	PT 84	PT 88	calcite	PT 97	PT 99	cal/dol	PT 93	PT 96	PT 99
Na ₂ O			Na ₂ O			Na ₂ O			Na ₂ O			
CaO	55.69	54.13	CaO	55.22	55.62	CaO	53.90	53.23	CaO	53.84	53.34	30.39
MgO		0.13	MgO	0.34	0.22	MgO	0.40	0.49	MgO	1.01	1.48	20.97
MnO		0.06	MnO	0.05		MnO	0.12	0.42	MnO	0.18	0.19	0.24
FeO	0.06	0.11	FeO	0.10	0.08	FeO	0.25	0.22	FeO	0.13	0.17	0.99
SrO	0.13	0.62	SrO	0.49	0.45	SrO	0.55	0.38	SrO	0.51	0.59	0.23
Ce ₂ O ₃			Ce ₂ O ₃			Ce ₂ O ₃			Ce ₂ O ₃			
CO ₂ *	43.80	42.99	CO ₂ *	44.01	44.13	CO ₂ *	43.20	42.87	CO ₂ *	43.76	43.95	47.60
TOTAL	99.68	98.04	TOTAL	100.21	100.50	TOTAL	98.42	97.61	TOTAL	99.43	99.72	100.42
Na			Na			Na			Na			
Ca	0.998	0.988	Ca	0.985	0.989	Ca	0.979	0.975	Ca	0.966	0.953	0.501
Mg		0.004	Mg	0.008	0.005	Mg	0.010	0.012	Mg	0.025	0.037	0.481
Mn		0.001	Mn	0.001		Mn	0.002	0.006	Mn	0.003	0.003	0.003
Fe	0.001	0.002	Fe	0.001	0.001	Fe	0.004	0.003	Fe	0.002	0.003	0.013
Sr	0.002	0.006	Sr	0.005	0.004	Sr	0.005	0.004	Sr	0.005	0.006	0.002
Ce			Ce			Ce			Ce			
C	1	1	C	1	1	C	1	1	C	1	1	1
TOTAL	2.000	2.000	TOTAL	2.000	2.000	TOTAL	2.000	2.000	TOTAL	2.000	2.000	2.000
O	3	3	O	3	3	O	3	3	O	3	3	3

APPENDIX Jvii CARBONATES (cont.)

C.18.60 <u>carbonatite</u>			C.23.270b <u>carbonatite</u>			C.23.300 <u>carbonatite</u>			C.49.40 <u>carbonatite</u>				
calcite	PT 97	PT 98	cal/dol	PT 92	PT 93	calcite	PT 91	PT 92	cal/dol	PT 93	PT 94	PT 95	PT 96
Na ₂ O	0.07	0.05	Na ₂ O			Na ₂ O			Na ₂ O				
CaO	52.43	51.93	CaO	55.10	25.72	CaO	54.35	53.56	CaO	53.15	53.18	25.92	29.58
MgO	0.67	0.70	MgO	0.13	21.04	MgO	0.39	0.38	MgO	0.91	1.04	21.93	21.76
MnO	0.09	0.10	MnO	0.08		MnO			MnO	0.20	0.21	0.29	0.25
FeO	0.28	0.28	FeO			FeO		0.09	FeO	0.27	0.30	1.59	1.42
SrO	0.76	0.78	SrO	0.11		SrO	0.46	0.50	SrO	0.61	0.72	0.30	0.28
Ce ₂ O ₃			Ce ₂ O ₃			Ce ₂ O ₃			Ce ₂ O ₃			2.18	
CO ₂ *	42.48	42.12	CO ₂ *	43.48	43.16	CO ₂ *	43.27	42.72	CO ₂ *	43.25	43.49	46.45	48.12
TOTAL	96.78	95.96	TOTAL	98.90	89.92	TOTAL	98.47	97.25	TOTAL	98.39	98.94	98.66	101.41
Na	0.002	0.002	Na			Na			Na				
Ca	0.969	0.968	Ca	0.995	0.468	Ca	0.986	0.984	Ca	0.964	0.960	0.438	0.482
Mg	0.017	0.018	Mg	0.003	0.532	Mg	0.010	0.010	Mg	0.023	0.026	0.516	0.494
Mn	0.001	0.001	Mn	0.001		Mn			Mn	0.003	0.003	0.004	0.003
Fe	0.004	0.004	Fe			Fe		0.001	Fe	0.004	0.004	0.021	0.018
Sr	0.008	0.008	Sr	0.001		Sr	0.005	0.005	Sr	0.006	0.007	0.003	0.002
Ce			Ce			Ce			Ce			0.013	
C	1	1	C	1	1	C	1	1	C	1	1	1	1
TOTAL	2.001	2.001	TOTAL	2.000	2.000	TOTAL	2.000	2.000	TOTAL	2.000	2.000	2.000	2.000
O	3	3	O	3	3	O	3	3	O	3	3	3	3

C.54.x <u>carbonatite</u>						C.100.b <u>carbonatite</u>			C.100.d <u>carbonatite</u>			C.100.y <u>carbonatite</u>			
cal/dol/str	PT 88	PT 92	PT 94	PT 93	PT 90	calcite	PT 97	PT 98	calcite	PT 92	PT 94	calcite	PT 95	PT 96	PT 94
Na ₂ O	0.08	0.08	0.01	0.03	9.71	Na ₂ O			Na ₂ O			Na ₂ O			
CaO	55.86	55.91	56.39	3.77	15.09	CaO	51.86	52.52	CaO	52.45	29.44	CaO	52.46	51.66	28.81
MgO		0.01	0.01		0.01	MgO	1.12	1.00	MgO	0.85	21.12	MgO	0.73	0.88	20.21
MnO			0.03	0.02		MnO	0.16	0.14	MnO	0.13	0.18	MnO	0.14	0.13	0.25
FeO	0.03	0.06	0.04	0.09	0.08	FeO	0.29	0.29	FeO	0.16	1.07	FeO	0.31	0.29	2.45
SrO	0.85	0.68	0.41	63.60	36.28	SrO	0.72	0.70	SrO	0.47	0.21	SrO	0.64	0.71	0.23
Ce ₂ O ₃			0.05	0.71	1.08	Ce ₂ O ₃			BaO			Ce ₂ O ₃			
CO ₂ *	44.27	44.27	44.51	30.35	34.64	CO ₂ *	42.50	42.87	CO ₂ *	42.47	47.02	CO ₂ *	42.52	42.06	46.43
TOTAL	101.09	101.01	101.45	98.57	96.89	TOTAL	96.65	97.52	TOTAL	96.53	99.04	TOTAL	96.80	95.73	98.38
Na	0.003	0.003		0.001	0.398	Na			Na			Na			
Ca	0.990	0.991	0.994	0.097	0.342	Ca	0.958	0.962	Ca	0.969	0.492	Ca	0.968	0.964	0.487
Mg						Mg	0.029	0.026	Mg	0.022	0.491	Mg	0.019	0.023	0.475
Mn						Mn	0.003	0.002	Mn	0.002	0.003	Mn	0.002	0.002	0.003
Fe		0.001	0.001	0.002	0.001	Fe	0.004	0.004	Fe	0.003	0.014	Fe	0.004	0.004	0.032
Sr	0.008	0.007	0.004	0.890	0.445	Sr	0.007	0.007	Sr	0.005	0.002	Sr	0.006	0.007	0.002
Ce				0.006	0.008	Ce			Ba			Ce			
C	1	1	1	1	1	C	1	1	C	1	1	C	1	1	1
TOTAL	2.000	2.001	2.000	1.998	2.295	TOTAL	2.000	2.000	TOTAL	2.000	2.000	TOTAL	2.000	2.000	2.000
O	3	3	3	3	3	O	3	3	O	3	3	O	3	3	3

APPENDIX Jvii CARBONATES (cont.)

N.63.4 <u>carbonatite</u>			N.64.18c <u>carbonatite</u>			N.64.18 <u>carb-mel</u>			N.64.19 <u>carbonatite</u>			
calcite	PT 89	PT 90	calcite	PT 85	PT 89	calcite	PT 94	PT 95	calcite	PT 79	PT 80	PT 83
Na ₂ O			Na ₂ O			Na ₂ O			Na ₂ O			
CaO	55.00	56.12	CaO	55.06	54.80	CaO	55.32	55.21	CaO	55.44	54.54	54.62
MgO	0.07	0.07	MgO	0.08	0.04	MgO			MgO			0.05
MnO		0.05	MnO			MnO			MnO			
FeO	0.08		FeO			FeO			FeO		0.05	
SrO	0.69	0.73	SrO	0.68	0.70	SrO	0.90	0.77	SrO	0.71	0.70	0.65
Ce ₂ O ₃			Ce ₂ O ₃			Ce ₂ O ₃			Ce ₂ O ₃			
CO ₂ *	43.58	44.46	CO ₂ *	43.59	43.35	CO ₂ *	43.8	43.65	CO ₂ *	43.81	43.13	43.20
TOTAL	99.42	101.43	TOTAL	99.41	98.89	TOTAL	100.02	99.63	TOTAL	99.96	98.42	98.52
Na			Na			Na			Na			
Ca	0.990	0.991	Ca	0.991	0.992	Ca	0.992	0.993	Ca	0.993	0.992	0.992
Mg	0.002	0.002	Mg	0.002	0.001	Mg			Mg			0.001
Mn		0.001	Mn			Mn			Mn			
Fe	0.001		Fe			Fe			Fe		0.001	
Sr	0.007	0.007	Sr	0.007	0.007	Sr	0.009	0.008	Sr	0.007	0.007	0.006
Ce			Ce			Ce			Ce			
C	1	1	C	1	1	C	1	1	C	1	1	1
TOTAL	2.000	2.000	TOTAL	2.000	2.000	TOTAL	2.000	2.000	TOTAL	2.000	2.000	2.000
O	3	3	O	3	3	O	3	3	O	3	3	3

N.187.100 <u>carbonatite</u>			TQ <u>turjite</u>			TD1 <u>carbonate - fluorite vein</u>						TD2 <u>aeg-aug-ap vein</u>			
calcite	PT 97	PT 98	calcite	PT 94	PT 95	cal/ank	PT 87	PT 90	PT 88	PT 85	PT 86	cal/ank	PT 89	PT 91	PT 88
Na ₂ O			Na ₂ O	0.06		Na ₂ O						Na ₂ O	0.03		
CaO	55.15	55.15	CaO	54.56	54.26	CaO	54.07	54.46	28.83	27.85	27.67	CaO	54.52	54.26	36.01
MgO			MgO			MgO	0.05	0.04	10.13	8.55	6.99	MgO			7.17
MnO	0.05		MnO			MnO	0.23	0.29	1.49	1.88	2.23	MnO		0.04	0.45
FeO			FeO	0.06	0.11	FeO	0.39	0.19	14.93	17.44	19.82	FeO	0.06	0.04	11.60
SrO	0.88	1.00	SrO	1.58	1.68	SrO	0.19	0.12	0.15	0.07		SrO	0.84	0.93	0.46
Ce ₂ O ₃			Ce ₂ O ₃			Ce ₂ O ₃						BaO	0.15		
CO ₂ *	43.69	43.71	CO ₂ *	43.57	43.36	CO ₂ *	42.95	43.13	43.82	43.07	42.87	CO ₂ *	43.24	43.03	43.67
TOTAL	99.77	99.86	TOTAL	99.83	99.41	TOTAL	97.88	98.23	99.35	98.86	99.58	TOTAL	98.84	98.30	99.36
Na			Na	0.002		Na						Na	0.001		
Ca	0.991	0.990	Ca	0.983	0.982	Ca	0.988	0.991	0.517	0.508	0.507	Ca	0.990	0.990	0.647
Mg			Mg			Mg	0.002	0.001	0.253	0.217	0.178	Mg			0.180
Mn	0.001		Mn			Mn	0.004	0.004	0.021	0.027	0.033	Mn		0.001	0.007
Fe			Fe	0.001	0.002	Fe	0.006	0.003	0.209	0.248	0.283	Fe	0.001	0.001	0.163
Sr	0.009	0.010	Sr	0.015	0.016	Sr	0.002	0.001	0.002	0.001		Sr	0.009	0.009	0.005
Ce			Ce			Ce						Ba	0.001		
C	1	1	C	1	1	C	1	1	1	1	1	C	1	1	1
TOTAL	2.000	2.000	TOTAL	2.001	2.000	TOTAL	2.000	2.000	2.000	2.000	2.000	TOTAL	2.001	2.000	2.000
O	3	3	O	3	3	O	3	3	3	3	3	O	3	3	3

APPENDIX Jviii CHLORITE

S.78.15 <u>melilitolite</u>			C.K.7 <u>phoscorite</u>			C.XC <u>carbonatite</u>		
PT 26	PT 32		pt13	pt14	pt15	pt17	pt18	PT 5 PT 22
SiO ₂	32.79	32.85	SiO ₂	38.93	38.58	37.59	36.32	37.20 37.50
TiO ₂			TiO ₂			0.06	0.05	0.11 0.14
Al ₂ O ₃	12.79	12.94	Al ₂ O ₃	1.82	2.33	1.34	1.72	3.10 2.67
FeO	0.53	0.93	FeO	7.40	7.55	8.82	9.89	7.11 7.76
MnO	0.17	0.26	MnO	0.11	0.11	0.14	0.11	0.08
MgO	36.87	36.45	MgO	36.45	35.46	36.56	35.92	34.55 34.23
SrO			SrO	0.09	0.12	0.10	0.09	
CaO		0.08	CaO	0.15	0.11	0.10	0.16	0.06 0.10
Na ₂ O	0.73	0.22	Na ₂ O	0.09	0.14		0.04	0.25 0.24
K ₂ O		0.12	K ₂ O	0.79	1.75		0.17	1.91 1.84
F			F					0.39
H ₂ O*			H ₂ O*					
O=F			O=F					-0.16
TOTAL	83.88	83.85	TOTAL	85.83	86.15	84.92	84.68	86.14 86.37
Si	3.178	3.187	Si	3.885	3.883	3.792	3.724	3.590 3.619
Ti			Ti			0.005	0.005	0.008 0.010
Al	1.461	1.481	Al	0.065	0.086	0.049	0.065	0.352 0.304
Fe ²⁺	0.044	0.075	Fe ²⁺	0.618	0.635	0.744	0.849	0.574 0.627
Mn	0.014	0.021	Mn	0.002	0.002	0.012	0.009	0.016 0.006
Mg	5.327	5.273	Mg	5.423	5.318	5.500	5.493	4.971 4.925
Sr			Sr	0.005	0.007	0.007	0.005	
Ca		0.009	Ca	0.016	0.012	0.012	0.019	0.006 0.011
Na	0.137	0.042	Na	0.016	0.028		0.007	0.047 0.045
K		0.014	K	0.100	0.224		0.023	0.235 0.226
F			F					0.119
H			H					
O			O					13.881 14.000
TOTAL	10.161	10.101	TOTAL	10.141	10.194	10.120	10.197	9.783 9.771
AN SUM	14	14	AN SUM	14	14	14	14	14 14
C.18.60 <u>carbonatite</u>			C.100.b <u>carbonatite</u>			C.100.y <u>carbonatite</u>		
PT 34	PT 40		PT 12	PT 16		PT 50	PT 51	
SiO ₂	38.55	39.34	SiO ₂	38.46	37.81	SiO ₂	41.57	40.05
TiO ₂	0.11	0.18	TiO ₂		0.11	TiO ₂	0.12	0.06
Al ₂ O ₃	2.96	2.80	Al ₂ O ₃	1.56	1.80	Al ₂ O ₃	1.89	1.99
FeO	8.68	7.27	FeO	9.20	9.77	FeO	3.07	3.15
MnO			MnO			MnO	0.06	0.06
MgO	34.53	34.64	MgO	35.35	33.15	MgO	38.20	38.36
SrO			SrO			SrO		
CaO	0.21	0.10	CaO	0.25	0.58	CaO	0.46	0.76
Na ₂ O	0.08	0.17	Na ₂ O	0.05	0.10	Na ₂ O	0.10	0.11
K ₂ O	1.28	2.24	K ₂ O	1.02	1.73	K ₂ O	0.81	0.66
F	0.49	0.42	F		0.47	F		
H ₂ O*	11.96	12.08	H ₂ O*			H ₂ O*	12.62	12.41
O=F	-0.21	-0.18	O=F		-0.20	O=F		
TOTAL	98.65	99.06	TOTAL	85.89	85.32	TOTAL	98.90	97.61
Si	3.791	3.843	Si	3.819	3.805	Si	3.950	3.869
Ti	0.008	0.013	Ti		0.009	Ti	0.009	0.004
Al	0.343	0.322	Al	0.182	0.214	Al	0.212	0.227
Fe ²⁺	0.714	0.594	Fe ²⁺	0.763	0.823	Fe ²⁺	0.244	0.254
Mn			Mn			Mn	0.005	0.005
Mg	5.063	5.044	Mg	5.231	4.974	Mg	5.411	5.524
Sr			Sr			Sr		
Ca	0.022	0.010	Ca	0.026	0.063	Ca	0.047	0.079
Na	0.015	0.032	Na	0.009	0.019	Na	0.018	0.021
K	0.161	0.279	K	0.130	0.222	K	0.098	0.081
F	0.152	0.130	F		0.149	F		
H	7.848	7.870	H			H	8.000	8.000
O	10.848	10.870	O	14.000	13.851	O	10.000	10.000
TOTAL	10.117	10.138	TOTAL	10.161	10.126	TOTAL	9.994	10.064
AN SUM	14	14	AN SUM	14	14	AN SUM	14	14

APPENDIX Jix CUSPIDINE

E.133.30 <u>melilitolite</u>			S.162.300 <u>melilitolite</u>			TD 2 <u>aeg-aug-ap vein</u>				TUR II <u>melilitolite</u>			
	PT 58	PT 42		PT 20	PT 21		PT 4	PT 7	PT 25		PT 31	PT 35	PT 67
SiO ₂	32.50	32.79	SiO ₂	32.54	32.56	SiO ₂	31.59	32.35	32.57	SiO ₂	32.95	32.88	33.30
FeO	0.11	0.12	FeO	0.09	0.10	Al ₂ O ₃			0.04	CaO	62.11	61.39	61.89
MnO			MnO	0.06		FeO	0.11	0.12	0.14	TOTAL	95.06	94.27	95.19
MgO	0.05	0.07	MgO	0.11	0.09	MgO	0.11	0.09	0.13				
SrO	0.59	0.35	SrO	0.32	0.34	BaO		0.19		Si	1.990	2.000	2.004
CaO	59.27	59.39	CaO	59.00	59.22	SrO	0.38	0.37	0.39	Ca	4.019	4.000	3.991
Na ₂ O	0.08	0.14	Na ₂ O	0.22	0.26	CaO	63.23	60.66	60.35	TOTAL	6.009	6.000	5.995
F	14.63	14.69	F	12.03	11.80	Na ₂ O	0.36	0.27	0.36	O	8	8	8
O=F	-6.16	-6.19	O=F	-5.07	-4.97	K ₂ O	3.04		0.10				
TOTAL	101.07	101.36	TOTAL	99.30	99.40	TOTAL	98.92	94.05	94.08				
Si	1.708	1.715	Si	1.759	1.761	Si	1.891	1.984	1.992				
Fe ²⁺	0.005	0.005	Fe ²⁺	0.004	0.004	Al			0.003				
Mn			Mn	0.003		Fe ²⁺	0.005	0.007	0.007				
Mg	0.004	0.005	Mg	0.009	0.007	Mg	0.009	0.008	0.012				
Sr	0.017	0.011	Sr	0.011	0.011	Ba		0.004					
Ca	3.337	3.328	Ca	3.417	3.432	Sr	0.013	0.013	0.013				
Na	0.008	0.015	Na	0.023	0.027	Ca	4.053	3.985	3.955				
F	2.432	2.429	F	2.056	2.019	Na	0.041	0.030	0.043				
O	5.568	5.571	O	5.944	5.981	K	0.232		0.008				
TOTAL	5.080	5.077	TOTAL	5.224	5.243	TOTAL	6.247	6.032	6.032				
AN SUM	8	8	AN SUM	8	8	AN SUM	8	8	8				

APPENDIX Jx**FELDSPAR**

<u>ijolite-fenite</u>			<u>C.190.y</u>		<u>ijolite-fenite</u>			TD2	<u>aeg-aug-ap vein</u>				
K-spar	PT 63	PT 67	K-spar	PT 5	PT 27	PT 29	PT 32	albite	PT 44	PT 51	K-spar	PT 36	PT 50
Na ₂ O	0.33	0.12	Na ₂ O	1.19	0.53	1.25	1.16	Na ₂ O	11.78	11.65		0.16	0.13
K ₂ O	18.16	17.87	K ₂ O	16.29	18.77	16.78	16.57	K ₂ O				17.20	17.02
BaO		0.20	BaO	1.52	0.52	0.55	1.29	BaO					
CaO			CaO					CaO	0.10	0.05		0.08	
FeO	0.09		FeO	0.42	0.07	0.09	0.15	FeO		0.18			0.10
Al ₂ O ₃	19.12	19.05	Al ₂ O ₃	19.50	19.25	19.37	19.36	Al ₂ O ₃	19.15	18.87		17.69	17.87
SiO ₂	66.33	66.32	SiO ₂	65.02	66.02	65.78	66.44	SiO ₂	67.97	67.30		64.08	64.44
TOTAL	104.03	103.56	TOTAL	103.94	105.16	103.82	104.97	TOTAL	99.00	98.05		99.21	99.56
Na	0.029	0.010	Na	0.104	0.046	0.109	0.100	Na	1.008	1.007		0.015	0.012
K	1.039	1.025	K	0.939	1.071	0.962	0.942	K				1.029	1.013
Ba		0.004	Ba	0.027	0.009	0.010	0.023	Ba					
Ca			Ca					Ca	0.005	0.002		0.004	
Fe ²⁺	0.003		Fe ²⁺	0.016	0.003	0.003	0.006	Fe ²⁺		0.007			0.004
Al	1.010	1.010	Al	1.039	1.015	1.026	1.017	Al	0.996	0.991		0.977	0.982
Si	2.974	2.982	Si	2.939	2.954	2.956	2.962	Si	2.999	3.000		3.004	3.005
TOTAL	5.055	5.031	TOTAL	5.064	5.098	5.066	5.050	TOTAL	5.007	5.008		5.029	5.016
O	8	8	O	8	8	8	8	O	8	8		8	8

NC.153.82fenite

albite	PT 55	PT 59	PT 74	K-spar	PT 58	PT 60	PT 61	PT 62	PT 75
Na ₂ O	11.58	11.54	11.43		0.32	4.90	1.69	5.18	2.22
K ₂ O	0.15	0.25	0.24		17.40	9.87	15.32	9.86	13.94
BaO					0.26	0.55	0.62	0.54	0.91
CaO						0.07	0.05	0.10	
FeO	0.45	0.42	0.39			0.07	0.06	0.06	0.10
Al ₂ O ₃	18.76	18.86	18.82		16.85	17.15	16.99	17.33	17.00
SiO ₂	68.21	67.80	68.47		66.74	66.97	65.46	66.36	65.63
TOTAL	99.15	98.87	99.35		101.57	99.58	100.19	99.43	99.80
Na	0.991	0.991	0.976		0.028	0.433	0.152	0.459	0.199
K	0.008	0.015	0.013		1.016	0.573	0.905	0.575	0.823
Ba					0.005	0.010	0.011	0.010	0.017
Ca						0.003	0.002	0.005	
Fe ²⁺	0.016	0.016	0.015			0.003	0.002	0.002	0.004
Al	0.976	0.984	0.976		0.909	0.921	0.927	0.933	0.928
Si	3.011	3.003	3.013		3.055	3.050	3.032	3.033	3.038
TOTAL	5.001	5.008	4.993		5.013	4.993	5.033	5.017	5.009
O	8	8	8		8	8	8	8	8

TU 120Aturjite

celsian	PT 1	PT 2	PT 3	PT 4
Na ₂ O	0.72	0.60	0.65	
K ₂ O	2.07	1.89	0.94	2.11
BaO	35.25	35.90	38.41	36.46
CaO		0.23	0.47	
FeO		0.38	0.54	
Al ₂ O ₃	25.06	25.48	25.86	25.74
SiO ₂	35.85	35.64	33.45	35.59
TOTAL	98.95	100.12	100.32	99.90
Na	0.170	0.139	0.154	
K	0.320	0.290	0.148	0.325
Ba	1.677	1.696	1.855	1.728
Ca		0.030	0.062	
Fe ²⁺		0.038	0.055	
Al	3.585	3.620	3.756	3.668
Si	4.351	4.295	4.122	4.303
TOTAL	10.103	10.108	10.152	10.024
O	16	16	16	16

APPENDIX Jxi   GARNET**C.90.29**pyroxenite

	PT 7	PT 12	PT 13	PT 15	PT 18
SiO ₂	34.15	34.12	33.17	32.12	34.58
TiO ₂	5.19	4.99	6.33	8.49	4.30
Al ₂ O ₃	1.26	0.96	1.00	1.81	0.89
Fe ₂ O ₃	24.39	24.49	23.86	21.28	24.22
FeO	2.17	1.78	2.04	3.05	2.10
MnO	0.31	0.34	0.33	0.34	0.31
MgO	0.32	0.23	0.39	0.44	0.21
SrO					
CaO	32.66	32.68	32.46	32.24	32.39
Na ₂ O	0.13	0.19	0.15	0.12	0.20
TOTAL	100.58	99.78	99.73	99.89	99.21

Si	5.710	5.750	5.604	5.414	5.854
Ti	0.653	0.632	0.804	1.076	0.547
Al	0.248	0.191	0.199	0.360	0.178
Fe ³⁺	3.069	3.106	3.033	2.699	3.086
Fe ²⁺	0.304	0.251	0.289	0.430	0.298
Mn	0.044	0.049	0.047	0.049	0.044
Mg	0.080	0.058	0.098	0.111	0.053
Sr					
Ca	5.851	5.901	5.876	5.822	5.875
Na	0.042	0.062	0.049	0.039	0.066

TOTAL	16.000	16.000	16.000	16.000	16.000

O	24	24	24	24	24

C.ND.500ijolite

	PT 11	PT 12	PT 18
SiO ₂	28.39	27.15	33.69
TiO ₂	14.47	15.92	4.68
Al ₂ O ₃	1.49	1.47	1.81
Fe ₂ O ₃	17.78	17.43	25.06
FeO	3.17	2.67	0.23
MnO	0.34	0.29	0.23
MgO	1.14	1.32	0.65
SrO			
CaO	31.82	31.75	33.46
Na ₂ O	0.14	0.17	
TOTAL	98.74	98.17	99.81

Si	4.863	4.683	5.650
Ti	1.864	2.065	0.590
Al	0.301	0.299	0.358
Fe ³⁺	2.292	2.262	3.162
Fe ²⁺	0.454	0.385	0.033
Mn	0.049	0.042	0.033
Mg	0.291	0.339	0.163
Sr			
Ca	5.840	5.868	6.012
Na	0.046	0.057	

TOTAL	16.000	16.000	16.000

O	24	24	24

C.AG.122melteigite

	PT 11	PT 12	PT 31	PT 32	PT 35	PT 42	PT 48
SiO ₂	31.03	33.66	33.77	29.89	31.02	28.73	27.00
TiO ₂	9.43	5.06	7.83	10.85	8.93	13.48	16.72
Al ₂ O ₃	2.25	0.92	2.50	2.79	2.51	2.34	1.88
Fe ₂ O ₃	21.44	25.96	19.57	21.26	22.48	18.91	16.68
FeO	1.80	0.27	1.06	0.95	0.81	2.27	3.43
MnO	0.26	0.22	0.25	0.24	0.22	0.18	0.28
MgO	0.85	0.40	2.80	1.01	0.83	1.13	1.11
SrO							
CaO	32.57	33.63	31.32	32.89	32.93	32.50	31.91
Na ₂ O	0.06	0.11	0.23	0.08	0.09	0.08	0.16
TOTAL	99.69	100.23	99.33	99.96	99.82	99.63	99.17

Si	5.231	5.649	5.595	5.022	5.215	4.861	4.615
Ti	1.195	0.639	0.976	1.371	1.129	1.715	2.149
Al	0.447	0.182	0.488	0.552	0.497	0.467	0.379
Fe ³⁺	2.720	3.279	2.439	2.688	2.844	2.408	2.145
Fe ²⁺	0.254	0.038	0.147	0.133	0.114	0.321	0.491
Mn	0.037	0.031	0.035	0.034	0.031	0.026	0.041
Mg	0.214	0.100	0.692	0.253	0.208	0.285	0.283
Sr							
Ca	5.883	6.047	5.560	5.920	5.932	5.891	5.844
Na	0.020	0.036	0.069	0.026	0.029	0.026	0.053

TOTAL	16.001	16.001	16.001	15.999	15.999	16.000	16.000

O	24	24	24	24	24	24	24

C.170.38ijolite

	PT 38	PT 42
SiO ₂	33.89	34.70
TiO ₂	4.27	3.72
Al ₂ O ₃	3.85	4.01
Fe ₂ O ₃	23.32	22.75
FeO	0.40	0.75
MnO	0.14	0.17
MgO	0.58	0.59
SrO		
CaO	33.40	33.46
Na ₂ O		
TOTAL	99.85	100.15

Si	5.632	5.732
Ti	0.532	0.464
Al	0.752	0.780
Fe ³⁺	2.916	2.828
Fe ²⁺	0.056	0.104
Mn	0.020	0.024
Mg	0.144	0.144
Sr		
Ca	5.948	5.924
Na		

TOTAL	16.000	16.000

O	24	24

C.IJ.14ijolite

	PT 7	PT 25
SiO ₂	30.05	32.72
TiO ₂	12.11	9.39
Al ₂ O ₃	0.84	0.30
Fe ₂ O ₃	20.00	21.16
FeO	3.32	3.42
MnO	0.26	0.26
MgO	0.80	0.33
SrO		
CaO	32.02	32.31
Na ₂ O	0.17	0.41
TOTAL	99.57	100.30

Si	5.113	5.508
Ti	1.550	1.189
Al	0.168	0.060
Fe ³⁺	2.561	2.680
Fe ²⁺	0.473	0.482
Mn	0.037	0.037
Mg	0.203	0.083
Sr		
Ca	5.838	5.828
Na	0.056	0.134

TOTAL	16.000	16.000

O	24	24

E.123.106ijolite

	PT 6	PT 22	PT 24
SiO ₂	33.29	31.80	32.48
TiO ₂	7.66	10.35	9.60
Al ₂ O ₃	0.38	0.39	0.39
Fe ₂ O ₃	21.33	19.59	19.43
FeO	3.44	4.02	3.70
MnO	0.38	0.38	0.40
MgO	0.05	0.18	0.11
ZrO ₂	0.33	0.49	0.40
CaO	31.77	31.74	31.74
Na ₂ O	0.49	0.48	0.59
TOTAL	99.12	99.42	98.85

Si	5.668	5.418	5.545
Ti	0.981	1.326	1.232
Al	0.076	0.078	0.078
Fe ³⁺	2.733	2.512	2.496
Fe ²⁺	0.490	0.573	0.529
Mn	0.055	0.055	0.058
Mg	0.013	0.046	0.028
Zr	0.027	0.041	0.033
Ca	5.796	5.794	5.805
Na	0.162	0.159	0.195

TOTAL	16.000	16.000	16.000

O	24	24	24

APPENDIX Jxi GARNET (cont.)

WKM	<u>ijolite</u>			C.190.105	<u>turjaite</u>				N.64.18m	<u>carbonatite-melilitolite contact</u>				
	PT 59	PT 65	PT 68		PT 61	PT 65	PT 74	PT 78		PT 27	PT 32	PT 33	PT 38	PT 39
SiO ₂	26.60	25.54	26.49	SiO ₂	35.24	32.54	33.94	31.36	SiO ₂	33.64	36.53	33.91	35.98	36.09
TiO ₂	15.48	16.73	15.60	TiO ₂	3.69	7.81	5.05	10.34	TiO ₂	4.41	0.16	4.02	0.40	1.12
Al ₂ O ₃	1.11	1.22	1.18	Al ₂ O ₃	3.51	4.01	3.76	2.05	Al ₂ O ₃	2.02	4.05	1.99	11.06	2.87
Fe ₂ O ₃	19.88	20.00	19.98	Fe ₂ O ₃	22.10	20.54	22.24	18.84	Fe ₂ O ₃	24.85	24.92	25.32	10.34	25.29
FeO	1.44	0.82	1.09	FeO	1.23	0.53	0.67	2.29	FeO			0.37		0.16
MnO	0.41	0.44	0.36	MnO	0.20	0.15	0.17	0.27	MnO	0.14	0.10			
MgO	1.14	1.24	1.19	MgO	0.50	0.94	0.65	0.98	MgO	0.74	0.13	0.45	2.88	0.34
SrO				SrO					SrO				0.21	0.10
CaO	31.97	32.14	32.27	CaO	33.67	33.83	33.66	33.16	CaO	33.39	34.19	33.56	34.73	33.82
Na ₂ O	0.19	0.20	0.17	Na ₂ O		0.05			Na ₂ O				0.05	
TOTAL	98.22	98.33	98.32	TOTAL	100.13	100.40	100.14	99.30	TOTAL	99.19	100.08	99.62	95.65	99.78
Si	4.608	4.424	4.58	Si	5.825	5.372	5.619	5.289	Si	5.665	6.026	5.695	5.937	5.998
Ti	2.016	2.18	2.028	Ti	0.459	0.970	0.629	1.311	Ti	0.558	0.020	0.508	0.050	0.140
Al	0.228	0.248	0.24	Al	0.684	0.780	0.734	0.407	Al	0.401	0.787	0.394	2.151	0.562
Fe ³⁺	2.592	2.608	2.6	Fe ³⁺	2.749	2.551	2.771	2.392	Fe ³⁺	3.149	3.093	3.200	1.283	3.162
Fe ²⁺	0.208	0.12	0.156	Fe ²⁺	0.169	0.073	0.093	0.323	Fe ²⁺			0.051		0.022
Mn	0.06	0.064	0.052	Mn	0.028	0.021	0.024	0.039	Mn	0.020	0.014			
Mg	0.296	0.32	0.308	Mg	0.123	0.231	0.160	0.246	Mg	0.186	0.032	0.113	0.708	0.084
Sr				Sr					Sr				0.020	0.010
Ca	5.932	5.968	5.98	Ca	5.963	5.984	5.971	5.992	Ca	6.024	6.043	6.039	6.140	6.022
Na	0.064	0.068	0.056	Na		0.016			Na				0.016	
TOTAL	16.000	16.000	16.000	TOTAL	16.000	16.000	16.000	16.000	TOTAL	16.002	16.014	16.000	16.305	16.000
O	24	24	24	O	24	24	24	24	O	24	24	24	24	24

N.64.18	<u>melilitolite</u>				E.133.30	<u>turjaite</u>		S.78.15	<u>melilitolite</u>		S.162.300	<u>ij-mel contact</u>		
	PT 44	PT 46	PT 47	PT 48		PT 37	PT 46		PT 25	PT 35		PT 7	PT 12	PT 28
SiO ₂	35.96	36.01	33.89	36.55	SiO ₂	37.93	37.63	SiO ₂	34.07	34.79	SiO ₂	32.22	27.66	31.68
TiO ₂	1.24	0.10	4.38	0.75	TiO ₂		0.12	TiO ₂	2.58	2.32	TiO ₂	8.91	15.39	8.76
Al ₂ O ₃	3.23	3.68	2.35	3.11	Al ₂ O ₃	9.31	9.60	Al ₂ O ₃	1.16	3.31	Al ₂ O ₃	0.55	0.83	0.98
Fe ₂ O ₃	25.03	25.38	25.19	25.49	Fe ₂ O ₃	17.58	16.63	Fe ₂ O ₃	25.39	23.49	Fe ₂ O ₃	21.66	19.03	22.59
FeO			0.26	0.32	FeO	0.28	0.43	FeO	0.01		FeO	2.50	2.81	1.41
MnO					MnO	0.13	0.09	MnO			MnO	0.24	0.33	0.24
MgO	0.25	0.14	0.56	0.24	MgO	0.05	0.08	MgO	0.46	0.44	MgO	0.61	1.15	0.90
SrO					ZrO ₂			SrO			SrO			
CaO	34.09	34.35	33.72	34.06	CaO	35.01	34.69	CaO	32.96	33.70	CaO	32.65	31.77	32.81
Na ₂ O					Na ₂ O			Na ₂ O			Na ₂ O	0.19	0.22	0.10
TOTAL	99.80	99.66	100.35	100.51	TOTAL	100.29	99.27	TOTAL	96.63	98.05	TOTAL	99.53	99.19	99.47
Si	5.97	5.99	5.64	6.02	Si	6.065	6.065	Si	5.894	5.872	Si	5.460	4.740	5.364
Ti	0.16	0.01	0.55	0.09	Ti		0.015	Ti	0.336	0.294	Ti	1.136	1.984	1.116
Al	0.63	0.72	0.46	0.60	Al	1.755	1.824	Al	0.236	0.658	Al	0.108	0.168	0.196
Fe ³⁺	3.13	3.18	3.16	3.16	Fe ³⁺	2.115	2.017	Fe ³⁺	3.305	2.984	Fe ³⁺	2.764	2.456	2.880
Fe ²⁺			0.04	0.04	Fe ²⁺	0.038	0.058	Fe ²⁺	0.002		Fe ²⁺	0.352	0.404	0.200
Mn					Mn	0.018	0.012	Mn			Mn	0.036	0.048	0.036
Mg	0.06	0.04	0.14	0.06	Mg	0.012	0.019	Mg	0.119	0.111	Mg	0.156	0.292	0.228
Sr					Zr			Sr			Sr			
Ca	6.06	6.12	6.02	6.02	Ca	5.998	5.991	Ca	6.109	6.094	Ca	5.928	5.836	5.952
Na					Na			Na			Na	0.064	0.072	0.032
TOTAL	16.00	16.05	16.00	16.00	TOTAL	16.000	16.000	TOTAL	16.000	16.013	TOTAL	16.000	16.000	16.000
O	24.00	24	24	24	O	24	24	O	24	24	O	24	24	24

APPENDIX Jxi GARNET (cont.)

TUR I <u>turjaite</u>									TUR II <u>"copiamel"</u>		C.54.x <u>carbonatite</u>			
	PT 19	PT 33	PT 40	PT 6	PT 7	PT 8	PT 19	PT 20		PT 27		PT 73	PT 75	PT 82
SiO ₂	37.19	35.59	37.40	31.61	32.36	34.18	36.17	35.47	SiO ₂	36.29	SiO ₂	35.14	35.39	34.76
TiO ₂	0.28	3.44	0.58	9.28	8.53	4.90	1.27	2.00	TiO ₂		TiO ₂	1.70	2.69	4.27
Al ₂ O ₃	6.38	4.34	5.92	3.37	3.56	4.03	4.12	4.09	Al ₂ O ₃	5.27	Al ₂ O ₃	1.41	2.91	4.18
Fe ₂ O ₃	21.53	20.19	20.68	19.50	19.92	21.82	24.59	23.99	Fe ₂ O ₃	23.32	Fe ₂ O ₃	27.01	23.18	21.51
FeO	0.55	2.19	1.24	1.49	1.69	1.33	0.19	0.21	FeO		FeO	0.40	1.03	0.19
MnO	0.10	0.08	0.07	0.15	0.17				MnO		MnO	0.13	0.15	0.10
MgO	0.18	0.51	0.41	1.12	1.17	0.64	0.26	0.32	MgO	0.33	MgO	0.39	0.57	0.84
SrO									SrO		SrO			
CaO	34.15	33.15	33.43	33.18	33.11	33.41	34.14	33.90	CaO	34.75	CaO	33.03	33.20	34.04
Na ₂ O			0.08						Na ₂ O		Na ₂ O			
TOTAL	100.36	99.49	99.81	99.69	100.51	100.31	100.74	99.98	TOTAL	99.96	TOTAL	99.22	99.12	99.90
Si	6.040	5.891	6.103	5.278	5.352	5.644	5.929	5.862	Si	5.968	Si	5.929	5.917	5.730
Ti	0.034	0.428	0.071	1.165	1.061	0.608	0.157	0.249	Ti		Ti	0.216	0.338	0.529
Al	2.631	0.847	1.138	0.663	0.694	0.784	0.796	0.797	Al	1.021	Al	0.280	0.573	0.812
Fe ³⁺	1.221	2.515	2.539	2.450	2.480	2.711	3.033	2.983	Fe ³⁺	2.886	Fe ³⁺	3.430	2.916	2.668
Fe ²⁺	0.074	0.303	0.169	0.207	0.233	0.184	0.026	0.029	Fe ²⁺		Fe ²⁺	0.057	0.145	0.027
Mn	0.014	0.011	0.010	0.021	0.024				Mn		Mn	0.019	0.021	0.014
Mg	0.044	0.126	0.100	0.279	0.288	0.158	0.064	0.079	Mg	0.081	Mg	0.098	0.142	0.206
Sr									Sr		Sr			
Ca	5.942	5.879	5.844	5.936	5.867	5.911	5.996	6.002	Ca	6.123	Ca	5.971	5.947	6.012
Na			0.025						Na		Na			
TOTAL	16.000	16.000	16.000	16.000	16.000	16.000	16.000	16.000	TOTAL	16.079	TOTAL	16.000	16.000	16.000
O	24	24	24	24	24	24	24	24	O	24	O	24	24	24

N.63.4 <u>carbonatite</u>				N.64.19 <u>carbonatite</u>		
	PT 77	PT 79	PT 83		PT 62	PT 63
SiO ₂	27.05	29.25	29.57	SiO ₂	36.27	36.34
TiO ₂	11.22	9.36	7.74	TiO ₂		
Al ₂ O ₃	0.93	1.74	1.77	Al ₂ O ₃	3.89	3.91
Fe ₂ O ₃	20.26	22.08	22.05	Fe ₂ O ₃	25.25	25.22
FeO				FeO		
MnO	0.35	0.31	0.30	MnO	0.09	0.10
MgO	1.41	1.06	1.11	MgO	0.17	0.18
SrO				SrO	0.10	
CaO	31.49	31.97	32.08	CaO	33.98	33.82
Na ₂ O	0.10	0.10	0.14	Na ₂ O		
TOTAL	92.81	95.87	94.76	TOTAL	99.75	99.57
Si	4.929	5.133	5.250	Si	6.015	6.030
Ti	1.537	1.235	1.033	Ti		
Al	0.200	0.360	0.370	Al	0.760	0.765
Fe ³⁺	2.778	2.916	2.946	Fe ³⁺	3.151	3.149
Fe ²⁺				Fe ²⁺		
Mn	0.054	0.046	0.045	Mn	0.013	0.014
Mg	0.383	0.277	0.294	Mg	0.042	0.045
Sr				Sr	0.010	
Ca	6.147	6.011	6.103	Ca	6.038	6.012
Na	0.035	0.034	0.032	Na		
TOTAL	16.063	16.011	16.074	TOTAL	16.029	16.014
O	24	24	24	O	24	24

N.187.100 <u>carbonatite</u>					
	PT 1	PT 6	PT 11	PT 14	PT 15
SiO ₂	33.27	32.82	33.41	33.24	33.26
TiO ₂	2.22	2.02	1.73	1.43	1.86
Al ₂ O ₃	1.53	1.80	1.46	1.57	2.62
Fe ₂ O ₃	25.21	24.84	25.65	25.45	24.01
FeO					
MnO	0.25	0.25	0.19	0.17	0.29
MgO	0.54	0.53	0.46	0.41	0.49
SrO					
CaO	33.81	33.38	33.44	33.47	33.38
Na ₂ O					
TOTAL	96.83	95.64	96.34	95.74	95.91
Si	5.775	5.764	5.826	5.834	5.793
Ti	0.290	0.267	0.227	0.189	0.244
Al	0.313	0.373	0.300	0.325	0.538
Fe ³⁺	3.292	3.282	3.366	3.361	3.146
Fe ²⁺					
Mn	0.037	0.037	0.028	0.025	0.043
Mg	0.140	0.139	0.120	0.107	0.127
Sr					
Ca	6.287	6.281	6.248	6.294	6.230
Na					
TOTAL	16.133	16.142	16.114	16.135	16.121
O	24	24	24	24	24

APPENDIX Jxi GARNET (cont.)

TU 119												C.190.x			
turiite												jiolite-fenite contact			
	PT 67	PT 70	PT 72	PT 73	PT 74	PT 75	PT 76	PT 77	PT 81	PT 85	PT 86		PT 19	PT 24	PT 34
SiO ₂	33.66	34.76	35.99	36.04	33.64	35.80	28.72	31.44	35.35	35.61	36.21	SiO ₂	30.72	28.49	28.44
TiO ₂	6.07	2.30	1.00	0.35	5.09	0.12	15.29	9.39	1.92	0.61	0.37	TiO ₂	9.57	13.69	15.16
Al ₂ O ₃	1.60	0.55	4.18	4.35	1.56	2.50	0.50	0.76	3.63	4.18	3.11	Al ₂ O ₃	1.06	1.38	1.38
Fe ₂ O ₃	22.73	27.99	23.48	24.38	23.95	27.00	15.13	21.53	23.30	23.92	25.95	Fe ₂ O ₃	21.33	19.13	16.51
FeO	0.73	0.51			0.39		2.79	1.34				FeO	2.22	2.42	3.82
MnO	0.33	0.36	0.24	0.21	0.31	0.32	0.43	0.29	0.24	0.23	0.36	MnO	0.30	0.34	0.29
MgO	0.64	0.14	0.45	0.31	0.55	0.13	1.05	0.91	0.63	0.36	0.21	MgO	0.56	0.80	0.96
SrO			0.09						0.11			SrO			
CaO	33.27	32.67	33.86	34.13	33.08	33.48	30.99	32.60	33.68	34.11	33.79	CaO	31.81	32.17	31.84
Na ₂ O	0.19	0.14			0.16		0.71	0.22				Na ₂ O	0.23	0.21	0.22
TOTAL	99.22	99.42	99.29	99.77	98.73	99.35	95.61	98.48	98.86	99.02	100.00	TOTAL	97.80	98.63	98.62
Si	5.665	5.890	5.974	5.964	5.696	6.008	5.048	5.370	5.908	5.943	6.010	Si	5.303	4.892	4.877
Ti	0.768	0.293	0.125	0.044	0.648	0.015	2.021	1.206	0.241	0.077	0.046	Ti	1.242	1.768	1.955
Al	0.317	0.110	0.818	0.848	0.311	0.494	0.104	0.153	0.715	0.822	0.608	Al	0.216	0.279	0.279
Fe ³⁺	2.879	3.569	2.933	3.036	3.052	3.409	2.001	2.767	2.931	3.004	3.241	Fe ³⁺	2.771	2.472	2.131
Fe ²⁺	0.102	0.073			0.055		0.410	0.191				Fe ²⁺	0.320	0.347	0.548
Mn	0.047	0.052	0.034	0.029	0.044	0.045	0.064	0.042	0.034	0.033	0.051	Mn	0.044	0.049	0.042
Mg	0.161	0.035	0.111	0.076	0.139	0.033	0.275	0.232	0.157	0.090	0.052	Mg	0.144	0.205	0.245
Sr			0.009						0.011			Sr			
Ca	5.999	5.932	6.022	6.051	6.002	6.020	5.836	5.966	6.031	6.099	6.009	Ca	5.883	5.918	5.850
Na	0.062	0.046			0.053		0.242	0.073				Na	0.077	0.070	0.073
TOTAL	16.000	16.000	16.026	16.048	16.000	16.024	16.001	16.000	16.028	16.068	16.017	TOTAL	16.000	16.000	16.000
O	24	24	24	24	24	24	24	24	24	24	24	O	24	24	24

TU 120A							
turiite							
	PT 5	PT 19	PT 20	PT 21	PT 26	PT 28	PT 31
SiO ₂	34.80	34.21	34.89	36.43	33.30	34.61	33.20
TiO ₂	3.26	4.69	3.43	0.74	7.42	3.40	7.72
Al ₂ O ₃	0.87	1.60	1.74	2.83	1.26	1.08	1.18
Fe ₂ O ₃	25.34	23.09	24.43	26.50	21.24	25.97	20.60
FeO	1.81	1.61	1.15	0.20	3.02	1.16	2.99
MnO	0.36	0.29	0.20	0.28	0.31	0.32	0.41
MgO		0.81	0.48	0.31	0.78	0.52	0.68
SrO							
CaO	33.07	32.61	33.25	33.71	32.60	32.81	32.80
Na ₂ O							
TOTAL	99.51	98.91	99.57	101.01	99.93	99.87	99.58
Si	5.886	5.778	5.853	5.993	5.595	5.820	5.597
Ti	0.415	0.596	0.433	0.092	0.938	0.430	0.979
Al	0.173	0.318	0.344	0.549	0.249	0.214	0.234
Fe ³⁺	3.225	2.935	3.085	3.281	2.686	3.287	2.613
Fe ²⁺	0.256	0.227	0.161	0.028	0.424	0.163	0.422
Mn	0.052	0.041	0.028	0.039	0.044	0.046	0.059
Mg		0.204	0.120	0.076	0.195	0.130	0.171
Sr							
Ca	5.993	5.901	5.976	5.942	5.868	5.911	5.925
Na							
TOTAL	16.000	16.000	16.000	16.000	16.000	16.000	16.000
O	24	24	24	24	24	24	24

TQ							
turiite							
	PT 29	PT 31	PT 35	PT 36	PT 40	PT 44	PT 49
SiO ₂	32.04	34.48	35.94	36.48	32.37	33.25	30.90
TiO ₂	10.67	5.59	2.80	0.66	9.73	7.94	12.67
Al ₂ O ₃	1.70	1.91	3.67	4.10	1.92	2.29	0.98
Fe ₂ O ₃	18.71	22.37	23.12	24.35	18.61	20.08	17.39
FeO	1.50	0.68			1.82	0.93	2.59
MnO	0.38	0.31	0.26	0.23	0.33	0.34	0.39
MgO	1.35	1.06	0.72	0.45	1.29	1.17	1.20
SrO	0.09						
CaO	33.67	33.53	34.15	34.10	33.31	33.77	32.97
Na ₂ O	0.09	0.09	0.05		0.07	0.06	0.21
TOTAL	100.19	100.02	100.71	100.37	99.45	99.82	99.30
Si	5.338	5.730	5.884	5.993	5.423	5.534	5.221
Ti	1.337	0.699	0.345	0.082	1.226	0.994	1.610
Al	0.334	0.374	0.708	0.794	0.379	0.449	0.195
Fe ³⁺	2.345	2.798	2.848	3.010	2.346	2.514	2.211
Fe ²⁺	0.209	0.094			0.255	0.129	0.367
Mn	0.054	0.044	0.036	0.032	0.047	0.048	0.056
Mg	0.335	0.263	0.176	0.110	0.322	0.290	0.302
Sr	0.009						
Ca	6.010	5.970	5.990	6.002	5.979	6.022	5.969
Na	0.029	0.029	0.016		0.023	0.019	0.069
TOTAL	16.000	16.000	16.002	16.023	16.000	16.000	16.000
O	24	24	24	24	24	24	24

APPENDIX Jxii MELILITE

C.5.118 mica turjaite

	PT 2	PT 8	PT 13	PT 14	PT 20	PT 85
SiO ₂	43.73	43.87	43.75	43.75	43.90	44.05
Al ₂ O ₃	8.43	8.27	8.20	8.20	7.50	8.16
Fe ₂ O ₃						
FeO	3.59	2.94	3.11	3.11	3.02	3.70
MnO	0.09	0.08	0.08	0.10	0.09	0.13
MgO	6.05	6.60	6.57	6.57	6.82	6.15
BaO			0.24			
SrO	0.63	0.58	0.60	0.57	0.58	0.59
CaO	32.13	32.31	32.04	31.82	32.86	32.03
K ₂ O		0.05			0.05	0.05
Na ₂ O	5.19	4.99	5.08	4.82	4.68	5.04
TOTAL	99.84	99.69	99.67	99.09	99.57	99.90

Si	3.969	3.974	3.974	3.997	3.990	3.992
Al	0.901	0.882	0.877	0.873	0.803	0.873
Fe ³⁺						
Fe ²⁺	0.273	0.222	0.236	0.215	0.229	0.280
Mn	0.007	0.007	0.007	0.007	0.007	0.009
Mg	0.820	0.891	0.889	0.924	0.933	0.831
Ba			0.009			
Sr	0.033	0.030	0.033	0.030	0.030	0.030
Ca	3.120	3.136	3.117	3.099	3.199	3.111
K		0.005			0.007	0.005
Na	0.910	0.877	0.894	0.849	0.824	0.887

TOTAL	10.038	10.026	10.036	9.991	10.024	10.017

O	14	14	14	14	14	14

C.190.105

turjaite

	PT 52	PT 64	PT 66	PT 71	PT 76	PT 79	PT 81
SiO ₂	44.21	43.83	44.05	44.31	43.71	43.85	43.97
Al ₂ O ₃	8.19	8.19	9.45	9.30	8.92	9.18	8.19
Fe ₂ O ₃			0.33			0.44	
FeO	3.06	3.10	2.41	3.68	3.61	3.29	3.10
MnO	0.10	0.08	0.09	0.18	0.20	0.16	0.10
MgO	6.89	6.90	6.29	4.91	5.25	5.12	6.88
BaO							
SrO	0.52	0.53	0.64	1.68	1.80	1.68	0.55
CaO	32.28	32.52	31.21	29.96	29.88	30.22	32.51
K ₂ O							
Na ₂ O	4.74	4.58	5.42	5.64	5.52	5.66	4.57
TOTAL	99.99	99.73	99.89	99.66	98.89	99.60	99.87

Si	3.986	3.967	3.960	4.024	4.010	3.992	3.973
Al	0.870	0.874	1.001	0.995	0.964	0.985	0.872
Fe ³⁺			0.023			0.030	
Fe ²⁺	0.231	0.235	0.181	0.280	0.277	0.250	0.234
Mn	0.008	0.006	0.007	0.014	0.016	0.012	0.008
Mg	0.926	0.931	0.843	0.665	0.718	0.695	0.927
Ba							
Sr	0.027	0.028	0.033	0.088	0.096	0.089	0.029
Ca	3.118	3.154	3.006	2.915	2.937	2.948	3.147
K							
Na	0.829	0.804	0.945	0.993	0.982	0.999	0.801

TOTAL	9.994	9.998	10.000	9.975	9.999	10.000	9.991

O	14	14	14	14	14	14	14

N.64.19 carbonatite

	PT 66	PT 74
SiO ₂	43.51	43.98
Al ₂ O ₃	6.02	5.45
Fe ₂ O ₃	1.39	
FeO	1.58	2.33
MnO	0.17	0.24
MgO	8.60	8.69
BaO		
SrO	0.47	0.58
CaO	34.28	34.76
K ₂ O		
Na ₂ O	3.67	3.35
TOTAL	99.70	99.38

Si	3.953	4.040
Al	0.645	0.586
Fe ³⁺	0.095	
Fe ²⁺	0.120	0.178
Mn	0.013	0.019
Mg	1.165	1.181
Ba		
Sr	0.025	0.031
Ca	3.337	3.396
K		
Na	0.647	0.592

TOTAL	10.000	9.993

O	14	14

N.64.18m melilitolite (contact with carbonatite)

	PT 31	PT 37	PT 41	PT 48	PT 8
SiO ₂	43.27	44.09	44.06	44.03	44.49
Al ₂ O ₃	5.14	5.59	6.01	6.14	6.06
Fe ₂ O ₃	0.37				
FeO	2.13	2.64	1.83	2.99	2.01
MnO	0.19	0.16	0.14	0.18	0.26
MgO	9.05	8.64	8.46	7.68	9.23
BaO					
SrO	0.49	0.52	0.74	0.53	
CaO	35.09	34.13	33.89	33.57	34.55
K ₂ O					
Na ₂ O	3.04	3.32	3.78	3.85	3.72
TOTAL	98.67	99.09	98.91	98.97	100.32

Si	3.980	4.025	4.021	4.028	3.992
Al	0.557	0.601	0.646	0.662	0.642
Fe ³⁺	0.026				
Fe ²⁺	0.156	0.202	0.140	0.229	0.152
Mn	0.015	0.012	0.011	0.014	0.019
Mg	1.241	1.176	1.151	1.047	1.234
Ba					
Sr	0.026	0.028	0.039	0.028	
Ca	3.458	3.338	3.314	3.291	3.323
K					
Na	0.542	0.588	0.669	0.683	0.646

TOTAL	10.000	9.969	9.990	9.982	10.008

O	14	14	14	14	14

N.64.18 calcite melilitolite

	PT 38	PT 42	PT 43	PT 45	PT 50
SiO ₂	44.08	43.64	44.22	43.71	44.17
Al ₂ O ₃	5.87	5.59	5.86	5.65	5.46
Fe ₂ O ₃	1.39	1.57	0.96	0.66	0.95
FeO	1.10	0.92	1.33	1.63	0.99
MnO	0.14	0.19	0.15	0.18	0.11
MgO	9.13	9.04	9.22	9.20	9.30
BaO					
SrO	0.42	0.34	0.38	0.37	0.45
CaO	34.73	35.27	34.72	35.05	35.13
K ₂ O					
Na ₂ O	3.67	3.44	3.63	3.29	3.52
TOTAL	100.53	100.01	100.47	99.75	100.09

Si	3.962	3.950	3.973	3.965	3.985
Al	0.622	0.596	0.621	0.604	0.581
Fe ³⁺	0.094	0.107	0.065	0.045	0.065
Fe ²⁺	0.082	0.070	0.100	0.124	0.075
Mn	0.011	0.015	0.011	0.014	0.008
Mg	1.223	1.220	1.235	1.244	1.251
Ba					
Sr	0.022	0.018	0.020	0.019	0.024
Ca	3.344	3.420	3.343	3.406	3.396
K					
Na	0.640	0.604	0.632	0.579	0.616

TOTAL	10.000	10.000	10.000	10.000	10.000

O	14	14	14	14	14

APPENDIX Jxii MELILITE

C.5.118 mica turjaite

	PT 2	PT 8	PT 13	PT 14	PT 20	PT 85
SiO ₂	43.73	43.87	43.75	43.75	43.90	44.05
Al ₂ O ₃	8.43	8.27	8.20	8.20	7.50	8.16
Fe ₂ O ₃						
FeO	3.59	2.94	3.11	3.11	3.02	3.70
MnO	0.09	0.08	0.08	0.10	0.09	0.13
MgO	6.05	6.60	6.57	6.57	6.82	6.15
BaO			0.24			
SrO	0.63	0.58	0.60	0.57	0.58	0.59
CaO	32.13	32.31	32.04	31.82	32.86	32.03
K ₂ O		0.05			0.05	0.05
Na ₂ O	5.19	4.99	5.08	4.82	4.68	5.04
TOTAL	99.84	99.69	99.67	99.09	99.57	99.90

Si	3.969	3.974	3.974	3.997	3.990	3.992
Al	0.901	0.882	0.877	0.873	0.803	0.873
Fe ³⁺						
Fe ²⁺	0.273	0.222	0.236	0.215	0.229	0.280
Mn	0.007	0.007	0.007	0.007	0.007	0.009
Mg	0.820	0.891	0.889	0.924	0.933	0.831
Ba			0.009			
Sr	0.033	0.030	0.033	0.030	0.030	0.030
Ca	3.120	3.136	3.117	3.099	3.199	3.111
K		0.005			0.007	0.005
Na	0.910	0.877	0.894	0.849	0.824	0.887

TOTAL	10.038	10.026	10.036	9.991	10.024	10.017
O	14	14	14	14	14	14

TUR I turjaite

	PT 1	PT 5	PT 7	PT 8	PT 9	PT 10	PT 11	PT 12
SiO ₂	44.67	45.15	44.69	44.64	44.79	45.29	45.11	45.07
Al ₂ O ₃	7.50	8.82	8.87	9.06	8.93	9.03	9.08	8.67
Fe ₂ O ₃								
FeO	2.38	2.65	2.70	2.72	2.17	2.63	2.54	2.32
MnO	0.09	0.09	0.08	0.10	0.09	0.07	0.07	0.09
MgO	7.59	6.33	6.25	6.39	6.82	6.32	6.36	6.56
BaO								
SrO	0.83	1.11	1.13	1.11	1.16	1.09	0.95	1.26
CaO	32.26	30.55	30.79	30.86	30.75	30.25	30.32	30.60
K ₂ O				0.04				
Na ₂ O	4.94	5.82	5.88	5.93	5.74	5.85	5.92	5.73
TOTAL	100.26	100.52	100.39	100.85	100.45	100.53	100.35	100.30

Si	4.014	4.035	4.009	3.989	4.004	4.040	4.030	4.036
Al	0.794	0.929	0.938	0.954	0.941	0.949	0.956	0.915
Fe ³⁺								
Fe ²⁺	0.179	0.198	0.203	0.203	0.162	0.196	0.190	0.174
Mn	0.007	0.007	0.006	0.008	0.007	0.005	0.005	0.007
Mg	1.017	0.843	0.836	0.851	0.909	0.870	0.847	0.876
Ba								
Sr	0.043	0.058	0.059	0.058	0.060	0.056	0.049	0.065
Ca	3.106	2.925	2.960	2.955	2.945	2.891	2.902	2.936
K				0.005				
Na	0.861	1.009	1.023	1.027	0.995	1.012	1.025	0.995

TOTAL	10.021	10.004	10.034	10.050	10.023	10.019	10.004	10.004
O	14	14	14	14	14	14	14	14

C.190.105 turjaite

	PT 52	PT 64	PT 66	PT 71	PT 76	PT 79	PT 81
SiO ₂	44.21	43.83	44.05	44.31	43.71	43.85	43.97
Al ₂ O ₃	8.19	8.19	9.45	9.30	8.92	9.18	8.19
Fe ₂ O ₃			0.33			0.44	
FeO	3.06	3.10	2.41	3.68	3.61	3.29	3.10
MnO	0.10	0.08	0.09	0.18	0.20	0.16	0.10
MgO	6.89	6.90	6.29	4.91	5.25	5.12	6.88
BaO							
SrO	0.52	0.53	0.64	1.68	1.80	1.68	0.55
CaO	32.28	32.52	31.21	29.96	29.88	30.22	32.51
K ₂ O							
Na ₂ O	4.74	4.58	5.42	5.64	5.52	5.66	4.57
TOTAL	99.99	99.73	99.89	99.66	98.89	99.60	99.87

Si	3.986	3.967	3.960	4.024	4.010	3.992	3.973
Al	0.870	0.874	1.001	0.995	0.964	0.985	0.872
Fe ³⁺			0.023			0.030	
Fe ²⁺	0.231	0.235	0.181	0.280	0.277	0.250	0.234
Mn	0.008	0.006	0.007	0.014	0.016	0.012	0.008
Mg	0.926	0.931	0.843	0.665	0.718	0.695	0.927
Ba							
Sr	0.027	0.028	0.033	0.088	0.096	0.089	0.029
Ca	3.118	3.154	3.006	2.915	2.937	2.948	3.147
K							
Na	0.829	0.804	0.945	0.993	0.982	0.999	0.801

TOTAL	9.994	9.998	10.000	9.975	9.999	10.000	9.991
O	14	14	14	14	14	14	14

TUR II "copiamelilite"

	PT 13	PT 14	PT 15	PT 42	PT 68	PT 86
SiO ₂	44.41	43.94	44.09	42.85	44.03	43.78
Al ₂ O ₃	7.69	7.59	7.74	7.84	8.04	7.78
Fe ₂ O ₃						
FeO	2.52	2.43	2.30	2.15	2.45	2.44
MnO	0.11					
MgO	7.76	7.77	7.83	7.58	7.63	7.81
BaO						
SrO						
CaO	34.13	34.38	34.24	33.57	34.15	34.26
K ₂ O						0.07
Na ₂ O	4.28	3.77	3.95	4.07	4.63	4.27
TOTAL	100.90	99.88	100.15	98.06	100.93	100.41

Si	3.962	3.958	3.957	3.932	3.932	3.932
Al	0.809	0.806	0.819	0.848	0.846	0.823
Fe ³⁺						
Fe ²⁺	0.188	0.183	0.173	0.165	0.183	0.183
Mn	0.009					
Mg	1.032	1.044	1.048	1.037	1.015	1.046
Ba						
Sr						
Ca	3.263	3.318	3.293	3.301	3.268	3.297
K						0.008
Na	0.741	0.659	0.688	0.724	0.802	0.743

TOTAL	10.004	9.968	9.978	10.007	10.046	10.032
O	14	14	14	14	14	14

APPENDIX Jxiii MICA

C.AG.48A							olivine melteigite			C.AG.122			ijolite		C.ND.500	C.170.38			ijolite
	PT 56	PT 73	PT 75	PT 78	PT 91	PT 92		PT 2	PT 50		PT 1		PT 45	PT 48	PT 50				
SiO ₂	40.81	40.38	37.72	39.53	40.24	42.54	SiO ₂	40.39	40.62		41.33		38.27	39.07	38.85				
TiO ₂	0.94	0.65	3.47	2.06	1.16	0.18	TiO ₂	0.54	0.55		0.32		0.47	0.69	0.94				
Al ₂ O ₃	13.51	12.39	14.17	13.28	13.29	11.55	Al ₂ O ₃	12.76	12.10		14.26		15.25	15.62	15.74				
Fe ₂ O ₃		0.33	0.47	0.50		1.40	Fe ₂ O ₃		0.26										
FeO	6.61	9.39	9.54	9.00	9.67	3.15	FeO	8.02	8.61		8.05		5.23	6.25	6.21				
MnO	0.13	0.15	0.17	0.15	0.18	0.08	MnO	0.29	0.25		0.34		0.16	0.14	0.13				
MgO	23.86	22.09	19.52	21.46	21.38	26.97	MgO	22.88	22.26		23.92		24.77	24.1	23.86				
CaO	0.00	0.08	0.00	0.00	0.00	0.00	CaO		0.07				0.12		0.07				
BaO	0.12	0.49	2.14	0.47	0.28	0.20	BaO	0.53	0.60				2.06	0.85	0.8				
Na ₂ O	1.21	1.27	1.20	1.04	1.28	1.38	Na ₂ O	0.25	0.30				0.8	0.57	0.65				
K ₂ O	9.10	8.55	8.10	9.01	8.73	8.86	K ₂ O	10.40	10.21		10.27		9.33	9.96	10.11				
H ₂ O	3.20	3.08	3.36	2.99	3.06	2.76	H ₂ O	2.71	2.81				4.18	4.23	4.23				
F	2.06	2.16	1.48	2.35	2.23	3.08	F	2.93	2.69										
Cl							Cl												
O=F+Cl	-0.69	-0.91	-0.62	-0.99	-0.94	-1.30	O=F+Cl	-1.23	-1.13										
TOTAL	100.68	100.07	100.66	100.80	100.56	100.71	TOTAL	100.47	100.18		98.49		100.64	101.48	101.59				
Si	5.787	5.849	5.509	5.692	5.796	5.949	Si	5.835	5.900		5.770		5.492	5.535	5.504				
²⁷ Al	2.213	2.115	2.439	2.254	2.204	1.904	²⁷ Al	2.165	2.071		2.230		2.508	2.465	2.496				
Fe ³⁺		0.036	0.051	0.054		0.148	Fe ³⁺		0.029										
²⁷ Al	0.045				0.520		²⁷ Al	0.007			0.117		0.071	0.143	0.132				
Ti	0.100	0.071	0.381	0.223	0.126	0.019	Ti	0.059	0.060		0.033		0.051	0.074	0.1				
Fe ²⁺	0.784	1.138	1.165	1.084	1.165	0.368	Fe ²⁺	0.969	1.046		0.940		0.628	0.74	0.736				
Mn	0.016	0.018	0.021	0.018	0.022	0.009	Mn	0.035	0.031		0.040		0.019	0.017	0.016				
Mg	5.044	4.770	4.250	4.607	4.590	5.622	Mg	4.927	4.820		4.978		5.299	5.089	5.04				
Ca		0.012					Ca		0.011				0.018		0.011				
Ba	0.007	0.028	0.122	0.027	0.016	0.011	Ba	0.030	0.034				0.116	0.047	0.044				
Na	0.333	0.357	0.340	0.290	0.357	0.374	Na	0.070	0.084				0.223	0.157	0.179				
K	1.646	1.580	1.509	1.655	1.604	1.581	K	1.917	1.892		1.830		1.708	1.8	1.827				
TOTAL	15.973	15.974	15.787	15.904	15.571	15.985	TOTAL	16.014	15.978		15.938		16.133	16.066	16.084				
O	22	22	22	22	22	22	O	22	22		22		22	22	22				

APPENDIX Jxiii MICA (cont.)

E.133.30 <u>melilitolite</u>					S.78.15 <u>melilitolite</u>				S.162.300 <u>melilitolite-ijolite contact</u>					
	PT 34	PT 39	PT 54	PT 57		PT 43	PT 45	PT 47		PT 15	PT 24	PT 25	PT 26	PT 31
SiO ₂	37.83	38.36	37.87	39.05	SiO ₂	37.75	37.91	37.92	SiO ₂	39	39.71	39.99	38.96	40.27
TiO ₂	1.13	1.67	1.90		TiO ₂	1.06	0.59	0.55	TiO ₂	0.87	0.24	0.33	1.59	0.14
Al ₂ O ₃	14.93	14.16	14.68	15.13	Al ₂ O ₃	13.61	14.18	14.49	Al ₂ O ₃	10.82	10.74	10.7	10.81	10.64
Fe ₂ O ₃					Fe ₂ O ₃				Fe ₂ O ₃	2.84	2.17	2.27	3.28	1.94
FeO	9.73	9.62	9.15	9.99	FeO	8.41	8.04	8.13	FeO	8.65	6.99	7.03	6.26	8.95
MnO	0.20	0.19	0.16	0.17	MnO	0.17	0.15	0.16	MnO	0.32	0.25	0.25	0.24	0.32
MgO	20.45	20.63	20.39	18.05	MgO	21.49	21.71	21.52	MgO	21.4	22.62	22.81	22.18	21.86
CaO	0.07	0.10	0.06	2.58	CaO	0.14	0.33	0.17	CaO	0.06	0.52	0.1	0.09	
BaO	1.62	0.82	0.88		BaO	0.52	0.21	0.34	BaO	0.3	0.13		0.34	0.16
Na ₂ O	0.35	0.37	0.50	0.97	Na ₂ O	0.24	0.21	0.20	Na ₂ O	0.33	0.36	0.33	0.31	0.27
K ₂ O	9.93	9.93	9.58	9.42	K ₂ O	10.88	10.88	10.63	K ₂ O	10.42	10.27	10.78	10.67	10.7
H ₂ O	4.09	3.68	3.70	4.09	H ₂ O	3.78	3.80	4.05	H ₂ O	3.86	4.03	3.82	3.85	4.05
F		0.88	0.80		F	0.53	0.52		F	0.32		0.48	0.38	
Cl					Cl				Cl					
O=F+Cl		-0.37	-0.34		O=F+Cl	-0.22	-0.22		O=F+Cl	-0.13		-0.2	-0.16	
TOTAL	100.33	100.04	99.33	99.45	TOTAL	98.36	98.31	98.16	TOTAL	99.06	98.03	98.69	98.8	99.3
Si	5.551	5.613	5.564	5.729	Si	5.616	5.614	5.615	Si	5.79	5.883	5.891	5.754	5.936
²⁷ Al	2.549	2.387	2.436	2.281	²⁷ Al	2.384	2.386	2.385	²⁷ Al	1.893	1.875	1.858	1.882	1.848
Fe ³⁺					Fe ³⁺				Fe ³⁺	0.317	0.242	0.251	0.364	0.215
²⁷ Al	0.033	0.055	0.106	0.335	²⁷ Al	0.002	0.089	0.144	²⁷ Al					
Ti	0.125	0.184	0.210		Ti	0.119	0.066	0.061	Ti	0.097	0.027	0.037	0.177	0.016
Fe ²⁺	1.194	1.177	1.124	1.226	Fe ²⁺	1.046	0.996	1.007	Fe ²⁺	1.074	0.867	0.866	0.773	1.104
Mn	0.025	0.024	0.020	0.021	Mn	0.021	0.019	0.020	Mn	0.04	0.031	0.031	0.03	0.04
Mg	4.473	4.500	4.466	3.947	Mg	4.766	4.793	4.750	Mg	4.736	4.995	5.009	4.883	4.804
Ca	0.011	0.016	0.009	0.406	Ca	0.022	0.052	0.027	Ca	0.01	0.083	0.016	0.014	
Ba	0.093	0.047	0.051		Ba	0.030	0.012	0.020	Ba	0.017	0.008		0.02	0.009
Na	0.100	0.105	0.142	0.276	Na	0.069	0.060	0.057	Na	0.095	0.103	0.094	0.089	0.077
K	1.859	1.854	1.796	1.763	K	2.065	2.055	2.008	K	1.973	1.941	2.026	2.01	2.012
TOTAL	16.013	15.961	15.924	15.983	TOTAL	16.140	16.142	16.094	TOTAL	16.042	16.055	16.079	15.996	16.061
O	22	22	22	22	O	22	22	22	O	22	22	22	22	22

APPENDIX Jxiii MICA (cont.)

E.133.30 <u>melilitolite</u>				
	PT 34	PT 39	PT 54	PT 57
SiO ₂	37.83	38.36	37.87	39.05
TiO ₂	1.13	1.67	1.90	
Al ₂ O ₃	14.93	14.16	14.68	15.13
Fe ₂ O ₃				
FeO	9.73	9.62	9.15	9.99
MnO	0.20	0.19	0.16	0.17
MgO	20.45	20.63	20.39	18.05
CaO	0.07	0.10	0.06	2.58
BaO	1.62	0.82	0.88	
Na ₂ O	0.35	0.37	0.50	0.97
K ₂ O	9.93	9.93	9.58	9.42
H ₂ O	4.09	3.68	3.70	4.09
F		0.88	0.80	
Cl				
O=F+Cl		-0.37	-0.34	
TOTAL	100.33	100.04	99.33	99.45

WKT <u>turjaite</u>				
	PT 3	PT 17	PT 18	PT 24
SiO ₂	36.97	37.57	36.71	36.52
TiO ₂	2.04	1.32	1.47	1.71
Al ₂ O ₃	15.96	15.49	16.27	16.66
Fe ₂ O ₃				
FeO	7.57	7.13	7.33	7.50
MnO	0.16	0.17	0.17	0.19
MgO	21.91	22.13	21.79	21.31
CaO		0.14		0.06
BaO	0.97	1.06	1.25	1.33
Na ₂ O	0.24	0.20	0.25	0.25
K ₂ O	10.08	10.28	10.21	9.98
H ₂ O	3.74	3.79	3.73	3.80
F	0.82	0.68	0.77	0.63
Cl				
O=F+Cl	-0.35	-0.29	-0.32	-0.27
TOTAL	100.11	99.68	99.63	99.68

Si	5.372	5.477	5.370	5.340
²⁶ Al	2.628	2.523	2.630	2.660
Fe ³⁺				
²⁷ Al	0.105	0.138	0.175	0.211
Ti	0.223	0.145	0.162	0.188
Fe ²⁺	0.920	0.869	0.897	0.917
Mn	0.020	0.021	0.021	0.024
Mg	4.746	4.809	4.752	4.645
Ca		0.022		0.009
Ba	0.055	0.061	0.072	0.076
Na	0.068	0.057	0.071	0.071
K	1.869	1.912	1.905	1.862
TOTAL	16.006	16.032	16.054	16.003
O	22	22	22	22

S.78.15 <u>melilitolite</u>			
	PT 43	PT 45	PT 47
SiO ₂	37.75	37.91	37.92
TiO ₂	1.06	0.59	0.55
Al ₂ O ₃	13.61	14.18	14.49
Fe ₂ O ₃			
FeO	8.41	8.04	8.13
MnO	0.17	0.15	0.16
MgO	21.49	21.71	21.52
CaO	0.14	0.33	0.17
BaO	0.52	0.21	0.34
Na ₂ O	0.24	0.21	0.20
K ₂ O	10.88	10.88	10.63
H ₂ O	3.78	3.80	4.05
F	0.53	0.52	
Cl			
O=F+Cl	-0.22	-0.22	
TOTAL	98.36	98.31	98.16

TUR I <u>turjaite</u>				
	PT 2	PT 23	PT 51	PT 52
SiO ₂	37.09	36.54	39.21	36.83
TiO ₂	2.44	2.67	0.44	2.91
Al ₂ O ₃	15.65	15.66	14.64	15.61
Fe ₂ O ₃				
FeO	7.27	7.45	6.38	7.82
MnO	0.12	0.08	0.12	0.13
MgO	21.52	21.28	23.59	20.89
CaO		0.05		0.25
BaO	1.09	1.42	1.05	1.52
Na ₂ O	0.60	0.61	0.67	0.69
K ₂ O	9.59	9.17	9.44	9.21
H ₂ O	3.76	3.73	3.80	3.88
F	0.75	0.75	0.77	0.49
Cl				
O=F+Cl	-0.32	-0.32	-0.32	-0.21
TOTAL	99.56	99.09	99.78	100.02

Si	5.407	5.364	5.651	5.372
²⁶ Al	2.593	2.636	2.349	2.628
Fe ³⁺				
²⁷ Al	0.096	0.074	0.138	0.055
Ti	0.268	0.295	0.048	0.319
Fe ²⁺	0.886	0.915	0.769	0.954
Mn	0.015	0.010	0.015	0.016
Mg	4.677	4.657	5.068	4.542
Ca		0.008		0.039
Ba	0.062	0.082	0.059	0.087
Na	0.170	0.174	0.187	0.195
K	1.784	1.717	1.736	1.714
TOTAL	15.957	15.932	16.019	15.922
O	22	22	22	22

S.162.300 <u>melilitolite-ijorte contact</u>					
	PT 15	PT 24	PT 25	PT 26	PT 31
SiO ₂	39	39.71	39.99	38.96	40.27
TiO ₂	0.87	0.24	0.33	1.59	0.14
Al ₂ O ₃	10.82	10.74	10.7	10.81	10.64
Fe ₂ O ₃	2.84	2.17	2.27	3.28	1.94
FeO	8.65	6.99	7.03	6.26	8.95
MnO	0.32	0.25	0.25	0.24	0.32
MgO	21.4	22.62	22.81	22.18	21.86
CaO	0.06	0.52	0.1	0.09	
BaO	0.3	0.13		0.34	0.16
Na ₂ O	0.33	0.36	0.33	0.31	0.27
K ₂ O	10.42	10.27	10.78	10.67	10.7
H ₂ O	3.86	4.03	3.82	3.85	4.05
F	0.32		0.48	0.38	
Cl					
O=F+Cl	-0.13		-0.2	-0.16	
TOTAL	99.06	98.03	98.69	98.8	99.3

TUR II <u>"copiamelilitite"</u>					
	PT 19	PT 30	PT 38	PT 61	PT 64
SiO ₂	37.73	37.39	37.25	38.27	37.86
TiO ₂	0.94	2.09	0.20	1.38	0.69
Al ₂ O ₃	16.02	16.06	17.81	16.17	17.30
Fe ₂ O ₃					
FeO	5.97	6.00	6.07	6.06	6.64
MnO			0.13	0.12	0.13
MgO	23.34	22.36	23.08	22.65	22.45
CaO			0.16		
BaO	1.98	1.11		1.07	0.35
Na ₂ O	0.91	0.86	0.66		
K ₂ O	9.47	9.27	9.92	9.74	10.07
H ₂ O					
F	na	na	na	na	na
Cl					
O=F+Cl					
TOTAL	96.36	95.14	95.28	95.46	95.49

Si	5.436	5.418	5.356	5.511	5.443
²⁶ Al	2.564	2.582	2.644	2.489	2.557
Fe ³⁺					
²⁷ Al	0.157	0.160	0.374	0.256	0.374
Ti	0.101	0.227	0.021	0.149	0.075
Fe ²⁺	0.720	0.727	0.730	0.730	0.798
Mn			0.016	0.015	0.016
Mg	5.014	4.830	4.946	4.863	4.811
Ca			0.024		
Ba	0.112	0.063		0.060	
Na	0.254	0.241	0.184		
K	1.741	1.713	1.819	1.780	1.847
TOTAL	16.099	15.961	16.114	15.853	15.921
O	22	22	22	22	22

APPENDIX Jxiii MICA (cont.)

C.K.7	phoscorite				C.3.93	phoscorite					C.18.26	phoscorite			
	PT 34	PT 53	PT 54	PT 55		PT 1	PT 3	PT 4	PT 6	PT 10		PT 33	PT 34	PT 36	PT 41
SiO ₂	40.87	41.86	41.95	41.98	SiO ₂	42.00	42.05	41.67	41.95	42.30	SiO ₂	39.16	39.10	40.45	39.46
TiO ₂	0.63	0.28	0.31	0.30	TiO ₂	0.27	0.31	0.19	0.30	0.29	TiO ₂	1.40	1.23	1.01	0.97
Al ₂ O ₃	12.29	9.84	9.97	9.89	Al ₂ O ₃	6.65	7.12	5.29	9.00	8.65	Al ₂ O ₃	13.47	14.19	11.85	13.36
Fe ₂ O ₃	0.84	3.41	2.94	3.31	Fe ₂ O ₃	7.24	6.38	8.93	4.11	4.38	Fe ₂ O ₃			0.96	
FeO	4.96	3.37	3.57	3.42	FeO	4.75	4.68	4.71	4.51	4.70	FeO	10.54	7.63	6.72	7.55
MnO	0.07	0.10	0.08	0.07	MnO	0.09	0.07	0.09	0.06	0.05	MnO	0.17	0.12	0.13	0.12
MgO	25.05	25.99	25.73	26.01	MgO	24.71	24.71	24.33	25.07	25.01	MgO	20.38	22.45	23.22	22.47
CaO		0.07			CaO	0.19	0.16	0.10	0.10	0.11	CaO		0.04	0.09	
BaO	0.21				BaO						BaO	0.30	0.43	0.15	0.45
Na ₂ O	1.09	0.90	1.02	1.01	Na ₂ O	0.41	0.44	0.38	0.39	0.42	Na ₂ O	1.06	1.16	1.06	1.06
K ₂ O	9.27	9.54	9.49	9.48	K ₂ O	9.74	9.65	10.03	10.12	9.99	K ₂ O	9.01	9.29	8.86	8.90
H ₂ O	4.21	4.20	4.21	4.20	H ₂ O						H ₂ O	4.10	4.15	3.95	4.11
F					F	0.86	0.91	0.87	1.20	1.16	F			0.36	
Cl					Cl						Cl				
O=F+Cl					O=F+Cl	-0.36	-0.38	-0.37	-0.51	-0.49	O=F+Cl			0.15	
TOTAL	99.49	99.56	99.27	99.67	TOTAL	96.55	96.10	96.22	96.30	96.57	TOTAL	99.59	99.79	98.96	98.45
Si	5.840	5.978	6.002	5.984	Si	6.078	6.090	6.103	6.031	6.066	Si	5.726	5.646	5.869	5.757
²⁷ Al	2.070	1.656	1.681	1.661	²⁷ Al	1.134	1.215	0.913	1.525	1.462	²⁷ Al	2.274	2.354	2.026	2.243
Fe ³⁺	0.090	0.366	0.316	0.355	Fe ³⁺	0.788	0.695	0.984	0.444	0.472	Fe ³⁺			0.104	
²⁷ Al					²⁷ Al						²⁷ Al	0.047	0.061		0.054
Ti	0.068	0.030	0.033	0.032	Ti	0.029	0.034	0.021	0.032	0.031	Ti	0.154	0.134	0.110	0.106
Fe ²⁺	0.593	0.402	0.428	0.407	Fe ²⁺	0.575	0.567	0.577	0.543	0.564	Fe ²⁺	1.289	0.921	0.815	0.921
Mn	0.008	0.012	0.010	0.008	Mn	0.011	0.009	0.011	0.007	0.006	Mn	0.021	0.015	0.016	0.015
Mg	5.336	5.533	5.488	5.527	Mg	5.331	5.335	5.312	5.373	5.346	Mg	4.442	4.833	5.023	4.887
Ca		0.011	0.000	0.000	Ca	0.029	0.025	0.016	0.015	0.017	Ca		0.006	0.014	
Ba	0.012				Ba						Ba	0.017	0.024	0.009	0.026
Na	0.302	0.249	0.283	0.279	Na	0.115	0.124	0.108	0.109	0.117	Na	0.300	0.325	0.298	0.300
K	1.690	1.738	1.732	1.724	K	1.798	1.783	1.874	1.856	1.828	K	1.681	1.711	1.640	1.657
TOTAL	16.009	15.975	15.973	15.977	TOTAL	15.888	15.877	15.919	15.935	15.909	TOTAL	15.951	16.030	15.924	15.966
O	22	22	22	22	O	22	22	22	22	22	O	22	22	22	22
C.19.163	phoscorite								C.TL.344	carb	C.XC	carbonatite			
	PT 1	PT 2	PT 4	PT 20	PT 21	PT 26	PT 27		PT 74			PT 4	PT 18	PT 20	
SiO ₂	41.93	42.19	42.84	41.79	42.11	41.51	41.19	SiO ₂	42.57	SiO ₂	39.92	39.70	39.70		
TiO ₂	0.30	0.15	0.20	0.34	0.14	0.43	0.44	TiO ₂	0.17	TiO ₂	0.50	0.50	0.49		
Al ₂ O ₃	11.46	6.97	6.69	11.37	6.90	10.46	12.35	Al ₂ O ₃	9.87	Al ₂ O ₃	11.62	11.58	11.54		
Fe ₂ O ₃	1.16	6.56	6.58	1.08	6.61	2.14	0.06	Fe ₂ O ₃	2.80	Fe ₂ O ₃	2.01	2.50	2.34		
FeO	3.11	2.90	3.42	3.62	3.13	5.57	4.79	FeO	3.23	FeO	5.64	4.80	5.27		
MnO						0.08		MnO		MnO	0.10	0.08	0.07		
MgO	26.23	25.88	25.56	25.77	25.71	24.25	25.17	MgO	26.18	MgO	24.43	24.99	24.57		
CaO				0.24	0.09			CaO	0.07	CaO					
BaO								BaO		BaO					
Na ₂ O	0.20	0.25	0.27	0.20	0.26	0.96	0.25	Na ₂ O	1.41	Na ₂ O	1.13	1.17	1.17		
K ₂ O	11.10	10.37	10.42	10.52	10.14	9.57	10.53	K ₂ O	8.90	K ₂ O	9.23	9.36	9.43		
H ₂ O	4.23	4.14	4.13	4.23	4.13	3.63	3.70	H ₂ O	4.23	H ₂ O	3.96	3.94	3.88		
F						1.11	1.06	F		F	0.32	0.36	0.48		
Cl								Cl		Cl					
O=F+Cl						-0.47	-0.45	O=F+Cl		O=F+Cl	-0.13	-0.15	-0.20		
TOTAL	99.72	99.41	100.11	99.16	99.22	99.24	99.09	TOTAL	99.43	TOTAL	98.73	98.83	98.74		
Si	5.957	6.099	6.156	5.970	6.101	5.989	5.906	Si	6.048	Si	5.793	5.750	5.768		
²⁷ Al	1.919	1.187	1.133	1.914	1.178	1.779	2.087	²⁷ Al	1.653	²⁷ Al	1.987	1.977	1.976		
Fe ³⁺	0.124	0.714	0.711	0.116	0.721	0.232	0.006	Fe ³⁺	0.299	Fe ³⁺	0.220	0.272	0.256		
²⁷ Al								²⁷ Al		²⁷ Al					
Ti	0.032	0.016	0.022	0.037	0.015	0.047	0.047	Ti	0.018	Ti	0.055	0.054	0.054		
Fe ²⁺	0.369	0.350	0.411	0.432	0.379	0.672	0.574	Fe ²⁺	0.384	Fe ²⁺	0.685	0.582	0.641		
Mn						0.010	0.000	Mn		Mn	0.012	0.010	0.009		
Mg	5.555	5.577	5.475	5.488	5.553	5.216	5.380	Mg	5.545	Mg	5.285	5.396	5.322		
Ca				0.037	0.014			Ca	0.011	Ca					
Ba								Ba		Ba					
Na	0.055	0.070	0.075	0.055	0.073	0.269	0.070	Na	0.388	Na	0.318	0.329	0.330		
K	2.012	1.912	1.910	1.917	1.874	1.762	1.926	K	1.613	K	1.709	1.730	1.748		
TOTAL	16.023	15.925	15.893	15.966	15.908	15.976	15.996	TOTAL	15.959	TOTAL	16.064	16.100	16.104		
O	22	22	22	22	22	22	22	O	22	O	22	22	22		

APPENDIX Jxiii MICA (cont.)

C.6.15						C.18.60						C.23.270b					
carbonatite						carbonatite						carbonatite					
	PT 5	PT 10	PT 11	PT 18	PT 19		PT 3	PT 8	PT 9	PT 16		PT 53	PT 55	PT 58			
SiO ₂	41.91	42.34	42.49	41.96	40.16	SiO ₂	42.18	41.22	40.29	40.37	SiO ₂	39.52	39.49	40.09			
TiO ₂	0.14	0.14	0.11	0.24	1.02	TiO ₂	0.81	1.10	1.15	0.92	TiO ₂	0.67	0.50	0.47			
Al ₂ O ₃	11.22	12.01	9.01	12.57	12.14	Al ₂ O ₃	10.79	11.57	11.22	12.14	Al ₂ O ₃	10.88	10.59	11.02			
Fe ₂ O ₃	1.59	0.69	4.45	0.38	0.70	Fe ₂ O ₃	0.91	0.85	0.95	0.67	Fe ₂ O ₃	2.82	3.11	2.61			
FeO	1.52	2.15	1.34	2.49	9.62	FeO	8.87	7.24	11.82	7.80	FeO	3.48	3.04	3.10			
MnO					0.19	MnO	0.13		0.15	0.07	MnO	0.10	0.07	0.05			
MgO	27.36	27.39	27.41	27.18	21.41	MgO	21.96	22.79	19.36	22.62	MgO	24.75	25.05	25.36			
CaO		0.05	0.10	0.07		CaO	0.22	0.06	0.06	0.12	CaO		0.07	0.06			
BaO		0.15		0.17	0.17	BaO					BaO	0.23	0.17	0.13			
Na ₂ O	0.32	0.33	0.43	0.35	0.71	Na ₂ O	0.68	0.74	0.69	0.76	Na ₂ O	0.98	0.99	1.02			
K ₂ O	10.78	10.71	10.44	10.82	9.81	K ₂ O	9.42	9.87	9.89	9.79	K ₂ O	10.01	9.95	10.20			
H ₂ O	3.89	3.98	3.94	3.91	3.90	H ₂ O	3.77	3.76	3.75	3.80	H ₂ O	4.05	3.87	3.96			
F	0.67	0.61	0.51	0.76	0.45	F	0.82	0.83	0.65	0.69	F		0.36	0.30			
Cl						Cl					Cl						
O=F+Cl	-0.28	-0.26	-0.21	-0.32	-0.19	O=F+Cl	-0.35	-0.35	-0.27	-0.29	O=F+Cl		-0.15	-0.13			
TOTAL	99.12	100.29	100.02	100.58	100.09	TOTAL	100.21	99.68	99.71	99.46	TOTAL	97.49	97.11	98.24			
Si	5.952	5.941	6.021	5.883	5.842	Si	6.071	5.942	5.944	5.852	Si	5.805	5.817	5.827			
²⁶ Al	1.878	1.986	1.505	2.077	2.081	²⁶ Al	1.830	1.966	1.951	2.074	²⁶ Al	1.884	1.838	1.888			
Fe ³⁺	0.169	0.073	0.475	0.040	0.077	Fe ³⁺	0.098	0.092	0.105	0.074	Fe ³⁺	0.311	0.345	0.285			
²⁷ Al						²⁷ Al					²⁷ Al						
Ti	0.015	0.015	0.012	0.025	0.112	Ti	0.088	0.119	0.128	0.100	Ti	0.074	0.055	0.051			
Fe ²⁺	0.181	0.252	0.159	0.292	1.170	Fe ²⁺	1.068	0.873	1.458	0.946	Fe ²⁺	0.427	0.375	0.377			
Mn					0.023	Mn	0.016		0.019	0.009	Mn	0.012	0.009	0.006			
Mg	5.793	5.729	5.790	5.681	4.643	Mg	4.712	4.898	4.258	4.888	Mg	5.420	5.501	5.495			
Ca		0.008	0.015	0.011		Ca	0.034	0.009	0.009	0.019	Ca		0.011	0.009			
Ba		0.008		0.009	0.010	Ba					Ba	0.013	0.010	0.007			
Na	0.088	0.090	0.118	0.095	0.200	Na	0.190	0.207	0.197	0.214	Na	0.279	0.283	0.287			
K	1.953	1.917	1.887	1.935	1.820	K	1.730	1.815	1.861	1.811	K	1.876	1.870	1.891			
TOTAL	16.029	16.019	15.982	16.048	15.978	TOTAL	15.837	15.921	15.930	15.987	TOTAL	16.101	16.114	16.123			
O	22	22	22	22	22	O	22	22	22	22	O	22	22	22			
C.23.300						C.49.40											
carbonatite						carbonatite											
	PT 43	PT 45	PT 47	PT 49			PT 34	PT 36	PT 39	PT 45	PT 52	PT 53	PT 54	PT 55	PT 56	PT 60	
SiO ₂	41.49	42.34	41.27	40.86	SiO ₂	43.04	42.11	40.85	42.38	41.29	42.92	42.40	42.30	42.55	42.78		
TiO ₂	0.52	0.50	0.56	0.38	TiO ₂	0.31	0.52	1.71	0.46	0.54	0.35	0.22	0.22	0.29	0.16		
Al ₂ O ₃	11.92	11.16	12.46	12.40	Al ₂ O ₃	10.86	11.88	11.18	11.98	12.94	10.68	2.74	3.14	11.17	7.53		
Fe ₂ O ₃	0.82	1.32	0.56	0.83	Fe ₂ O ₃	1.59	0.83	1.19	0.64		1.86	11.80	11.69	1.03	5.89		
FeO	3.94	4.02	4.15	3.65	FeO	2.84	3.66	10.42	3.85	4.53	2.89	4.16	3.73	4.14	2.81		
MnO	0.07	0.07	0.07	0.06	MnO			0.18				0.06	0.07	0.06			
MgO	25.64	25.65	25.57	25.85	MgO	26.75	26.19	20.06	26.26	25.58	26.42	24.58	25.03	25.75	26.25		
CaO	0.13	0.05			CaO						0.15	0.26	0.23	0.23	0.13		
BaO	0.21		0.32	0.54	BaO						0.30						
Na ₂ O	1.18	1.23	1.45	1.27	Na ₂ O	0.30	0.31	0.51	0.34	0.30	0.26	0.28	0.28	0.26	0.28		
K ₂ O	8.97	9.07	9.12	9.47	K ₂ O	10.60	10.47	10.02	10.42	10.64	10.69	9.23	9.41	10.37	10.33		
H ₂ O	3.98	4.22	4.22	4.20	H ₂ O	4.09	3.71	4.10	3.94	4.23	3.79	4.03	3.98	4.23	3.86		
F	0.46				F		0.99		0.62		0.89				0.60		
Cl					Cl		0.14						0.17				
O=F+Cl	-0.19				O=F+Cl		-0.45		-0.26		-0.37					-0.25	
TOTAL	99.14	99.63	99.75	99.51	TOTAL	100.38	100.36	100.22	100.63	100.05	100.83	99.76	100.25	100.08	100.37		
Si	5.911	5.996	5.856	5.827	Si	6.037	5.938	5.950	5.950	5.851	6.033	6.223	6.176	6.026	6.102		
²⁶ Al	2.001	1.863	2.084	2.084	²⁶ Al	1.795	1.974	1.919	1.982	2.149	1.769	0.474	0.540	1.864	1.266		
Fe ³⁺	0.088	0.141	0.060	0.089	Fe ³⁺	0.167	0.088	0.130	0.068		0.197	1.303	1.284	0.110	0.632		
²⁷ Al					²⁷ Al					0.011							
Ti	0.056	0.053	0.060	0.041	Ti	0.033	0.055	0.187	0.049	0.058	0.037	0.024	0.024	0.023	0.017		
Fe ²⁺	0.470	0.476	0.493	0.436	Fe ²⁺	0.334	0.432	1.269	0.453	0.537	0.340	0.511	0.456	0.490	0.335		
Mn	0.008	0.008	0.008	0.007	Mn		0.022					0.007	0.009	0.007	0.000		
Mg	5.445	5.415	5.409	5.495	Mg	5.594	5.505	4.356	5.496	5.403	5.537	5.378	5.448	5.436	5.582		
Ca	0.020	0.008			Ca		0.000	0.000	0.000	0.000	0.023	0.041	0.036	0.035	0.020		
Ba	0.012		0.018	0.030	Ba						0.017						
Na	0.326	0.338	0.399	0.351	Na	0.082	0.085	0.144	0.093	0.082	0.071	0.080	0.079	0.071	0.077		
K	1.630	1.639	1.651	1.723	K	1.897	1.883	1.862	1.866	1.923	1.917	1.728	1.753	1.873	1.880		
TOTAL	15.967	15.937	16.038	16.083	TOTAL	15.939	15.960	15.839	15.957	16.014	15.941	15.769	15.805	15.935	15.911		
O	22	22	22	22	O	22	22	22	22	22	22	22	22	22	22	22	

APPENDIX Jxiii MICA (cont.)

C.100.a carbonatite					C.100.b					C.100.d carbonatite					C.100.y carbonatite				
	PT 58	PT 61	PT 62	PT 63		PT 11		PT 20	PT 21	PT 31		PT 28	PT 32	PT 33	PT 41				
SiO ₂	42.39	39.23	38.76	38.13		40.31	SiO ₂	42.43	41.50	42.08	SiO ₂	40.05	40.92	40.78	40.98				
TiO ₂	0.09	1.63	1.50	1.60		0.83	TiO ₂	0.08	0.16	0.12	TiO ₂	1.08	1.32	1.12	0.61				
Al ₂ O ₃	11.21	12.39	12.04	12.14		11.18	Al ₂ O ₃	11.56	11.55	11.32	Al ₂ O ₃	11.17	9.78	10.47	11.19				
Fe ₂ O ₃	1.95	0.86	1.09	1.76		1.26	Fe ₂ O ₃	0.93	1.27	1.55	Fe ₂ O ₃	1.73	2.70	2.06	0.94				
FeO	1.63	13.09	12.73	11.87		9.25	FeO	2.12	2.14	1.89	FeO	8.92	9.90	9.31	8.88				
MnO		0.21	0.20	0.20		0.10	MnO				MnO	0.05	0.13	0.05					
MgO	27.74	18.87	18.74	19.18		21.53	MgO	27.30	26.87	27.31	MgO	21.38	20.27	21.01	22.04				
CaO	0.12			0.12		0.11	CaO				CaO				0.14				
BaO	0.18	0.13					BaO	0.18	0.27	0.19	BaO								
Na ₂ O	0.28	0.78	0.80	0.88		0.64	Na ₂ O	0.25	0.28	0.25	Na ₂ O	0.80	0.76	0.75	0.63				
K ₂ O	10.69	9.36	9.50	9.50		9.39	K ₂ O	10.70	10.77	10.83	K ₂ O	9.95	9.89	10.01	9.64				
H ₂ O	3.75	3.89	3.76	3.66		3.74	H ₂ O	3.96	4.00	4.00	H ₂ O	3.55	3.61	3.63	3.51				
F	1.08	0.39	0.55	0.74		0.69	F	0.59	0.41	0.12	F	1.09	0.95	0.94	1.26				
Cl							Cl				Cl								
O=F+Cl	-0.45	-0.16	-0.23	-0.31		-0.29	O=F+Cl	-0.25	-0.17	-0.20	O=F+Cl	-0.46	-0.40	-0.40	-0.53				
TOTAL	100.66	100.67	99.44	99.47		98.74	TOTAL	99.85	99.05	99.46	TOTAL	99.31	99.83	99.73	99.29				
Si	5.943	5.761	5.766	5.674		5.924	Si	5.981	5.922	5.949	Si	5.877	6.009	5.968	5.974				
²⁷ Al	1.852	2.144	2.111	2.129		1.936	²⁷ Al	1.921	1.942	1.886	²⁷ Al	1.932	1.693	1.806	1.923				
Fe ³⁺	0.205	0.095	0.122	0.196		0.140	Fe ³⁺	0.098	0.136	0.165	Fe ³⁺	0.191	0.298	0.226	0.103				
²⁶ Al							²⁶ Al				²⁶ Al								
Ti	0.009	0.180	0.168	0.179		0.092	Ti	0.008	0.017	0.013	Ti	0.119	0.146	0.123	0.067				
Fe ²⁺	0.191	1.608	1.583	1.477		1.137	Fe ²⁺	0.250	0.256	0.223	Fe ²⁺	1.095	1.216	1.140	1.083				
Mn		0.026	0.025	0.025		0.012	Mn				Mn	0.006	0.016	0.006					
Mg	5.797	4.131	4.156	4.255		4.717	Mg	5.737	5.716	5.756	Mg	4.677	4.438	4.584	4.790				
Ca	0.018			0.019		0.017	Ca				Ca				0.022				
Ba	0.010	0.007					Ba	0.010	0.015	0.011	Ba								
Na	0.076	0.222	0.231	0.254		0.182	Na	0.068	0.077	0.069	Na	0.228	0.216	0.213	0.178				
K	1.912	1.753	1.803	1.804		1.760	K	1.924	1.961	1.953	K	1.863	1.853	1.869	1.793				
TOTAL	16.088	15.962	16.011	16.085		15.917	TOTAL	15.997	16.042	16.025	TOTAL	15.988	15.885	15.935	15.933				
O	22	22	22	22		22	O	22	22	22	O	22	22	22	22				
N.63.4 carb					N.64.19 carbonatite					N.187.100 carbonatite									
	PT 84		PT 52	PT 54	PT 55	PT 61	PT 78		PT 2	PT 9	PT 13	PT 17	PT 19		PT 12				
SiO ₂	38.98	SiO ₂	39.48	39.95	40.04	40.27	39.52	SiO ₂	39.25	36.42	35.99	35.73	39.07		36.67				
TiO ₂	2.09	TiO ₂	0.50	0.38	0.37	0.22	0.49	TiO ₂		0.70	0.81	0.73	0.09		0.75				
Al ₂ O ₃	13.62	Al ₂ O ₃	14.11	13.71	13.86	13.92	14.37	Al ₂ O ₃	10.59	13.31	13.87	13.37	11.07		10.04				
Fe ₂ O ₃	0.5	Fe ₂ O ₃						Fe ₂ O ₃	2.17	0.69		0.31	2.30		0.28				
FeO	8.97	FeO	5.57	5.84	5.51	5.58	6.41	FeO	2.51	4.37	7.35	7.69	1.89		26.67				
MnO	0.16	MnO	0.22	0.25	0.20	0.21	0.28	MnO	0.58	0.15	0.51	0.30	0.53		0.13				
MgO	20.85	MgO	24.55	24.47	24.74	24.76	23.78	MgO	25.05	23.55	21.32	21.14	25.47		4.63				
CaO	0.93	CaO	0.05	0.08	0.05			CaO	0.10										
BaO	0.46	BaO	0.50	0.34	0.24	0.30	0.44	BaO		0.42	0.28	0.15	0.24						
Na ₂ O	0.24	Na ₂ O	0.87	0.81	0.81	0.78	0.93	Na ₂ O	0.24	0.25	0.09	0.13	0.25		1.57				
K ₂ O	10.19	K ₂ O	9.36	9.46	9.67	9.64	9.44	K ₂ O	10.68	11.33	11.50	11.45	11.45		7.93				
H ₂ O	4.104	H ₂ O						H ₂ O	3.39	3.57	3.60	3.54	3.53		3.50				
F		F	0.39	0.38				F	1.23	0.77	0.66	0.71	1.01						
Cl		Cl						Cl											
O=F+Cl		O=F+Cl	-0.16	-0.16				O=F+Cl	-0.52	-0.32	-0.28	-0.30	-0.43						
TOTAL	101.09	TOTAL	95.44	95.51	95.49	95.68	95.66	TOTAL	95.27	95.21	95.70	94.95	96.47		92.17				
Si	5.628	Si	5.671	5.732	5.724	5.743	5.671	Si	5.884	5.536	5.513	5.527	5.805		6.282				
²⁷ Al	2.318	²⁷ Al	2.329	2.268	2.276	2.257	2.329	²⁷ Al	1.871	2.385	2.487	2.437	1.938		2.027				
Fe ³⁺	0.054	Fe ³⁺						Fe ³⁺	0.245	0.079		0.036	0.257						
²⁶ Al		²⁶ Al	0.060	0.051	0.059	0.083	0.101	²⁶ Al			0.017								
Ti	0.227	Ti	0.054	0.041	0.040	0.024	0.053	Ti		0.080	0.093	0.085	0.010		0.097				
Fe ²⁺	1.083	Fe ²⁺	0.669	0.701	0.659	0.666	0.769	Fe ²⁺	0.315	0.555	0.942	0.995	0.235		3.821				
Mn	0.020	Mn	0.027	0.030	0.024	0.025	0.034	Mn	0.074	0.019	0.066	0.039	0.067		0.019				
Mg	4.488	Mg	5.257	5.234	5.272	5.264	5.087	Mg	5.598	5.337	4.869	4.875	5.641		1.183				
Ca	0.144	Ca	0.008	0.012	0.008			Ca	0.016						0.051				
Ba	0.026	Ba	0.028	0.019	0.013	0.017	0.025	Ba		0.025	0.017	0.009	0.014						
Na	0.067	Na	0.242	0.225	0.225	0.216	0.259	Na	0.070	0.074	0.027	0.039	0.072		0.522				
K	1.877	K	1.715	1.732	1.764	1.754	1.728	K	2.043	2.197	2.247	2.259	2.170		1.733				
TOTAL	15.932	TOTAL	16.060	16.045	16.064	16.049	16.056	TOTAL	16.116	16.287	16.278	16.301	16.209		15.735				
O	22	O	22	22	22	22	22	O	22	22	22	22	22		22				

APPENDIX Jxiii MICA (cont.)

TU 119

turite

	PT 22	PT 23	PT 24	PT 25	PT 26	PT 30	PT 43	PT 44
SiO ₂	38.87	39.92	38.82	39.02	37.74	39.17	37.45	38.33
TiO ₂	2.00	0.20	1.93	1.26	2.23	0.69	2.45	1.94
Al ₂ O ₃	11.85	11.20	12.67	10.63	13.07	10.17	5.62	5.13
Fe ₂ O ₃	1.55	1.43	1.09	1.24	1.19	1.71	4.34	4.03
FeO	9.41	10.22	8.64	17.35	8.46	15.69	35.45	35.24
MnO	0.29	0.52	0.23	0.57	0.23	0.59	0.11	0.16
MgO	20.10	21.11	21.10	15.13	20.78	16.49	0.96	1.22
CaO	0.07	0.05	0.05	0.22		0.05	0.22	0.42
BaO	0.49	0.27	0.61	0.13	0.86			
Na ₂ O	0.42	0.28	0.45	0.19	0.47	0.07	0.12	0.09
K ₂ O	9.89	10.34	9.88	9.69	9.85	10.35	8.22	8.18
H ₂ O								
F								
Cl								
O=F+Cl								
TOTAL	94.94	95.54	95.47	95.43	94.88	94.98	94.94	94.74

Si	5.758	5.893	5.691	5.948	5.587	5.976	6.329	6.468
²⁷ Al	2.069	1.948	2.189	1.910	2.280	1.829	1.119	1.020
Fe ³⁺	0.173	0.159	0.120	0.142	0.132	0.196	0.551	0.512
²⁷ Al								
Ti	0.223	0.022	0.213	0.144	0.248	0.079	0.311	0.246
Fe ²⁺	1.166	1.262	1.059	2.212	1.048	2.002	5.010	4.973
Mn	0.036	0.065	0.029	0.074	0.029	0.076	0.016	0.023
Mg	4.439	4.645	4.611	3.438	4.586	3.750	0.242	0.307
Ca	0.011	0.008	0.008	0.036		0.008	0.040	0.076
Ba	0.028	0.016	0.035	0.008	0.050			
Na	0.121	0.080	0.128	0.056	0.135	0.021	0.039	0.029
K	1.869	1.947	1.848	1.884	1.860	2.014	1.772	1.761
TOTAL	15.893	16.045	15.931	15.852	15.955	15.951	15.429	15.415

O 22 22 22 22 22 22 22 22 22

TQ

turite chilled margin

	PT 28	PT 30	PT 33	PT 42	PT 45	PT 46	PT 63	PT 69	PT 70	PT 73	PT 80
SiO ₂	38.92	39.40	39.91	39.34	39.47	39.16	35.72	37.09	38.04	38.95	39.27
TiO ₂	1.04	1.00	2.40	2.31	2.29	0.20	2.29	1.86	1.12	0.13	0.50
Al ₂ O ₃	11.55	11.75	11.17	11.28	11.05	13.32	16.77	15.97	16.26	13.71	14.00
Fe ₂ O ₃	1.69	1.53	2.50	2.63	2.65						
FeO	10.31	10.37	8.74	8.56	9.15	8.08	9.70	9.10	8.72	10.78	10.84
MnO	0.75	0.68	0.31	0.31	0.34	0.76	0.19	0.17	0.17	0.29	0.31
MgO	20.21	20.47	20.49	20.59	20.19	22.64	19.63	20.41	21.05	20.74	20.54
CaO		0.07	0.06	0.04				0.09	0.12	0.12	
BaO	0.20	0.16	0.29	0.31	0.28	0.41	0.65	0.81	0.13	0.58	0.58
Na ₂ O	0.19	0.20	0.35	0.36	0.36	0.23	0.41	0.40	0.46	0.33	0.34
K ₂ O	10.33	10.51	10.20	10.18	10.09	10.52	9.72	9.81	9.92	10.02	10.12
H ₂ O	3.73	3.70	3.43	3.44	3.41	4.10	3.77	3.86	3.90	3.65	3.89
F	0.61	0.76	1.40	1.32	1.37		0.60	0.51	0.52	0.91	0.48
Cl											
O=F+Cl	-0.26	-0.32	-0.39	-0.56	-0.58		-0.25	-0.21	-0.22	-0.38	-0.20
TOTAL	99.27	100.28	100.66	100.11	100.07	99.42	99.20	99.87	100.19	99.83	100.67

Si	5.787	5.794	5.809	5.763	5.795	5.728	5.277	5.424	5.496	5.729	5.721
²⁷ Al	2.024	2.036	1.916	1.948	1.912	2.272	2.723	2.576	2.504	2.271	2.279
Fe ³⁺	0.189	0.169	0.274	0.289	0.293						
²⁷ Al						0.024	0.197	0.176	0.265	0.106	0.125
Ti	0.116	0.111	0.263	0.254	0.253	0.022	0.254	0.205	0.122	0.014	0.055
Fe ²⁺	1.282	1.276	1.064	1.049	1.124	0.988	1.198	1.113	1.054	1.326	1.321
Mn	0.094	0.085	0.038	0.038	0.042	0.094	0.024	0.021	0.021	0.036	0.038
Mg	4.480	4.488	4.446	4.497	4.419	4.937	4.323	4.449	4.534	4.547	4.461
Ca		0.011	0.009	0.006				0.014	0.019	0.019	
Ba	0.012	0.009	0.017	0.018	0.016	0.023	0.038	0.046	0.007	0.033	0.033
Na	0.055	0.057	0.099	0.102	0.102	0.065	0.117	0.113	0.129	0.094	0.096
K	1.959	1.972	1.894	1.903	1.890	1.963	1.832	1.830	1.828	1.880	1.881
TOTAL	15.998	16.008	15.829	15.867	15.846	16.116	15.983	15.967	15.977	16.056	16.010

O	22	22	22	22	22	22	22	22	22	22	22

TU 120A

turite

	PT 1	PT 2	PT 3	PT 4	PT 5	PT 6
SiO ₂	39.89	39.80	34.38	34.78	38.40	38.75
TiO ₂	0.81	0.18	3.30	3.38	1.50	1.36
Al ₂ O ₃	10.77	11.19	14.05	14.45	12.77	11.17
Fe ₂ O ₃	1.00		0.49	0.54		1.60
FeO	11.61	13.29	13.04	10.51	12.09	12.29
MnO	0.75	0.72	0.55	0.31	0.55	0.62
MgO	19.14	18.45	16.11	18.34	18.82	18.59
CaO	0.16	0.14				0.10
BaO		0.29	1.42	1.26	0.66	
Na ₂ O						
K ₂ O	10.14	10.23	9.47	9.27	10.15	9.96
H ₂ O						
F	n.a.	n.a.	n.a.	n.a.	n.a.	n.a.
Cl						
O=F+Cl						
TOTAL	94.27	94.29	92.81	92.84	94.94	94.44

Si	5.983	6.011	5.361	5.329	5.747	5.836
²⁷ Al	1.904	1.989	2.582	2.609	2.252	1.983
Fe ³⁺	0.113		0.057	0.062		0.181
²⁷ Al		0.003				
Ti	0.091	0.020	0.387	0.389	0.169	0.154
Fe ²⁺	1.456	1.679	1.701	1.347	1.513	1.548
Mn	0.095	0.092	0.073	0.040	0.070	0.079
Mg	4.279	4.154	3.745	4.189	4.199	4.174
Ca	0.026	0.023				0.016
Ba		0.017	0.087	0.076	0.039	
Na						
K	1.940	1.971	1.884	1.812	1.938	1.914
TOTAL	15.887	15.959	15.877	15.853	15.927	15.885

O	22	22	22	22	22	22

TD2

aeg-aug-apatite vein

	PT 43	PT 47	PT 48
SiO ₂	39.41	41.02	42.52
TiO ₂	0.78	0.55	0.52
Al ₂ O ₃	10.23	10.30	9.82
Fe ₂ O ₃	1.29	0.68	0.50
FeO	13.77	11.15	8.97
MnO	0.94	0.97	0.88
MgO	17.15	19.17	20.57
CaO	0.19	0.09	
BaO			
Na ₂ O	0.12	0.14	0.25
K ₂ O	10.14	10.66	10.80
H ₂ O			
F	0.78	0.55	0.52
Cl	3.08	3.03	2.55
O=F+Cl	-0.74	-0.87	-1.36
TOTAL	97.14	97.44	96.54

Si	6.013	6.114	6.245
²⁷ Al	1.839	1.809	1.700
Fe ³⁺	0.148	0.076	0.055
²⁷ Al			
Ti	0.089	0.062	0.057
Fe ²⁺	1.757	1.390	1.102
Mn	0.121	0.122	0.109
Mg	3.901	4.260	4.504
Ca	0.031	0.014	
Ba			
Na	0.035	0.040	0.071
K	1.974	2.027	2.024
TOTAL	15.908	15.914	15.867

O	22	22	22

APPENDIX Jxiv NEPHELINE

C.90.29 <u>pyroxenite</u>				C.AG.48A <u>ol. melteigite</u>			C.AG.122 <u>melteigite</u>			C.ND.500 <u>ijolite</u>			
	PT 4	PT 16	PT 24		PT 25	PT 66		PT 41	PT 45		PT 5	PT 6	PT 15
SiO ₂	41.98	42.70	43.25	SiO ₂	41.64	41.87	SiO ₂	41.34	41.16	SiO ₂	42.23	42.09	42.09
Al ₂ O ₃	34.01	33.94	33.64	Al ₂ O ₃	34.20	33.39	Al ₂ O ₃	34.69	34.84	Al ₂ O ₃	35.06	35.32	35.21
FeO	1.22	0.81	0.92	FeO	0.76	0.80	FeO	0.69	0.73	FeO	1.03	0.91	0.85
MnO				MnO			MnO			MnO			
MgO				MgO			MgO			MgO	0.05	0.04	
CaO				CaO	0.25	0.15	CaO	0.06		CaO	0.08	0.12	0.07
SrO				SrO			SrO			SrO			
K ₂ O	6.84	6.12	5.91	K ₂ O	3.51	3.52	K ₂ O	6.87	6.97	K ₂ O	5.75	5.75	5.98
Na ₂ O	16.59	16.41	16.85	Na ₂ O	18.87	18.59	Na ₂ O	16.44	16.46	Na ₂ O	17.33	17.65	17.27
TOTAL	100.64	99.98	100.57	TOTAL	99.23	98.32	TOTAL	100.2	100.16	TOTAL	101.53	101.91	101.47
Si	1.014	1.030	1.037	Si	1.009	1.022	Si	1.002	0.998	Si	1.006	1.000	1.003
Al	0.968	0.965	0.951	Al	0.977	0.961	Al	0.991	0.996	Al	0.984	0.989	0.989
Fe ²⁺	0.025	0.016	0.018	Fe ²⁺	0.015	0.016	Fe ²⁺	0.014	0.015	Fe ²⁺	0.021	0.018	0.017
Mn				Mn			Mn			Mn			
Mg				Mg			Mg			Mg	0.002	0.001	
Ca				Ca	0.005	0.004	Ca	0.001		Ca	0.002	0.004	0.002
Sr				Sr			Sr			Sr			
K	0.211	0.188	0.181	K	0.109	0.109	K	0.212	0.216	K	0.175	0.174	0.181
Na	0.777	0.767	0.783	Na	0.886	0.880	Na	0.773	0.774	Na	0.800	0.813	0.796
TOTAL	2.995	2.966	2.970	TOTAL	3.001	2.993	TOTAL	2.995	2.999	TOTAL	2.989	2.999	2.983
O	4	4	4	O	4	4	O	4	4	O	4	4	4

C.170.38 <u>ijolite</u>			C.IJ.14 <u>ijolite</u>				E.123.106 <u>ijolite</u>			WKM <u>ijolite</u>			
	PT 51	PT 46		PT 36	PT 50	PT 54		PT 23	PT 28		PT 57	PT 69	PT 76
SiO ₂	41.72	40.90	SiO ₂	42.27	42.28	42.02	SiO ₂	42.02	40.88	SiO ₂	39.79	40.34	40.21
Al ₂ O ₃	31.54	34.28	Al ₂ O ₃	35.14	34.93	34.75	Al ₂ O ₃	33.73	33.63	Al ₂ O ₃	36.48	36.34	36.57
FeO	0.26	0.57	FeO	1.11	1.09	1.18	FeO	0.70	1.01	FeO	1.28	0.85	1.28
MnO			MnO				MnO			MnO			
MgO			MgO				MgO			MgO	0.05	0.05	0.04
CaO	0.07	0.07	CaO	0.06		0.05	CaO	0.04	0.07	CaO	0.07	0.09	0.11
SrO			SrO	0.05	0.06	0.06	SrO			SrO			
K ₂ O	6.38	5.47	K ₂ O	6.87	6.95	6.78	K ₂ O	7.52	7.26	K ₂ O	7.77	7.44	7.69
Na ₂ O	17.56	18.00	Na ₂ O	16.88	16.74	16.57	Na ₂ O	16.15	16.21	Na ₂ O	16.00	16.03	15.84
TOTAL	97.53	99.29	TOTAL	102.38	102.05	101.41	TOTAL	100.16	99.06	TOTAL	101.44	101.14	101.74
Si	1.039	0.998	Si	1.004	1.007	1.007	Si	1.020	1.006	Si	0.960	0.972	0.966
Al	0.926	0.986	Al	0.983	0.981	0.981	Al	0.965	0.976	Al	1.038	1.031	1.035
Fe ²⁺	0.006	0.011	Fe ²⁺	0.022	0.022	0.024	Fe ²⁺	0.014	0.021	Fe ²⁺	0.026	0.017	0.026
Mn			Mn				Mn			Mn			
Mg			Mg				Mg			Mg	0.002	0.002	0.001
Ca	0.002	0.002	Ca	0.002		0.002	Ca	0.001	0.002	Ca	0.002	0.002	0.003
Sr			Sr	0.001	0.015	0.001	Sr			Sr			
K	0.203	0.170	K	0.208	0.211	0.207	K	0.233	0.228	K	0.239	0.229	0.236
Na	0.848	0.852	Na	0.777	0.773	0.770	Na	0.760	0.774	Na	0.749	0.749	0.737
TOTAL	3.023	3.020	TOTAL	2.997	2.995	2.991	TOTAL	2.994	3.007	TOTAL	3.015	3.001	3.003
O	4	4	O	4	4	4	O	4	4	O	4	4	4

APPENDIX Jxiv   NEPHELINE (cont.)

ij-fen	(contact)		C.190.x	(ijolite-fenite contact)			
	PT 54	PT 30		PT 22	PT 36	PT 39	PT 75
SiO ₂	41.95	42.72	SiO ₂	40.38	40.95	40.25	40.25
Al ₂ O ₃	32.25	33.02	Al ₂ O ₃	33.48	33.25	33.21	33.06
Fe ₂ O ₃	1.08	0.24	Fe ₂ O ₃	0.89	0.83	0.80	1.11
MnO			MnO				
MgO			MgO				
CaO			CaO	0.04	0.05		0.10
SrO	0.09		SrO	0.08	0.08		0.09
K ₂ O	7.42	7.27	K ₂ O	9.04	8.61	8.61	7.77
Na ₂ O	15.96	16.37	Na ₂ O	16.15	16.12	16.28	16.38
TOTAL	98.75	99.62	TOTAL	100.06	99.89	99.15	98.76
Si	1.033	1.038	Si	0.994	1.006	0.997	0.999
Al	0.936	0.946	Al	0.971	0.963	0.970	0.967
Fe ³⁺	0.020	0.005	Fe ³⁺	0.017	0.015	0.015	0.021
Mn			Mn				
Mg			Mg				
Ca			Ca	0.001	0.001		0.003
Sr	0.001		Sr	0.001	0.001		0.001
K	0.233	0.225	K	0.284	0.270	0.272	0.246
Na	0.762	0.771	Na	0.771	0.768	0.782	0.788
TOTAL	2.986	2.985	TOTAL	3.039	3.024	3.037	3.025
O	4	4	O	4	4	4	4

APPENDIX Jxiv OLIVINE + MONTICELLITE

C.AG.48A <u>olivine melteigite</u>					N.64.18m <u>carb-mel. contact</u>				TUR II	N.64.19 <u>carbonatite</u>			
olivine	PT 35	PT 38	PT 60	PT 71	mont.	PT 7	PT 96	PT 99	PT 84	mont.	PT 50	PT 53	PT 68
SiO ₂	39.37	39.32	39.42	39.03	SiO ₂	35.64	36.87	36.94	37.60	SiO ₂	35.59	36.14	35.65
Al ₂ O ₃					Al ₂ O ₃					Al ₂ O ₃			
FeO	16.73	14.68	13.78	16.38	FeO	13.15	9.24	9.92	10.64	FeO	11.95	12.01	12.05
MnO	0.33	0.26	0.22	0.42	MnO	1.81	1.48	1.27	0.31	MnO	2.16	2.25	2.11
MgO	42.59	44.18	45.16	42.70	MgO	15.24	18.39	17.95	18.31	MgO	16.22	16.30	16.13
CaO	0.57	0.47	0.48	0.58	CaO	33.37	34.31	33.76	35.27	CaO	33.22	33.48	33.00
SrO					SrO					SrO	0.14	0.12	0.12
TOTAL	99.59	98.91	99.06	99.11	TOTAL	99.21	100.29	99.84	102.13	TOTAL	99.28	100.30	99.06
Si	1.002	0.999	0.995	0.998	Si	1.002	1.004	1.011	1.007	Si	0.997	1.001	1.000
Al					Al					Al			
Fe ²⁺	0.356	0.312	0.291	0.351	Fe ²⁺	0.309	0.210	0.227	0.238	Fe ²⁺	0.280	0.278	0.283
Mn	0.007	0.005	0.005	0.009	Mn	0.043	0.034	0.029	0.007	Mn	0.051	0.053	0.050
Mg	1.616	1.673	1.700	1.628	Mg	0.639	0.747	0.732	0.731	Mg	0.677	0.673	0.674
Ca	0.015	0.013	0.013	0.016	Ca	1.005	1.001	0.990	1.011	Ca	0.997	0.993	0.992
Sr					Sr					Sr	0.002	0.002	0.002
TOTAL	2.997	3.001	3.005	3.002	TOTAL	2.999	2.996	2.989	2.994	TOTAL	3.003	2.999	3.000
O	4	4	4	4	O	4	4	4	4	O	4	4	4

APPENDIX Jxvi Fe-Ti OXIDES

C.AG.347 <u>pyroxenite</u>				C.90.29 <u>pyroxenite</u>				C.AG.48A <u>ol. melteigite</u>						
PT 60	PT 67	PT 72		PT 34	PT 35	PT 45		PT 23	PT 24	PT 32	PT 50	PT 62	PT 69	
CaO		0.41	0.13	CaO	0.11	0.10		CaO						
MgO	3.86	1.19	2.80	MgO	1.98	2.00	1.64	MgO	5.88	4.93	6.39	0.97	0.84	1.96
MnO	0.66	0.81	0.56	MnO	0.67	0.70	0.69	MnO	0.49	0.64	0.55	1.11	0.99	0.55
FeO	31.24	29.19	30.38	FeO	31.85	31.95	31.89	FeO	27.24	31.93	28.89	37.55	39.05	35.10
Fe ₂ O ₃	55.10	65.93	59.66	Fe ₂ O ₃	57.72	57.62	59.43	Fe ₂ O ₃	23.29	37.29	35.79	50.50	46.72	44.43
Al ₂ O ₃	2.89	0.60	2.26	Al ₂ O ₃	1.95	2.01	1.84	Al ₂ O ₃	8.50	5.29	6.62	0.13	0.71	0.35
V ₂ O ₃	0.11	0.15	0.15	V ₂ O ₃	0.14	0.17	0.17	V ₂ O ₃	0.09	0.11	0.09	0.08	0.22	0.07
Cr ₂ O ₃	0.10		0.20	Cr ₂ O ₃	0.12	0.17	0.17	Cr ₂ O ₃	29.45	11.18	14.15		0.05	8.38
SiO ₂		0.33	0.09	SiO ₂	0.17		0.09	SiO ₂		0.09	0.11	0.18	0.17	0.14
TiO ₂	6.60	0.95	4.05	TiO ₂	4.63	4.92	3.95	TiO ₂	4.33	8.51	7.27	9.52	10.84	8.11
Nb ₂ O ₅				Nb ₂ O ₅				Nb ₂ O ₅						
TOTAL	100.56	99.56	100.28	TOTAL	99.34	99.63	99.87	TOTAL	99.27	99.97	99.86	100.04	99.59	99.09
Ca		0.017	0.005	Ca	0.004	0.004		Ca						
Mg	0.209	0.068	0.154	Mg	0.111	0.112	0.092	Mg	0.305	0.262	0.333	0.055	0.047	0.110
Mn	0.020	0.026	0.018	Mn	0.021	0.022	0.022	Mn	0.014	0.019	0.016	0.036	0.032	0.018
Fe ²⁺	0.951	0.929	0.939	Fe ²⁺	1.001	1.001	1.001	Fe ²⁺	0.794	0.950	0.845	1.187	1.236	1.108
Fe ³⁺	1.509	1.889	1.659	Fe ³⁺	1.632	1.624	1.679	Fe ³⁺	0.611	0.999	0.943	1.437	1.330	1.262
Al	0.124	0.027	0.098	Al	0.086	0.089	0.081	Al	0.349	0.222	0.273	0.006	0.032	0.016
V	0.003	0.005	0.004	V	0.004	0.005	0.005	V	0.003	0.003	0.003	0.002	0.007	0.002
Cr	0.003		0.006	Cr	0.004	0.005	0.005	CR	0.811	0.315	0.392		0.001	0.250
Si		0.013	0.003	Si	0.006		0.003	Si		0.003	0.004	0.007	0.006	0.005
Ti	0.181	0.027	0.113	Ti	0.131	0.139	0.111	Ti	0.113	0.228	0.191	0.271	0.308	0.230
Nb				Nb				Nb						
TOTAL	3.000	3.000	3.000	TOTAL	3.000	3.000	3.000	TOTAL	3.000	3.000	3.000	3.000	3.000	3.000
O	4	4	4	O	4	4	4	O	4	4	4	4	4	4

C.ND.500 <u>jiolite</u>			C.IJ.14 <u>urtite</u>			E.133.30 <u>jiolite</u>				WKM <u>melteigite</u>			
PT 8	PT 17		PT 8	PT 9		PT 33	PT 51	PT 56		PT 62	PT 63	PT 64	
CaO			CaO		0.10	CaO		0.37	0.05	CaO	0.07	0.06	0.06
MgO	0.63	0.97	MgO	0.09	0.13	MgO	1.17	1.22	1.22	MgO	0.57	0.84	
MnO	1.45	1.32	MnO	0.63	8.03	MnO	1.13	1.15	1.02	MnO	1.56	1.70	14.43
FeO	30.88	30.43	FeO	33.10	38.36	FeO	36.66	37.52	35.22	FeO	34.87	34.97	32.28
Fe ₂ O ₃	64.69	63.49	Fe ₂ O ₃	62.15		Fe ₂ O ₃	51.44	48.06	53.89	Fe ₂ O ₃	56.75	55.19	
Al ₂ O ₃	0.37	0.69	Al ₂ O ₃	0.14		Al ₂ O ₃	1.28	1.40	1.55	Al ₂ O ₃	0.60	0.61	
V ₂ O ₃	0.16	0.19	V ₂ O ₃	0.18	0.29	V ₂ O ₃	0.25	0.22	0.28	V ₂ O ₃	0.41	0.41	0.17
Ce ₂ O ₃			Ce ₂ O ₃		0.20	Ce ₂ O ₃				Ce ₂ O ₃			
SiO ₂	0.24	0.16	SiO ₂		0.07	SiO ₂	0.16	0.24	0.15	SiO ₂			0.12
TiO ₂	1.92	2.21	TiO ₂	3.20	53.79	TiO ₂	8.53	9.96	7.04	TiO ₂	6.34	7.12	52.88
Nb ₂ O ₅			Nb ₂ O ₅			Nb ₂ O ₅				Nb ₂ O ₅			0.13
TOTAL	100.34	99.46	TOTAL	99.49	100.97	TOTAL	100.62	100.15	100.42	TOTAL	101.17	100.90	100.07
Ca			Ca		0.003	Ca		0.011	0.002	Ca	0.003	0.002	0.002
Mg	0.036	0.055	Mg	0.005	0.005	Mg	0.065	0.068	0.068	Mg	0.032	0.047	
Mn	0.047	0.043	Mn	0.021	0.169	Mn	0.036	0.036	0.032	Mn	0.050	0.054	0.307
Fe ²⁺	0.982	0.972	Fe ²⁺	1.067	0.797	Fe ²⁺	1.145	1.174	1.101	Fe ²⁺	1.095	1.097	0.678
Fe ³⁺	1.851	1.824	Fe ³⁺	1.803		Fe ³⁺	1.445	1.353	1.516	Fe ³⁺	1.603	1.559	
Al	0.017	0.031	Al	0.006		Al	0.056	0.062	0.068	Al	0.027	0.027	
V	0.005	0.006	V	0.006	0.006	V	0.007	0.007	0.008	V	0.012	0.012	0.003
Ce			Cr		0.002	Ce				Ce			
Si	0.009	0.006	Si		0.002	Si	0.006	0.009	0.006	Si			0.003
Ti	0.055	0.063	Ti	0.093	1.006	Ti	0.239	0.280	0.198	Ti	0.179	0.201	0.999
Nb			Nb			Nb				Nb			0.001
TOTAL	3.000	3.000	TOTAL	3.000	1.989	TOTAL	3.000	3.000	3.000	TOTAL	3.000	3.000	1.994
O	4	4	O	4	3	O	4	4	4	O	4	4	3

APPENDIX Jxvi Fe-Ti OXIDES (cont.)

C.5.118 <u>melilitolite</u>				C.190.105 <u>melilitolite</u>				N.64.18c <u>melilitolite</u>				N.64.18 <u>melilitolite</u>			
PT 9	PT 10	PT 28		PT 58	PT 60	PT 68		PT 34	PT 44			PT 39	PT 49	PT 51	
CaO				CaO				CaO				CaO	0.13	0.05	
MgO	1.27	1.44	1.22	MgO	0.66	0.70	0.81	MgO	1.29	2.80		MgO	3.45	1.40	3.59
MnO	1.07	1.12	1.06	MnO	0.97	0.92	0.99	MnO	1.07	2.34		MnO	2.05	1.13	1.67
FeO	32.96	33.78	32.66	FeO	34.03	33.58	33.68	FeO	31.34	31.98		FeO	29.10	31.39	27.18
Fe ₂ O ₃	60.36	57.83	60.94	Fe ₂ O ₃	59.24	59.91	59.02	Fe ₂ O ₃	61.92	55.13		Fe ₂ O ₃	56.42	60.15	62.19
Al ₂ O ₃	0.77	0.96	0.81	Al ₂ O ₃	1.25	1.52	1.59	Al ₂ O ₃	0.99	2.86		Al ₂ O ₃	2.64	1.15	2.53
V ₂ O ₃	0.22	0.19	0.23	V ₂ O ₃	0.16	0.19	0.16	V ₂ O ₃	0.10	0.05		V ₂ O ₃	0.07	0.10	0.09
Ce ₂ O ₃				Ce ₂ O ₃				Ce ₂ O ₃				Ce ₂ O ₃			
SiO ₂	0.33	0.26	0.28	SiO ₂	0.06	0.09	0.11	SiO ₂	0.50	0.94		SiO ₂	0.11	0.10	0.07
TiO ₂	4.29	5.60	3.97	TiO ₂	4.69	4.13	4.51	TiO ₂	2.74	5.70		TiO ₂	5.44	3.91	2.98
Nb ₂ O ₅				Nb ₂ O ₅				Nb ₂ O ₅				Nb ₂ O ₅			
TOTAL	101.28	101.17	101.18	TOTAL	101.05	101.04	100.87	TOTAL	99.95	101.80		TOTAL	99.28	99.46	100.35
Ca				Ca				Ca				Ca	0.005	0.002	
Mg	0.071	0.080	0.068	Mg	0.037	0.039	0.045	Mg	0.073	0.151		Mg	0.190	0.079	0.196
Mn	0.034	0.035	0.034	Mn	0.031	0.029	0.031	Mn	0.034	0.072		Mn	0.064	0.036	0.052
Fe ²⁺	1.028	1.051	1.021	Fe ²⁺	1.067	1.051	1.054	Fe ²⁺	0.990	0.966		Fe ²⁺	0.901	0.995	0.834
Fe ³⁺	1.694	1.619	1.713	Fe ³⁺	1.671	1.688	1.663	Fe ³⁺	1.760	1.499		Fe ³⁺	1.572	1.715	1.718
Al	0.034	0.042	0.036	Al	0.055	0.067	0.070	Al	0.044	0.122		Al	0.115	0.051	0.109
V	0.007	0.006	0.007	V	0.005	0.006	0.005	V	0.003	0.001		V	0.002	0.003	0.003
Ce				Ce				Ce				Ce			
Si	0.012	0.010	0.010	Si	0.002	0.003	0.004	Si	0.019	0.034		Si	0.004	0.004	0.003
Ti	0.120	0.157	0.112	Ti	0.132	0.116	0.127	Ti	0.078	0.155		Ti	0.151	0.111	0.082
Nb				Nb				Nb				Nb			
TOTAL	3.000	3.000	3.000	TOTAL	3.000	3.000	3.000	TOTAL	3.000	3.000		TOTAL	3.000	3.000	3.000
O	4	4	4	O	4	4	4	O	4	4		O	4	4	4

S.78.15 <u>melilitolite</u>					WK T <u>turjaite</u>					TURI <u>turjaite</u>				
PT 23	PT 28	PT 29	PT 33		PT 2	PT 7	PT 13	PT 27		PT 56	PT 1	PT 2	PT 4	
CaO					CaO			0.06		CaO				0.13
MgO	2.85	2.72	2.99	1.78	MgO	1.58	1.85	1.62	2.11	MgO	1.49	1.46	1.54	1.31
MnO	1.05	1.14	1.12	1.26	MnO	1.33	1.30	1.29	1.16	MnO	0.91	1.11	0.84	0.76
FeO	32.67	31.85	32.69	33.77	FeO	31.93	31.60	31.65	30.85	FeO	29.96	30.12	31.52	30.56
Fe ₂ O ₃	47.08	51.43	49.40	52.16	Fe ₂ O ₃	58.80	59.79	58.41	59.39	Fe ₂ O ₃	63.23	62.92	61.72	64.32
Al ₂ O ₃	1.35	1.35	1.56	1.46	Al ₂ O ₃	1.80	0.80	1.60	1.90	Al ₂ O ₃	1.35	1.44	1.76	1.74
V ₂ O ₃	0.11	0.09	0.09	0.04	V ₂ O ₃	0.15	0.15	0.18	0.19	V ₂ O ₃	0.24	0.13	0.17	
Ce ₂ O ₃					Ce ₂ O ₃					Ce ₂ O ₃				
SiO ₂					SiO ₂	0.10	0.08	0.11	0.11	SiO ₂			0.28	0.18
TiO ₂	9.47	7.63	8.78	7.62	TiO ₂	4.48	4.72	4.54	4.10	TiO ₂	2.24	2.49	2.96	1.81
Nb ₂ O ₅					Nb ₂ O ₅					Nb ₂ O ₅				
TOTAL	94.60	96.21	96.63	98.09	TOTAL	100.17	100.29	99.46	99.81	TOTAL	99.42	99.67	100.79	100.81
Ca					Ca			0.002		Ca				0.005
Mg	0.166	0.156	0.170	0.101	Mg	0.088	0.103	0.091	0.118	Mg	0.084	0.082	0.085	0.073
Mn	0.035	0.037	0.036	0.041	Mn	0.042	0.041	0.041	0.037	Mn	0.029	0.036	0.026	0.024
Fe ²⁺	1.067	1.028	1.046	1.077	Fe ²⁺	1.000	0.991	0.998	0.965	Fe ²⁺	0.950	0.953	0.982	0.955
Fe ³⁺	1.384	1.493	1.422	1.496	Fe ³⁺	1.656	1.688	1.658	1.672	Fe ³⁺	1.805	1.791	1.730	1.808
Al	0.062	0.061	0.070	0.066	Al	0.079	0.035	0.071	0.084	Al	0.060	0.064	0.078	0.077
V	0.003	0.003	0.003	0.001	V	0.005	0.005	0.005	0.006	V	0.007	0.004	0.005	
Ce					Ce					Ce				
Si					Si	0.004	0.003	0.004	0.004	Si			0.011	0.007
Ti	0.278	0.221	0.253	0.218	Ti	0.126	0.133	0.129	0.115	Ti	0.064	0.071	0.083	0.051
Nb					Nb					Nb				
TOTAL	3.000	3.000	3.000	3.000	TOTAL	3.000	3.000	3.000	3.000	TOTAL	3.000	3.000	3.000	3.000
O	4	4	4	4	O	4	4	4	4	O	4	4	4	4

C.3.93phoscorite

	PT 20	PT 22	PT 24	PT 21	PT 23	PT 25
CaO						
MgO	0.54	0.81	0.66	4.58	6.48	5.82
MnO	0.17	0.27	0.18	3.60	3.34	2.67
FeO	31.62	31.32	31.82	36.76	34.47	34.91
Fe ₂ O ₃	68.69	67.90	67.72	1.05	1.15	4.01
Al ₂ O ₃				0.09		
V ₂ O ₃	0.51	0.51	0.49	0.11	0.06	0.02
Ce ₂ O ₃						
SiO ₂	0.14	0.18	0.15	0.17	0.12	0.14
TiO ₂	0.71	1.01	1.18	53.79	54.78	53.18
Nb ₂ O ₅						
TOTAL	102.38	102.00	102.20	100.16	100.39	100.75
Ca						
Mg	0.030	0.045	0.037	0.166	0.231	0.208
Mn	0.005	0.009	0.006	0.074	0.068	0.054
Fe ²⁺	0.990	0.981	0.996	0.748	0.690	0.701
Fe ³⁺	1.934	1.914	1.908	0.019	0.021	0.073
Al				0.003		
V	0.015	0.015	0.015	0.002	0.001	
Ce						
Si	0.005	0.007	0.006	0.004	0.003	0.003
Ti	0.020	0.028	0.033	0.984	0.986	0.960
Nb						
TOTAL	3.000	3.000	3.000	2.000	2.000	2.000
O	4	4	4	3	3	3

C.3.93phoscorite

	PT 20	PT 22	PT 24	PT 21	PT 23	PT 25
CaO						
MgO	0.54	0.81	0.66	4.58	6.48	5.82
MnO	0.17	0.27	0.18	3.60	3.34	2.67
FeO	31.62	31.32	31.82	36.76	34.47	34.91
Fe ₂ O ₃	68.69	67.90	67.72	1.05	1.15	4.01
Al ₂ O ₃				0.09		
V ₂ O ₃	0.51	0.51	0.49	0.11	0.06	0.02
Ce ₂ O ₃						
SiO ₂	0.14	0.18	0.15	0.17	0.12	0.14
TiO ₂	0.71	1.01	1.18	53.79	54.78	53.18
Nb ₂ O ₅						
TOTAL	102.38	102.00	102.20	100.16	100.39	100.75
Ca						
Mg	0.030	0.045	0.037	0.166	0.231	0.208
Mn	0.005	0.009	0.006	0.074	0.068	0.054
Fe ²⁺	0.990	0.981	0.996	0.748	0.690	0.701
Fe ³⁺	1.934	1.914	1.908	0.019	0.021	0.073
Al				0.003		
V	0.015	0.015	0.015	0.002	0.001	
Ce						
Si	0.005	0.007	0.006	0.004	0.003	0.003
Ti	0.020	0.028	0.033	0.984	0.986	0.960
Nb						
TOTAL	3.000	3.000	3.000	2.000	2.000	2.000
O	4	4	4	3	3	3

C.19.163phoscorite

	PT 8	PT 10	PT 15	PT 7	PT 14	PT 16
CaO						
MgO	0.34	0.22	0.60	2.98	3.87	0.90
MnO	0.15	0.10	0.21	2.03	2.10	2.17
FeO	32.99	32.63	32.54	40.28	39.77	42.34
Fe ₂ O ₃	64.94	64.58	63.85	1.45		2.41
Al ₂ O ₃			0.06			
V ₂ O ₃	0.58	0.46	0.55	0.13	0.18	0.15
Cr ₂ O ₃	0.06	0.07	0.07		0.17	
SiO ₂	0.73	0.74	1.08	0.64	0.93	0.84
TiO ₂	1.38	1.14	1.21	52.14	53.26	50.20
Nb ₂ O ₅						
TOTAL	101.17	99.94	100.18	99.66	100.28	99.01
Ca						
Mg	0.019	0.013	0.034	0.110	0.141	0.034
Mn	0.005	0.003	0.007	0.043	0.043	0.046
Fe ²⁺	1.043	1.045	1.035	0.833	0.811	0.895
Fe ³⁺	1.847	1.861	1.827	0.027		0.046
Al			0.003			
V	0.018	0.014	0.017	0.003	0.004	0.003
Cr	0.002	0.002	0.002		0.002	
Si	0.028	0.028	0.041	0.016	0.023	0.021
Ti	0.039	0.033	0.035	0.969	0.976	0.954
Nb						
TOTAL	3.000	3.000	3.000	2.000	1.999	2.000
O	4	4	4	3	3	3

C.19.163phoscorite

	PT 8	PT 10	PT 15	PT 7	PT 14	PT 16
CaO						
MgO	0.34	0.22	0.60	2.98	3.87	0.90
MnO	0.15	0.10	0.21	2.03	2.10	2.17
FeO	32.99	32.63	32.54	40.28	39.77	42.34
Fe ₂ O ₃	64.94	64.58	63.85	1.45		2.41
Al ₂ O ₃			0.06			
V ₂ O ₃	0.58	0.46	0.55	0.13	0.18	0.15
Cr ₂ O ₃	0.06	0.07	0.07		0.17	
SiO ₂	0.73	0.74	1.08	0.64	0.93	0.84
TiO ₂	1.38	1.14	1.21	52.14	53.26	50.20
Nb ₂ O ₅						
TOTAL	101.17	99.94	100.18	99.66	100.28	99.01
Ca						
Mg	0.019	0.013	0.034	0.110	0.141	0.034
Mn	0.005	0.003	0.007	0.043	0.043	0.046
Fe ²⁺	1.043	1.045	1.035	0.833	0.811	0.895
Fe ³⁺	1.847	1.861	1.827	0.027		0.046
Al			0.003			
V	0.018	0.014	0.017	0.003	0.004	0.003
Cr	0.002	0.002	0.002		0.002	
Si	0.028	0.028	0.041	0.016	0.023	0.021
Ti	0.039	0.033	0.035	0.969	0.976	0.954
Nb						
TOTAL	3.000	3.000	3.000	2.000	1.999	2.000
O	4	4	4	3	3	3

APPENDIX Jxvi Fe-Ti OXIDES (cont.)

C.18.26 <u>phoscorite</u>				C.TL.344 <u>carbonatite</u>							C.XC <u>carbonatite</u>			
PT 31	PT 39	PT 43		PT 64	PT 65	PT 75	PT 63	PT 68	PT 77		PT 10	PT 13	PT 24	
CaO				CaO	0.39						CaO			0.05
MgO	0.85	0.97	1.52	MgO	1.50	9.61	9.38	1.41	2.24	1.56	MgO	0.96	0.79	0.11
MnO	0.96	0.97	1.10	MnO	4.25	5.41	4.22	0.52	0.91	0.57	MnO	0.47	0.36	0.15
FeO	34.27	35.15	33.95	FeO	40.01	27.26	29.94	29.77	33.23	30.24	FeO	32.15	31.04	34.71
Fe ₂ O ₃	56.62	55.49	55.53	Fe ₂ O ₃	1.09			65.57	57.06	66.53	Fe ₂ O ₃	63.19	67.32	60.91
Al ₂ O ₃	0.63	0.13	0.42	Al ₂ O ₃			0.00	0.15	0.20	0.39	Al ₂ O ₃	0.16	0.18	
V ₂ O ₃	0.04	0.10	0.03	V ₂ O ₃	0.18	0.23	0.15	0.18	0.17	0.16	V ₂ O ₃	0.28	0.31	0.31
Ce ₂ O ₃				Ce ₂ O ₃		0.21	0.15				Ce ₂ O ₃			
SiO ₂				SiO ₂	0.52	0.52	0.31	0.61	0.67	0.32	SiO ₂			
TiO ₂	6.02	7.09	6.94	TiO ₂	52.12	56.69	56.69	0.92	5.92	1.42	TiO ₂	3.15	1.25	4.21
Nb ₂ O ₅				Nb ₂ O ₅							Nb ₂ O ₅			
TOTAL	99.39	99.90	99.49	TOTAL	100.06	99.93	100.84	99.13	100.40	101.20	TOTAL	100.36	101.24	100.45
Ca				Ca	0.010						Ca			0.002
Mg	0.048	0.055	0.086	Mg	0.056	0.335	0.326	0.080	0.125	0.087	Mg	0.054	0.044	0.006
Mn	0.031	0.031	0.035	Mn	0.090	0.107	0.083	0.017	0.029	0.018	Mn	0.015	0.012	0.005
Fe ²⁺	1.093	1.116	1.077	Fe ²⁺	0.832	0.533	0.583	0.953	1.038	0.947	Fe ²⁺	1.02	0.98	1.108
Fe ³⁺	1.625	1.586	1.585	Fe ³⁺	0.020			1.888	1.604	1.874	Fe ³⁺	1.805	1.912	1.749
Al	0.028	0.006	0.019	Al				0.007	0.009	0.017	Al	0.007	0.008	
V	0.001	0.003	0.001	V	0.004	0.004	0.003	0.006	0.005	0.005	V	0.009	0.009	0.009
Ce				Ce		0.002	0.001				Ce			
Si				Si	0.013	0.012	0.007	0.023	0.025	0.012	Si			
Ti	0.173	0.203	0.198	Ti	0.975	0.996	0.993	0.026	0.166	0.040	Ti	0.09	0.035	0.121
Nb				Nb							Nb			
TOTAL	3.000	3.000	3.000	TOTAL	2.000	1.989	1.997	3.000	3.000	3.000	TOTAL	3	3	3
O	4	4	4	O	3	3	3	4	4	4	O	4	4	4

C.XC (cont) <u>carbonatite</u>				C.6.15 <u>carbonatite</u>								C.18.60 <u>carbonatite</u>		
PT 9	PT 12	PT 14		PT 14	PT 22	PT 2	PT 13	PT 16	PT 20	PT 23		PT 2	PT 15	
CaO	0.05			CaO								CaO	0.26	0.22
MgO	5.55	6.71	5.05	MgO	10.48	11.57	1.34	1.46	1.28	0.92	1.49	MgO	0.13	0.09
MnO	3.85	5.1	4.48	MnO	2.60	2.27	0.49	0.33	0.43	0.39	0.28	MnO	0.06	
FeO	34.25	32.97	35.1	FeO	29.82	27.98	31.85	30.94	31.84	32.09	30.60	FeO	31.01	30.99
Fe ₂ O ₃	3.03	0.03	1.26	Fe ₂ O ₃	0.84	1.50	62.98	64.00	63.32	62.90	64.45	Fe ₂ O ₃	67.37	67.24
Al ₂ O ₃				Al ₂ O ₃			0.13	0.17	0.16	0.09	0.18	Al ₂ O ₃		
V ₂ O ₃	0.23	0.15	0.18	V ₂ O ₃	0.54	0.56	0.25	0.22	0.23	0.31	0.29	V ₂ O ₃	0.75	0.59
Ce ₂ O ₃			0.15	Ce ₂ O ₃								Ce ₂ O ₃		
SiO ₂			0.1	SiO ₂								SiO ₂		
TiO ₂	53.29	55.51	53.96	TiO ₂	56.87	56.61	3.44	2.70	3.25	3.12	2.38	TiO ₂	0.79	0.50
Nb ₂ O ₅	0.17	0.17		Nb ₂ O ₅								Nb ₂ O ₅		
TOTAL	100.42	100.64	100.28	TOTAL	101.15	100.49	100.48	99.81	100.51	99.81	99.67	TOTAL	100.37	99.64
Ca	0.001			Ca								Ca	0.014	0.009
Mg	0.2	0.239	0.182	Mg	0.361	0.398	0.075	0.083	0.072	0.052	0.085	Mg	0.007	0.005
Mn	0.079	0.103	0.092	Mn	0.051	0.044	0.016	0.011	0.014	0.013	0.009	Mn	0.002	
Fe ²⁺	0.691	0.658	0.712	Fe ²⁺	0.576	0.540	1.007	0.984	1.006	1.025	0.975	Fe ²⁺	0.992	1.000
Fe ³⁺	0.055	0.001	0.023	Fe ³⁺	0.015	0.026	1.791	1.831	1.801	1.807	1.847	Fe ³⁺	1.939	1.953
Al				Al			0.006	0.008	0.007	0.004	0.008	Al		
V	0.004	0.003	0.003	V	0.010	0.010	0.008	0.007	0.007	0.009	0.009	V	0.023	0.018
Ce			0.001	Ce								Ce		
Si			0.002	Si								Si		
Ti	0.967	0.996	0.984	Ti	0.987	0.982	0.098	0.077	0.092	0.090	0.068	Ti	0.023	0.015
Nb	0.002	0.002		Nb								Nb		
TOTAL	2	2	2	TOTAL	2.000	2.000	3.000	3.000	3.000	3.000	3.000	TOTAL	3.000	3.000
O	3	3	3	O	3	3	4	4	4	4	4	O	4	4

APPENDIX Jxvi Fe-Ti OXIDES (cont.)

C.23.270b <u>carbonatite</u>					C.23.300 <u>carbonatite</u>					C.49.40 <u>carbonatite</u>				
	PT 59	PT 61	PT 60	PT 62		PT 42	PT 53	PT 54	PT 44		PT 31	PT 43	PT 48	
CaO					CaO	0.16				CaO	0.08			
MgO	9.12	10.23	1.26	1.18	MgO	0.80	1.77	2.41	12.04	MgO	0.15	0.44	1.45	
MnO	4.25	5.75	0.45	0.42	MnO	0.28	0.42	1.03	4.25	MnO	0.12	0.11	0.39	
FeO	29.38	26.37	29.36	29.66	FeO	30.75	28.82	33.09	25.23	FeO	33.35	31.53	31.85	
Fe ₂ O ₃	0.81	0.16	66.06	66.17	Fe ₂ O ₃	64.61	67.15	54.27		Fe ₂ O ₃	64.94	66.94	63.60	
Al ₂ O ₃			0.27	0.33	Al ₂ O ₃		0.33	0.06		Al ₂ O ₃				
V ₂ O ₃	0.50	0.52	0.18	0.17	V ₂ O ₃	0.18	0.22	0.16	0.69	V ₂ O ₃	0.25	0.24	0.30	
Ce ₂ O ₃					Ce ₂ O ₃					Ce ₂ O ₃			0.18	
SiO ₂					SiO ₂			0.17		SiO ₂	0.29	0.12	0.16	
TiO ₂	55.54	56.08	1.15	1.17	TiO ₂	1.95	1.03	7.50	57.83	TiO ₂	2.12	1.06	3.14	
Nb ₂ O ₅					Nb ₂ O ₅					Nb ₂ O ₅				
TOTAL	99.60	99.11	98.73	99.10	TOTAL	98.72	99.75	98.70	100.04	TOTAL	101.30	100.44	101.06	
Ca					Ca	0.007				Ca	0.003			
Mg	0.322	0.359	0.072	0.068	Mg	0.046	0.100	0.136	0.413	Mg	0.008	0.025	0.081	
Mn	0.085	0.115	0.015	0.014	Mn	0.009	0.014	0.033	0.083	Mn	0.004	0.004	0.012	
Fe ²⁺	0.581	0.520	0.946	0.953	Fe ²⁺	0.995	0.916	1.051	0.485	Fe ²⁺	1.056	1.006	1.001	
Fe ³⁺	0.014	0.003	1.916	1.912	Fe ³⁺	1.881	1.920	1.551		Fe ³⁺	1.850	1.923	1.799	
Al			0.012	0.015	Al		0.015	0.003		Al				
V	0.009	0.010	0.006	0.005	V	0.006	0.007	0.005	0.013	V	0.008	0.007	0.009	
Ce					Ce					Ce			0.002	
Si					Si			0.006		Si	0.011	0.005	0.006	
Ti	0.988	0.994	0.033	0.034	Ti	0.057	0.029	0.214	1.000	Ti	0.060	0.030	0.089	
Nb					Nb					Nb				
TOTAL	2.000	2.000	3.000	3.000	TOTAL	3.000	3.000	3.000	1.994	TOTAL	3.000	3.000	3.000	
O	3	3	4	4	O	4	4	4	3	O	4	4	4	

C.100.b <u>carbonatite</u>						C.100.d <u>carbonatite</u>				C.100.y <u>carbonatite</u>					
	PT 4	PT 14	PT 3	PT 6	PT 15		PT 19	PT 29		PT 19	PT 23	PT 26	PT 18	PT 25	
CaO			0.15			CaO		0.35	CaO	0.23	0.17	0.14			
MgO	1.56	3.43	0.19	0.38	0.38	MgO	1.81	0.24	MgO	0.06	0.07		0.55	0.56	
MnO	1.41	1.65	0.07	0.09	0.13	MnO	0.23	0.06	MnO				0.70	0.77	
FeO	42.45	40.02	32.11	31.92	32.40	FeO	29.36	30.73	FeO	31.80	31.65	31.39	43.37	43.86	
Fe ₂ O ₃	2.73	2.28	65.27	66.14	65.23	Fe ₂ O ₃	66.10	68.16	Fe ₂ O ₃	65.32	65.96	66.91	5.60	3.42	
Al ₂ O ₃				0.09	0.07	Al ₂ O ₃	0.25	0.05	Al ₂ O ₃						
V ₂ O ₃	0.42	0.58	0.45	0.31	0.30	V ₂ O ₃	0.21	0.33	V ₂ O ₃	0.73	0.75	0.58	0.44	0.47	
Ce ₂ O ₃	0.18	0.19				Ce ₂ O ₃			Ce ₂ O ₃						
SiO ₂						SiO ₂			SiO ₂						
TiO ₂	51.27	52.08	1.99	1.51	2.04	TiO ₂	1.56	0.45	TiO ₂	1.37	1.09	0.66	46.76	48.84	
Nb ₂ O ₅	0.51	0.90				Nb ₂ O ₅			Nb ₂ O ₅				2.79	1.59	
TOTAL	100.52	101.13	100.23	100.44	100.56	TOTAL	99.52	100.37	TOTAL	99.51	99.69	99.67	100.21	99.51	
Ca			0.009			Ca		0.014	Ca	0.010	0.007	0.006			
Mg	0.058	0.125	0.011	0.022	0.022	Mg	0.103	0.014	Mg	0.003	0.004		0.021	0.021	
Mn	0.030	0.034	0.002	0.003	0.004	Mn	0.007	0.002	Mn				0.015	0.017	
Fe ²⁺	0.885	0.818	1.027	1.019	1.033	Fe ²⁺	0.935	0.983	Fe ²⁺	1.027	1.021	1.013	0.922	0.934	
Fe ³⁺	0.051	0.042	1.879	1.900	1.871	Fe ³⁺	1.893	1.962	Fe ³⁺	1.898	1.914	1.944	0.107	0.066	
Al				0.004	0.003	Al	0.011	0.002	Al						
V	0.008	0.011	0.014	0.009	0.009	V	0.006	0.010	V	0.023	0.023	0.018	0.009	0.010	
Ce	0.002	0.002				Ce			Ce						
Si						Si			Si						
Ti	0.961	0.958	0.057	0.043	0.058	Ti	0.045	0.013	Ti	0.040	0.032	0.019	0.894	0.935	
Nb	0.006	0.010				Nb			Nb				0.032	0.018	
TOTAL	2.000	2.000	3.000	3.000	3.000	TOTAL	3.000	3.000	TOTAL	3.000	3.000	3.000	2.000	2.000	
O	3	3	4	4	4	O	4	4	O	4	4	4	3	3	

APPENDIX Jxvi Fe-Ti OXIDES (cont.)

N.63.4 <u>carbonatite</u>			N.64.19 <u>carbonatite</u>			TU 120A <u>turjiite</u>			TD1		TD 2	
PT 71		PT 72	PT 72		PT 75	PT 33		PT 34	PT 55		PT 49	
CaO			CaO			CaO			CaO	0.27	CaO	0.52
MgO	1.54	1.47	MgO	1.90	2.11	MgO			MgO		MgO	
MnO	1.97	1.95	MnO	2.24	2.44	MnO	4.17	0.63	MnO		MnO	
FeO	30.62	30.31	FeO	30.68	30.06	FeO	38.25	45.34	FeO	0.53	FeO	1.37
Fe ₂ O ₃	60.05	60.97	Fe ₂ O ₃	62.92	60.63	Fe ₂ O ₃			Fe ₂ O ₃		Fe ₂ O ₃	
Al ₂ O ₃	1.06	0.92	Al ₂ O ₃	1.26	2.05	Al ₂ O ₃			Al ₂ O ₃	0.06	Al ₂ O ₃	
V ₂ O ₃	0.12	0.07	V ₂ O ₃	0.05	0.06	V ₂ O ₃			V ₂ O ₃	1.72	V ₂ O ₃	0.76
Ce ₂ O ₃			Ce ₂ O ₃			Ce ₂ O ₃			Ce ₂ O ₃	0.28	Ce ₂ O ₃	0.25
SiO ₂	0.07		SiO ₂	0.14	0.07	SiO ₂			SiO ₂	0.96	SiO ₂	0.36
TiO ₂	4.10	3.78	TiO ₂	3.77	4.10	TiO ₂	51.92	51.67	TiO ₂	89.43	TiO ₂	96.61
Nb ₂ O ₅			Nb ₂ O ₅			Nb ₂ O ₅			Nb ₂ O ₅	5.51	Nb ₂ O ₅	2.34
TOTAL	99.53	99.47	TOTAL	102.96	101.51	TOTAL	94.34	97.64	TOTAL	98.65	TOTAL	102.15
Ca			Ca			Ca			Ca	0.003	Ca	0.001
Mg	0.087	0.083	Mg	0.103	0.116	Mg			Mg		Mg	
Mn	0.063	0.063	Mn	0.069	0.076	Mn	0.093	0.014	Mn		Mn	
Fe ²⁺	0.969	0.962	Fe ²⁺	0.936	0.924	Fe ²⁺	0.845	0.979	Fe ²⁺	0.006	Fe ²⁺	0.015
Fe ³⁺	1.710	1.741	Fe ³⁺	1.727	1.678	Fe ³⁺			Fe ³⁺		Fe ³⁺	
Al	0.047	0.041	Al	0.054	0.089	Al			Al	0.001	Al	
V	0.004	0.002	V	0.001	0.002	V			V	0.019	V	0.008
Ce			Ce			Ce			Ce	0.001	Ce	0.001
Si	0.003		Si	0.005	0.003	Si			Si	0.013	Si	0.005
Ti	0.117	0.108	Ti	0.103	0.113	Ti	1.031	1.003	Ti	0.923	Ti	0.959
Nb			Nb			Nb			Nb	0.034	Nb	0.014
TOTAL	3.000	3.000	TOTAL	3.000	3.000	TOTAL	1.969	1.996	TOTAL	1.001	TOTAL	1.009
O	4	4	O	4	4	O	3	3	O	2	O	2

APPENDIX Jxvii PEROVSKITE

C.AG.347				C.90.29				C.AG.48A ol.melt		C.AG.122				
pyroxenite				pyroxenite						ijolite				
PT 71	PT 78	PT 88		PT 33	PT 36	PT 46		PT 64		PT 26	PT 29	PT 30	PT 33	
Nb ₂ O ₅		0.30		Nb ₂ O ₅				Nb ₂ O ₅	1.05	Nb ₂ O ₅	0.30	0.58	0.77	0.87
TiO ₂	58.47	57.10	58.09	TiO ₂	56.77	56.71	57.10	TiO ₂	55.73	TiO ₂	56.30	55.96	54.01	54.44
SiO ₂		1.08		SiO ₂				SiO ₂	0.06	SiO ₂			0.24	
Al ₂ O ₃	0.08	0.21	0.18	Al ₂ O ₃	0.13	0.11	0.11	Al ₂ O ₃	0.09	Al ₂ O ₃	0.11	0.10	0.15	0.16
V ₂ O ₃	0.22	0.22	0.32	V ₂ O ₃				V ₂ O ₃	0.25	V ₂ O ₃	0.15	0.18	0.19	0.21
La ₂ O ₃			0.50	La ₂ O ₃	0.22			La ₂ O ₃	0.65	La ₂ O ₃	0.50	0.91	1.09	1.37
Ce ₂ O ₃	0.24	0.15	1.18	Ce ₂ O ₃	0.89	0.23	0.48	Ce ₂ O ₃	1.60	Ce ₂ O ₃	1.19	1.72	2.72	2.93
FeO	1.46	1.18	0.94	FeO	0.70	1.15	1.00	FeO	0.84	FeO	0.90	1.14	1.67	1.52
MgO		0.36		MgO				MgO		MgO				
SrO				SrO	0.40	0.41	0.35	SrO		SrO				
CaO	40.48	40.07	39.13	CaO	39.21	39.99	39.73	CaO	37.12	CaO	39.58	37.80	36.73	36.10
Na ₂ O	0.14	0.18	0.19	Na ₂ O	0.22	0.11	0.11	Na ₂ O	0.66	Na ₂ O	0.48	0.60	0.70	0.84
TOTAL	101.09	100.55	100.84	TOTAL	98.54	98.71	98.88	TOTAL	98.06	TOTAL	98.52	98.99	98.27	98.44
Nb	0.000	0.000	0.003	Nb				Nb	0.011	Nb	0.003	0.006	0.008	0.009
Ti	0.990	0.966	0.991	Ti	0.991	0.986	0.991	Ti	0.983	Ti	0.985	0.981	0.962	0.971
Si		0.024		Si				Si	0.001	Si			0.006	
Al	0.002	0.006	0.005	Al	0.004	0.003	0.003	Al	0.002	Al	0.003	0.003	0.004	0.004
V	0.004	0.004	0.006	V				V	0.005	V	0.003	0.003	0.004	0.004
La			0.004	La	0.002			La	0.006	La	0.004	0.008	0.010	0.012
Ce	0.002	0.001	0.010	Ce	0.008	0.002	0.004	Ce	0.014	Ce	0.010	0.015	0.024	0.025
Fe ²⁺	0.027	0.022	0.018	Fe ²⁺	0.014	0.022	0.019	Fe ²⁺	0.016	Fe ²⁺	0.018	0.022	0.033	0.030
Mg		0.012		Mg				Mg		Mg				
Sr				Sr	0.005	0.005	0.005	Sr		Sr				
Ca	0.977	0.966	0.951	Ca	0.975	0.991	0.982	Ca	0.933	Ca	0.962	0.944	0.933	0.917
Na	0.006	0.008	0.009	Na	0.010	0.005	0.005	Na	0.030	Na	0.022	0.027	0.032	0.039
TOTAL	2.009	2.008	1.996	TOTAL	2.008	2.014	2.008	TOTAL	2.001	TOTAL	2.011	2.009	2.015	2.012
C.IJ.14			WKM		C.5.118			C.190.105						
ijolite					melilitolite			melilitolite						
PT 27	PT 44		PT 61		PT 6	PT 11	PT 12	PT 53	PT 54	PT 56	PT 59	PT 77		
Nb ₂ O ₅	6.82	0.78	Nb ₂ O ₅	3.20	Nb ₂ O ₅	0.63	0.38	Nb ₂ O ₅	0.51	0.88	0.98	0.85	0.95	
TiO ₂	47.97	55.36	TiO ₂	54.07	TiO ₂	56.03	56.74	58.84	TiO ₂	55.55	55.50	55.40	55.35	54.57
SiO ₂			SiO ₂		SiO ₂	0.25	0.17	0.16	SiO ₂					
Al ₂ O ₃	0.10	0.11	Al ₂ O ₃	0.06	Al ₂ O ₃	0.18	0.21	0.04	Al ₂ O ₃	0.18	0.14	0.15	0.26	0.25
V ₂ O ₃	0.37	0.18	V ₂ O ₃	0.07	V ₂ O ₃	0.13	0.04	0.08	V ₂ O ₃			0.11	0.10	
La ₂ O ₃	6.95	1.08	La ₂ O ₃	0.28	La ₂ O ₃	0.91	0.91	0.22	La ₂ O ₃	0.86	0.99	0.80	0.91	0.90
Ce ₂ O ₃	8.35	2.27	Ce ₂ O ₃	1.32	Ce ₂ O ₃	2.17	2.37	0.44	Ce ₂ O ₃	1.85	2.10	1.79	2.00	2.14
FeO	0.31	1.27	FeO	1.27	FeO	1.20	1.33	0.66	FeO	1.04	1.24	1.20	1.54	1.52
MgO			MgO		MgO				MgO					
SrO			SrO	0.25	SrO				SrO	0.34	0.39	0.35	0.37	0.35
CaO	16.60	37.15	CaO	37.72	CaO	38.29	37.41	40.21	CaO	38.22	37.62	38.31	37.54	37.78
Na ₂ O	5.16	0.66	Na ₂ O	0.78	Na ₂ O	0.49	0.48	0.25	Na ₂ O	0.33	0.38	0.39	0.41	0.44
TOTAL	92.63	98.86	TOTAL	99.02	TOTAL	100.28	100.04	100.90	TOTAL	98.88	99.24	99.48	99.33	98.90
Nb	0.083	0.008	Nb	0.034	Nb	0.007	0.004	Nb	0.005	0.009	0.010	0.009	0.010	
Ti	0.967	0.976	Ti	0.950	Ti	0.971	0.985	0.996	Ti	0.977	0.975	0.970	0.972	0.965
Si			Si		Si	0.006	0.004	0.004	Si					
Al	0.002	0.003	Al	0.002	Al	0.005	0.006	0.001	Al	0.005	0.004	0.004	0.007	0.007
V	0.008	0.003	V	0.001	V	0.002	0.001	0.001	V			0.002	0.002	
La	0.069	0.009	La	0.002	La	0.008	0.008	0.002	La	0.007	0.009	0.007	0.008	0.008
Ce	0.082	0.019	Ce	0.011	Ce	0.018	0.020	0.004	Ce	0.016	0.018	0.015	0.017	0.018
Fe ³⁺	0.007	0.025	Fe ²⁺	0.025	Fe ²⁺	0.023	0.026	0.012	Fe ²⁺	0.020	0.024	0.023	0.030	0.030
Mg			Mg		Mg				Mg					
Sr			Sr	0.003	Sr				Sr	0.005	0.005	0.005	0.005	0.005
Ca	0.477	0.934	Ca	0.944	Ca	0.946	0.925	0.970	Ca	0.958	0.942	0.955	0.939	0.951
Na	0.268	0.030	Na	0.035	Na	0.022	0.021	0.011	Na	0.015	0.017	0.018	0.019	0.020
TOTAL	1.963	2.009	TOTAL	2.008	TOTAL	2.007	1.999	2.001	TOTAL	2.008	2.004	2.009	2.007	2.014
O	3	3	O	3	O	3	3	3	O	3	3	3	3	3

APPENDIX Jxvii PEROVSKITE (cont.)

N.18.64m						E.133.30					S.78.15					
calcite melilitolite						melilitolite					melilitolite					
	PT 28	PT 29	PT 35	PT 36	PT 52		PT 32	PT 44	PT 45	PT 49		PT 34	PT 49			
Nb ₂ O ₅	2.69	3.63	4.16	2.97		Nb ₂ O ₅		1.34	1.97	1.37	Nb ₂ O ₅	0.14	0.97			
TiO ₂	50.54	49.01	48.39	48.47	56.66	TiO ₂	56.77	53.91	52.89	54.15	TiO ₂	57.02	56.25			
SiO ₂	0.44	0.41	0.42	0.45	0.06	SiO ₂	0.08	0.11	0.09	0.07	SiO ₂					
Al ₂ O ₃	0.28	0.46	0.40	0.56	0.16	Al ₂ O ₃	0.10	0.10	0.13	0.12	Al ₂ O ₃	0.05	0.09			
V ₂ O ₃	0.03	0.09	0.07	0.10	0.10	V ₂ O ₃	0.15		0.13		V ₂ O ₃	0.35	0.41			
La ₂ O ₃	0.98	0.83	1.16	0.80		La ₂ O ₃	0.63	1.22	1.21	1.17	La ₂ O ₃		0.75			
Ce ₂ O ₃	2.97	3.25	3.53	2.98	0.15	Ce ₂ O ₃	1.10	2.72	3.00	2.29	Ce ₂ O ₃		1.22			
FeO	2.78	3.01	3.17	3.29	2.05	FeO	1.11	1.29	1.48	1.21	FeO	0.66	0.88			
MgO						MgO					MgO	0.05				
SrO					0.26	SrO	0.65	0.58	0.59	0.61	SrO	0.44	0.36			
CaO	35.77	35.08	34.98	35.04	40.69	CaO	39.39	36.37	35.62	36.46	CaO	40.22	38.97			
Na ₂ O	0.61	0.68	0.68	0.53	0.40	Na ₂ O	0.40	0.70	0.82	0.67	Na ₂ O	0.24	0.43			
TOTAL	97.09	96.45	96.96	95.19	100.53	TOTAL	100.38	98.34	97.93	98.12	TOTAL	99.17	100.33			
Nb	0.029	0.040	0.046	0.033		Nb		0.014	0.021	0.015	Nb	0.001	0.010			
Ti	0.920	0.902	0.891	0.903	0.971	Ti	0.980	0.964	0.953	0.967	Ti	0.985	0.972			
Si	0.011	0.010	0.010	0.011	0.001	Si	0.002	0.003	0.002	0.002	Si					
Al	0.008	0.013	0.012	0.016	0.004	Al	0.003	0.003	0.004	0.003	Al	0.001	0.002			
V	0.001	0.002	0.001	0.002	0.002	V	0.003		0.002		V	0.006	0.008			
La	0.009	0.007	0.010	0.007		La	0.005	0.011	0.011	0.010	La		0.006			
Ce	0.026	0.029	0.032	0.027	0.001	Ce	0.009	0.024	0.026	0.020	Ce		0.010			
Fe ²⁺	0.056	0.062	0.065	0.068	0.039	Fe ²⁺	0.021	0.026	0.030	0.024	Fe ²⁺	0.013	0.017			
Mg						Mg					Mg	0.002				
Sr					0.003	Sr	0.009	0.008	0.008	0.008	Sr	0.006	0.005			
Ca	0.928	0.920	0.918	0.930	0.993	Ca	0.968	0.926	0.915	0.928	Ca	0.990	0.959			
Na	0.029	0.032	0.032	0.025	0.018	Na	0.018	0.032	0.038	0.031	Na	0.011	0.019			
TOTAL	2.017	2.018	2.018	2.023	2.033	TOTAL	2.017	2.010	2.010	2.008	TOTAL	2.015	2.009			
O	3	3	3	3	3	O	3	3	3	3	O	3	3			
WK T	turjaite					TUR I	turjaite					TUR II	"copiamelilite"			
	PT 5	PT 6	PT 28			PT 1	PT 2	PT 4	PT 6			PT 6	PT 7	PT 9	PT 21	PT 74
Nb ₂ O ₅	1.28	1.15	0.61		Nb ₂ O ₅	0.67	0.62	0.57		Nb ₂ O ₅						1.01
TiO ₂	55.11	55.04	56.71		TiO ₂	54.50	54.23	54.68	56.68	TiO ₂	57.77	57.49	57.76	58.37	56.31	
SiO ₂		0.05	0.06		SiO ₂		0.18			SiO ₂	0.23		0.24	0.21	0.21	
Al ₂ O ₃	0.18	0.16	0.11		Al ₂ O ₃	0.52	0.32	0.32	0.33	Al ₂ O ₃	0.22	0.21		0.21	0.37	
V ₂ O ₃	0.01	0.06	0.03		V ₂ O ₃					V ₂ O ₃						
La ₂ O ₃	0.93	0.80	0.76		La ₂ O ₃	1.21	0.93	1.00	0.71	La ₂ O ₃						
Ce ₂ O ₃	1.97	1.49	1.47		Ce ₂ O ₃	3.74	3.52	3.34	2.19	Ce ₂ O ₃		0.46	0.41			1.74
FeO	1.35	1.28	1.10		FeO	1.76	1.74	1.64	1.25	FeO	1.78	0.96	1.05	0.79	1.35	
MgO					MgO					MgO						
SrO	0.26	0.37	0.36		SrO					SrO						
CaO	37.63	38.33	39.01		CaO	36.08	35.86	36.10	38.07	CaO	40.55	40.04	40.25	40.72	38.42	
Na ₂ O	0.43	0.44	0.33		Na ₂ O	0.55	0.78	0.62		Na ₂ O						
TOTAL	99.15	99.17	100.55		TOTAL	99.03	98.18	98.27	99.23	TOTAL	100.55	99.16	99.71	100.30	99.41	
Nb	0.014	0.012	0.006		Nb	0.007	0.007	0.006	0.000	Nb						0.011
Ti	0.969	0.966	0.978		Ti	0.968	0.969	0.975	0.989	Ti	0.982	0.992	0.990	0.991		0.976
Si		0.001	0.001		Si	0.000	0.004	0.000	0.000	Si	0.005	0.000	0.005	0.005	0.005	
Al	0.005	0.004	0.003		Al	0.015	0.009	0.009	0.009	Al	0.006	0.006		0.006	0.010	
V		0.001	0.001		V	0.000	0.000	0.000	0.000	V						
La	0.008	0.007	0.006		La	0.011	0.008	0.009	0.006	La						
Ce	0.017	0.013	0.012		Ce	0.032	0.031	0.029	0.019	Ce		0.004	0.004			0.015
Fe ²⁺	0.026	0.025	0.021		Fe ²⁺	0.035	0.035	0.033	0.024	Fe ²⁺	0.034	0.018	0.020	0.015	0.026	
Mg					Mg					Mg						
Sr	0.004	0.005	0.005		Sr					Sr						
Ca	0.943	0.958	0.959		Ca	0.913	0.913	0.917	0.947	Ca	0.982	0.984	0.983	0.985	0.949	
Na	0.019	0.020	0.015		Na	0.025	0.036	0.029	0.000	Na						
TOTAL	2.005	2.012	2.007		TOTAL	2.005	2.011	2.006	1.994	TOTAL	2.009	2.004	2.003	2.001	1.991	
O	3	3	3		O	3	3	3	3	O	3	3	3	3	3	

APPENDIX Jxvii PEROVSKITE (cont.)

C.18.26 <u>phosconite</u>							N.64.19 <u>carb.</u>		TQ <u>turjite chilled margin</u>			
	PT 32	PT 44	PT 45	PT 46	PT 47	PT 48		PT 76		PT 27	PT 39	PT 43
Nb ₂ O ₅	1.67	0.28	0.50	1.07	1.63	0.69	Nb ₂ O ₅	3.51	Nb ₂ O ₅	1.34	1.35	1.35
TiO ₂	53.34	56.91	55.57	53.46	52.68	55.05	TiO ₂	46.46	TiO ₂	53.31	52.17	52.69
SiO ₂							SiO ₂		SiO ₂	0.16	0.55	0.51
Al ₂ O ₃	0.39	0.17	0.23	0.41	0.37	0.29	Al ₂ O ₃	0.71	Al ₂ O ₃	0.16	0.16	0.15
V ₂ O ₃	0.44	0.45	0.41	0.41	0.39	0.36	V ₂ O ₃	0.04	V ₂ O ₃	0.13		
La ₂ O ₃	1.21	0.71	1.14	1.21	1.56	0.99	La ₂ O ₃	1.07	La ₂ O ₃	0.84	0.83	0.76
Ce ₂ O ₃	2.86	1.25	1.81	3.00	3.33	2.22	Ce ₂ O ₃	3.46	Ce ₂ O ₃	1.89	1.84	1.86
FeO	1.89	0.97	1.22	1.89	2.08	1.44	FeO	3.75	FeO	2.25	2.44	2.49
MgO							MgO	0.05	MgO			
SrO	0.27	0.24	0.23	0.28	0.29	0.27	SrO		SrO	0.53	0.54	0.53
CaO	36.82	39.53	38.62	36.27	36.03	37.93	CaO	34.41	CaO	36.99	36.42	36.89
Na ₂ O	0.41	0.18	0.26	0.47	0.54	0.23	Na ₂ O	0.49	Na ₂ O	0.90	1.08	1.02
TOTAL	99.30	100.69	99.99	98.47	98.90	99.47	TOTAL	93.95	TOTAL	98.50	97.38	98.25
Nb	0.018	0.003	0.005	0.011	0.018	0.007	Nb	0.040	Nb	0.014	0.015	0.014
Ti	0.946	0.978	0.968	0.956	0.944	0.966	Ti	0.887	Ti	0.950	0.940	0.940
Si							Si		Si	0.004	0.013	0.012
Al	0.011	0.005	0.006	0.011	0.010	0.008	Al	0.021	Al	0.004	0.005	0.004
V	0.008	0.008	0.008	0.008	0.007	0.007	V	0.001	V	0.002		
La	0.011	0.006	0.010	0.011	0.014	0.009	La	0.010	La	0.007	0.007	0.007
Ce	0.025	0.010	0.015	0.026	0.029	0.019	Ce	0.032	Ce	0.016	0.016	0.016
Fe ²⁺	0.037	0.019	0.024	0.038	0.041	0.028	Fe ²⁺	0.080	Fe ²⁺	0.045	0.049	0.049
Mg							Mg	0.002	Mg			
Sr	0.004	0.003	0.003	0.004	0.004	0.004	Sr		Sr	0.007	0.007	0.007
Ca	0.931	0.968	0.959	0.924	0.920	0.949	Ca	0.936	Ca	0.939	0.935	0.938
Na	0.019	0.008	0.012	0.022	0.025	0.010	Na	0.024	Na	0.041	0.050	0.047
TOTAL	2.009	2.007	2.010	2.010	2.012	2.007	TOTAL	2.033	TOTAL	2.030	2.036	2.036
O	3	3	3	3	3	3	O	3	O	3	3	3

APPENDIX Jxviii PYROXENE**C.AG.347**pyroxenite

	PT 54	PT 56	PT 57	PT 63	PT 65	PT 68	PT 70	PT 74	PT 76
SiO ₂	46.39	50.78	46.42	46.47	46.23	46.98	50.62	49.35	46.72
TiO ₂	2.92	1.20	2.62	2.81	2.93	2.70	1.45	1.82	2.80
Al ₂ O ₃	6.18	3.44	6.14	6.22	6.63	6.34	3.39	4.85	6.15
Fe ₂ O ₃	5.31	3.25	5.63	4.92	5.22	4.40	3.25	3.33	5.18
FeO	0.76	0.95	0.96	1.79	1.69	1.61	2.00	1.37	0.99
MnO	0.08		0.09	0.09		0.06	0.07		0.05
MgO	13.29	15.32	13.31	13.05	12.76	13.33	14.68	14.52	13.51
SrO	0.10		0.10	0.16		0.10		0.09	0.09
CaO	24.37	25.24	24.29	24.11	24.43	24.21	24.45	25.11	24.32
Na ₂ O	0.49	0.26	0.41	0.42	0.47	0.45	0.49	0.25	0.44
TOTAL	99.89	100.44	99.96	100.04	100.36	100.18	100.40	100.69	100.25
Si	1.726	1.857	1.728	1.731	1.717	1.741	1.859	1.808	1.731
²⁷ Al	0.271	0.143	0.269	0.269	0.283	0.259	0.141	0.192	0.269
⁵⁶ Fe ³⁺	0.003		0.003						
²⁷ Al		0.005		0.004	0.007	0.018	0.006	0.017	
⁵⁶ Fe ³⁺	0.146	0.089	0.016	0.138	0.146	0.123	0.090	0.092	0.144
Ti	0.082	0.033	0.073	0.079	0.082	0.075	0.040	0.050	0.078
Fe ²⁺	0.024	0.029	0.030	0.056	0.053	0.050	0.061	0.042	0.031
Mn	0.003		0.003	0.003		0.002	0.002		0.002
Mg	0.737	0.835	0.739	0.725	0.706	0.736	0.804	0.793	0.746
Sr	0.002		0.002	0.003		0.002		0.002	0.002
Ca	0.972	0.989	0.969	0.962	0.972	0.961	0.962	0.986	0.966
Na	0.035	0.018	0.030	0.030	0.034	0.032	0.035	0.018	0.032
TOTAL	4.000	4.000	4.000	4.000	4.000	4.000	4.000	4.000	4.000
O	6	6	6	6	6	6	6	6	6

C.AG.48Aolivine melteigite

	PT 28	PT 36	PT 43	PT 44	PT 45	PT 46	PT 47	PT 48	PT 49	PT 53	PT 54	PT 55	PT 57	PT 58	PT 72	PT 74
SiO ₂	49.78	47.94	50.54	48.10	49.51	51.58	50.38	51.19	49.10	50.61	50.04	51.26	49.23	52.53	50.14	47.70
TiO ₂	1.63	2.03	1.01	2.03	1.82	0.90	1.21	0.86	1.64	1.07	0.83	0.28	1.56	0.85	1.33	2.02
Al ₂ O ₃	3.46	5.01	2.50	4.37	3.26	2.34	2.78	2.36	3.25	1.95	1.38	0.76	2.94	1.14	3.03	4.96
Fe ₂ O ₃	4.28	3.61	3.51	4.55	3.06	1.16	3.04	1.85	3.75	2.93	6.61	10.11	3.71	1.93	3.07	4.33
FeO	2.96	3.56	1.72	3.43	4.92	3.15	6.15	2.54	4.29	6.35	12.95	10.91	4.63	2.34	4.51	3.02
MnO	0.10	0.11	0.00	0.12	0.19	0.08	0.16	0.05	0.16	0.22	0.44	0.59	0.19	0.08	0.16	0.11
MgO	13.16	12.78	15.06	12.65	12.51	15.45	11.63	15.54	12.51	12.03	5.35	4.88	12.39	15.62	13.18	12.94
SrO	0.13	0.17	0.16	0.11	0.20	0.12	0.20	0.12	0.14	0.22	0.50	0.16	0.21	0.24	0.13	0.13
CaO	23.25	23.15	23.00	23.03	23.09	23.35	22.49	23.33	23.20	22.89	17.64	15.64	23.10	24.17	22.74	23.03
Na ₂ O	0.99	0.63	0.69	0.79	0.80	0.37	1.15	0.37	0.78	0.86	3.17	4.57	0.78	0.47	0.81	0.66
TOTAL	99.74	98.99	98.19	99.18	99.36	98.50	99.19	98.22	98.82	99.12	98.91	99.16	98.74	99.37	99.11	98.90
Si	1.854	1.803	1.892	1.810	1.862	1.921	1.903	1.911	1.856	1.916	1.967	1.999	1.866	1.942	1.882	1.796
²⁷ Al	0.146	0.197	0.108	0.190	0.138	0.079	0.097	0.089	0.144	0.084	0.033	0.001	0.131	0.050	0.118	0.204
⁵⁶ Fe ³⁺													0.003	0.008		
²⁷ Al	0.006	0.025	0.002	0.004	0.007	0.024	0.027	0.015	0.001	0.003	0.031	0.034			0.016	0.016
⁵⁶ Fe ³⁺	0.120	0.102	0.099	0.129	0.087	0.032	0.086	0.052	0.107	0.083	0.195	0.297	0.103	0.046	0.087	0.123
Ti	0.046	0.057	0.028	0.057	0.051	0.025	0.034	0.024	0.047	0.030	0.025	0.008	0.044	0.024	0.038	0.057
Fe ²⁺	0.092	0.112	0.054	0.108	0.155	0.098	0.194	0.079	0.136	0.201	0.426	0.356	0.147	0.072	0.142	0.095
Mn	0.003	0.004	0.000	0.004	0.006	0.003	0.005	0.002	0.005	0.007	0.015	0.019	0.006	0.003	0.005	0.004
Mg	0.731	0.717	0.840	0.710	0.701	0.858	0.655	0.865	0.705	0.679	0.313	0.284	0.700	0.861	0.737	0.726
Sr	0.003	0.004	0.003	0.002	0.004	0.003	0.004	0.003	0.003	0.005	0.011	0.004	0.005	0.005	0.003	0.003
Ca	0.928	0.933	0.923	0.929	0.930	0.932	0.910	0.933	0.940	0.928	0.743	0.653	0.938	0.957	0.914	0.929
Na	0.071	0.046	0.050	0.058	0.058	0.027	0.084	0.027	0.057	0.063	0.242	0.345	0.057	0.034	0.059	0.048
TOTAL	4.000	4.000	4.000	4.000	4.000	4.000	4.000	4.000	4.000	4.000	4.000	4.000	4.000	4.000	4.000	4.000
O	6	6	6	6	6	6	6	6	6	6	6	6	6	6	6	6

C.90.29pyroxenite

	PT 39	PT 42	PT 44
SiO ₂	46.45	46.04	45.28
TiO ₂	2.40	2.42	2.55
Al ₂ O ₃	6.21	6.05	6.61
Fe ₂ O ₃	3.82	4.44	5.63
FeO	3.27	2.76	2.14
MnO	0.11	0.08	0.10
MgO	12.00	12.02	11.83
SrO			0.08
CaO	23.71	23.77	23.89
Na ₂ O	0.55	0.54	0.54
TOTAL	98.14	97.68	98.08
Si	1.759	1.751	1.719
²⁷ Al	0.241	0.249	0.281
⁵⁶ Fe ³⁺			
²⁷ Al	0.036	0.023	0.015
⁵⁶ Fe ³⁺	0.109	0.127	0.161
Ti	0.068	0.069	0.073
Fe ²⁺	0.104	0.088	0.068
Mn	0.004	0.003	0.003
Mg	0.677	0.682	0.669
Sr			0.005
Ca	0.962	0.969	0.972
Na	0.040	0.040	0.040
TOTAL	4.000	4.000	4.000
O	6	6	6

APPENDIX Jxviii PYROXENE (cont.)

WKM	<u>melteigite</u>				N.64.18	<u>calcite melilitolite</u>						E.133.30	<u>turjaite</u>	
	PT 58	PT 70	PT 77	PT 80		PT 45	PT 1	PT 2	PT 3	PT 4	PT 5		PT 41	PT 48
SiO ₂	51.29	50.39	50.33	50.19	SiO ₂	52.17	55.56	55.44	54.99	55.34	55.18	SiO ₂	53.80	50.23
TiO ₂	0.35	0.62	0.62	0.59	TiO ₂	0.35						TiO ₂		1.16
Al ₂ O ₃	0.66	1.19	1.15	1.14	Al ₂ O ₃	0.39	0.26	0.31		0.37	0.47	Al ₂ O ₃	0.88	2.57
Fe ₂ O ₃	4.79	5.22	4.83	6.22	Fe ₂ O ₃	28.34	0.59	0.96	0.93		0.41	Fe ₂ O ₃		2.62
FeO	3.30	2.39	2.64	1.86	FeO	3.37	0.26			0.66		FeO	5.31	4.79
MnO	0.41	0.35	0.33	0.36	MnO	0.07	0.13	0.12	0.22			MnO	0.23	0.20
MgO	13.28	13.47	13.47	13.46	MgO	0.95	18.12	18.27	18.06	18.08	18.51	MgO	14.68	12.78
SrO	0.09				SrO	0.14						SrO	0.09	0.11
CaO	23.65	23.90	23.78	23.68	CaO	1.26	26.34	26.32	26.03	25.90	26.22	CaO	24.74	23.39
Na ₂ O	0.84	0.74	0.71	0.86	Na ₂ O	11.94						Na ₂ O	0.17	0.71
TOTAL	98.66	98.27	97.85	98.36	TOTAL	99.14	101.26	101.42	100.23	100.35	100.79	TOTAL	99.90	98.55
Si	1.938	1.909	1.914	1.901	Si	2.020	1.987	1.979	1.987	1.993	1.978	Si	1.988	1.899
^{iv} Al	0.029	0.053	0.052	0.051	^{iv} Al		0.011	0.013		0.007	0.020	^{iv} Al	0.012	0.101
^{iv} Fe ³⁺	0.033	0.038	0.034	0.048	^{iv} Fe ³⁺		0.002	0.008	0.013		0.002	^{iv} Fe ³⁺		
^{vi} Al					^{vi} Al	0.018				0.009		^{vi} Al	0.026	0.013
^{vi} Fe ³⁺	0.003	0.111	0.101	0.129	^{vi} Fe ³⁺	0.826	0.014	0.018	0.012		0.009	^{vi} Fe ³⁺		0.074
Ti	0.010	0.018	0.018	0.017	Ti	0.010						Ti		0.033
Fe ²⁺	0.104	0.076	0.084	0.059	Fe ²⁺	0.109	0.008			0.020		Fe ²⁺	0.164	0.151
Mn	0.013	0.011	0.011	0.012	Mn	0.002	0.004	0.004	0.007			Mn	0.007	0.006
Mg	0.748	0.761	0.763	0.760	Mg	0.055	0.966	0.972	0.973	0.971	0.989	Mg	0.809	0.720
Sr	0.002				Sr	0.003						Sr	0.002	0.002
Ca	0.957	0.970	0.969	0.961	Ca	0.052	1.009	1.007	1.008	1.000	1.007	Ca	0.979	0.947
Na	0.062	0.054	0.052	0.063	Na	0.896						Na	0.012	0.052
TOTAL	4.000	4.000	4.000	4.000	TOTAL	4.000	4.000	4.001	4.000	3.999	4.006	TOTAL	3.999	4.000
O	6	6	6	6	O	6	6	6	6	6	6	O	6	6

S.162.300 melilitolite - ijolite contact

	PT 13	PT 14	PT 16	PT 27	PT 33	PT 35	PT 36	PT 37
SiO ₂	54.25	51.30	53.52	53.30	52.94	54.10	52.94	53.11
TiO ₂	0.17	1.02	0.20	0.13	0.11	0.09	0.11	0.14
Al ₂ O ₃	0.32	0.84	0.52	0.42	0.33	0.31	0.37	0.38
Fe ₂ O ₃	1.24	4.96	2.47	2.45	2.50	1.13	2.07	2.49
FeO	3.52	7.90	3.44	4.16	6.37	4.71	6.62	6.25
MnO	0.21	0.47	0.26	0.34	0.53	0.42	0.44	0.44
MgO	15.59	9.90	14.77	14.29	12.65	14.66	12.62	12.81
SrO		0.10	0.11		0.14	0.11	0.16	0.15
CaO	24.45	20.88	24.53	24.27	23.49	23.99	23.18	23.21
Na ₂ O	0.47	2.03	0.57	0.60	0.81	0.58	0.87	0.92
TOTAL	100.21	99.41	100.39	99.96	99.87	100.10	99.38	99.90
Si	1.988	1.956	1.969	1.975	1.984	1.996	1.991	1.986
^{iv} Al	0.012	0.038	0.023	0.018	0.015	0.004	0.009	0.014
^{iv} Fe ³⁺		0.006	0.008	0.007	0.001			
^{vi} Al	0.002					0.009	0.007	0.003
^{vi} Fe ³⁺	0.034	0.136	0.060	0.061	0.070	0.031	0.059	0.070
Ti	0.005	0.029	0.006	0.004	0.003	0.002	0.003	0.004
Fe ²⁺	0.108	0.252	0.106	0.129	0.200	0.145	0.208	0.195
Mn	0.007	0.015	0.008	0.011	0.017	0.013	0.014	0.014
Mg	0.852	0.563	0.810	0.789	0.707	0.806	0.708	0.714
Sr		0.002	0.002		0.003	0.002	0.003	0.003
Ca	0.960	0.853	0.967	0.963	0.943	0.948	0.934	0.930
Na	0.033	0.150	0.041	0.043	0.059	0.041	0.063	0.067
TOTAL	4.000	4.000	4.000	4.000	4.000	4.000	4.000	4.000
O	6	6	6	6	6	6	6	6

TUR I turjaite

	PT 6	PT 1	PT 2	PT 4	PT 5	PT 6
SiO ₂	52.73	49.69	49.34	50.30	50.26	50.35
TiO ₂	0.50	1.05	0.75	1.04	0.62	0.99
Al ₂ O ₃	2.68	3.95	4.29	3.56	3.92	3.95
Fe ₂ O ₃	2.39	4.64	6.35	1.62	2.10	1.97
FeO	4.31	3.55	1.73	4.96	5.11	5.06
MnO	0.17	0.17	0.19			
MgO	14.36	12.57	12.92	13.84	13.45	13.42
SrO	0.15					
CaO	24.16	24.51	24.13	24.55	24.64	25.07
Na ₂ O	0.51	0.61	0.82			
TOTAL	101.96	100.75	100.53	99.87	100.10	100.81
Si	1.914	1.842	1.826	1.870	1.868	1.859
^{iv} Al	0.086	0.158	0.174	0.130	0.132	0.141
^{iv} Fe ³⁺						
^{vi} Al	0.029	0.014	0.014	0.029	0.039	0.031
^{vi} Fe ³⁺	0.065	0.129	0.177	0.045	0.059	0.055
Ti	0.014	0.029	0.021	0.029	0.017	0.027
Fe ²⁺	0.131	0.110	0.054	0.154	0.159	0.156
Mn	0.005	0.005	0.006			
Mg	0.777	0.695	0.713	0.767	0.745	0.739
Sr	0.003					
Ca	0.940	0.973	0.957	0.978	0.981	0.991
Na	0.036	0.044	0.059			
TOTAL	4.000	3.999	4.001	4.002	4.000	3.999
O	6	6	6	6	6	6

APPENDIX Jxviii PYROXENE (cont.)

C.K.7	phoscorite							C.18.26	phoscorite						
	PT 28	PT 30	PT 35	PT 42	PT 43	PT 46	PT 50		PT 26	PT 27	PT 29	PT 35	PT 38	PT 40	PT 49
SiO ₂	55.37	55.06	55.29	54.67	55.35	54.49	54.57	SiO ₂	49.03	49.53	52.97	52.85	53.78	53.28	51.70
TiO ₂	0.23	0.21	0.18	0.49	0.26	0.42	0.24	TiO ₂	1.00	0.53	0.42	0.38	0.33	0.56	0.64
Al ₂ O ₃	0.19	0.18	0.07	0.18	0.10	0.21	0.33	Al ₂ O ₃	4.30	3.99	1.12	1.02	0.69	1.25	2.52
Fe ₂ O ₃	0.22		0.34	0.74	0.30	0.21	0.42	Fe ₂ O ₃	5.60	4.85	2.01	4.31	3.05	1.95	2.01
FeO	2.22	2.51	2.17	3.86	2.13	3.42	2.43	FeO	2.29	3.15	3.51	1.67	4.20	3.49	2.82
MnO	0.10	0.13	0.13	0.17	0.11	0.15	0.14	MnO	0.24	0.26	0.18	0.21	0.19	0.15	0.17
MgO	17.18	16.91	17.20	15.81	17.24	16.03	16.72	MgO	12.75	12.81	14.52	14.78	13.52	14.72	14.84
SrO	0.13	0.11		0.09	0.12		0.11	SrO	0.09				0.09		
CaO	24.93	24.71	24.84	23.39	25.10	23.60	24.68	CaO	24.38	24.05	23.90	24.41	22.38	23.94	24.38
Na ₂ O	0.31	0.26	0.32	0.77	0.26	0.68	0.30	Na ₂ O	0.64	0.57	0.76	0.87	1.59	0.79	0.37
TOTAL	100.88	100.08	100.54	100.16	100.97	99.21	99.94	TOTAL	100.32	99.74	99.39	100.50	99.83	100.13	99.45
Si	1.998	2.003	2.000	2.000	1.996	2.005	1.991	Si	1.823	1.850	1.963	1.939	1.990	1.959	1.913
^{iv} Al	0.002				0.004		0.009	^{iv} Al	0.177	0.150	0.037	0.044	0.010	0.041	0.087
^{iv} Fe ³⁺								^{iv} Fe ³⁺				0.017			
^{vi} Al	0.006	0.008	0.003	0.008		0.009	0.005	^{vi} Al	0.011	0.026	0.012		0.02	0.013	0.023
^{iv} Fe ³⁺	0.006		0.009	0.020	0.008	0.006	0.011	^{iv} Fe ³⁺	0.157	0.136	0.056	0.102	0.085	0.054	0.056
Ti	0.006	0.006	0.005	0.013	0.007	0.012	0.007	Ti	0.028	0.015	0.012	0.010	0.009	0.015	0.018
Fe ²⁺	0.067	0.076	0.066	0.118	0.064	0.105	0.074	Fe ²⁺	0.071	0.098	0.109	0.051	0.130	0.107	0.087
Mn	0.003	0.004	0.004	0.005	0.003	0.005	0.004	Mn	0.008	0.008	0.006	0.007	0.006	0.005	0.005
Mg	0.924	0.917	0.928	0.862	0.927	0.879	0.910	Mg	0.707	0.713	0.802	0.808	0.746	0.807	0.818
Sr	0.003	0.002		0.002	0.003		0.002	Sr	0.002				0.002		
Ca	0.964	0.963	0.963	0.917	0.970	0.930	0.965	Ca	0.971	0.962	0.949	0.959	0.887	0.943	0.966
Na	0.022	0.018	0.022	0.055	0.018	0.049	0.021	Na	0.046	0.041	0.055	0.062	0.114	0.056	0.027
TOTAL	4.000	3.997	4.000	4.000	4.000	4.000	4.000	TOTAL	4.000	4.000	4.000	4.000	4.000	4.000	4.000
O	6	6	6	6	6	6	6	O	6	6	6	6	6	6	6

C.TL.344	carbonatite			C.XC	carbonatite			N.63.4	carbonatite					
	PT 66	PT 72	PT 81		PT 1	PT 2	PT 21		PT 62	PT 63	PT 64	PT 66	PT 67	PT 70
SiO ₂	54.55	53.72	54.36	SiO ₂	52.27	52.00	51.62	SiO ₂	51.80	53.26	51.12	52.33	49.00	50.12
TiO ₂	0.31	0.41	0.40	TiO ₂	0.16	0.22	0.36	TiO ₂	0.74	0.35	0.81	0.58	1.32	1.19
Al ₂ O ₃	0.37	0.45	0.39	Al ₂ O ₃	0.09	0.29	0.28	Al ₂ O ₃	2.39	0.91	2.52	1.97	4.08	3.28
Fe ₂ O ₃	0.52	0.93	1.11	Fe ₂ O ₃	4.09	5.20	5.04	Fe ₂ O ₃	2.35	1.72	3.05	1.58	5.18	3.88
FeO	2.51	2.30	2.09	FeO	0.87		0.13	FeO	2.00	1.65	2.01	2.41	1.41	2.13
MnO	0.12	0.08	0.12	MnO	0.24	0.16	0.17	MnO	0.26	0.24	0.31	0.27	0.28	0.29
MgO	16.44	16.41	16.41	MgO	15.22	15.12	15.02	MgO	15.32	16.04	14.73	15.44	13.55	14.12
SrO	0.00	0.11	0.14	SrO	0.08			SrO	0.09	0.13	0.15	0.11		0.13
CaO	24.86	24.23	24.97	CaO	23.91	23.23	23.93	CaO	24.83	25.22	24.77	24.85	24.56	24.74
Na ₂ O	0.37	0.40	0.39	Na ₂ O	0.80	1.20	0.93	Na ₂ O	0.25	0.24	0.31	0.21	0.40	0.35
TOTAL	100.05	99.03	100.38	TOTAL	97.73	97.42	97.49	TOTAL	100.02	99.76	99.78	99.75	99.98	100.23
Si	1.990	1.980	1.979	Si	1.965	1.957	1.946	Si	1.904	1.955	1.891	1.927	1.821	1.854
^{iv} Al	0.010	0.020	0.017	^{iv} Al	0.004	0.013	0.012	^{iv} Al	0.096	0.039	0.109	0.073	0.179	0.143
^{iv} Fe ³⁺			0.004	^{iv} Fe ³⁺	0.031	0.030	0.042	^{iv} Fe ³⁺		0.006				0.003
^{vi} Al	0.006			^{vi} Al				^{vi} Al	0.008		0.001	0.012		
^{iv} Fe ³⁺	0.014	0.026	0.026	^{iv} Fe ³⁺	0.085	0.117	0.101	^{iv} Fe ³⁺	0.065	0.042	0.085	0.044	0.145	0.105
Ti	0.009	0.011	0.011	Ti	0.005	0.006	0.010	Ti	0.020	0.010	0.023	0.016	0.037	0.033
Fe ²⁺	0.077	0.071	0.064	Fe ²⁺	0.027		0.004	Fe ²⁺	0.061	0.051	0.062	0.074	0.044	0.066
Mn	0.004	0.002	0.004	Mn	0.008	0.005	0.005	Mn	0.008	0.007	0.010	0.008	0.009	0.009
Mg	0.894	0.902	0.891	Mg	0.853	0.848	0.844	Mg	0.840	0.878	0.812	0.848	0.751	0.779
Sr	0.000	0.002	0.003	Sr	0.002			Sr	0.002	0.003	0.003	0.002		0.003
Ca	0.971	0.957	0.974	Ca	0.963	0.937	0.967	Ca	0.978	0.992	0.982	0.980	0.978	0.980
Na	0.026	0.029	0.028	Na	0.058	0.088	0.068	Na	0.018	0.017	0.022	0.015	0.029	0.025
TOTAL	4.000	4.000	4.000	TOTAL	4.000	4.001	4.000	TOTAL	4.000	4.000	4.000	4.000	4.000	4.000
O	6	6	6	O	6	6	6	O	6	6	6	6	6	6

APPENDIX Jxviii PYROXENE (cont.)

N.64.19	carbonatite			N.187.100	carbonatite				TD 2	aeg-aug-ap vein			
	PT 56	PT 58	PT 60		PT 3	PT 4	PT 5	PT 20		PT 34	PT 37	PT 40	PT 45
SiO ₂	54.13	53.06	53.50	SiO ₂	49.08	49.36	48.35	51.80	SiO ₂	52.21	52.24	52.58	52.07
TiO ₂	0.14	0.25	0.22	TiO ₂	0.50	0.58	0.69		TiO ₂	0.18	0.28	0.18	0.30
Al ₂ O ₃	0.63	1.33	1.24	Al ₂ O ₃	1.60	1.68	1.81	0.38	Al ₂ O ₃	0.56	0.59	0.46	0.59
Fe ₂ O ₃	1.39	1.67	1.66	Fe ₂ O ₃	5.81	5.54	6.29	3.25	Fe ₂ O ₃	15.77	10.29	9.71	8.43
FeO	1.40	2.09	2.42	FeO	1.67	3.52	3.00		FeO	3.08	6.38	5.47	7.54
MnO	0.24	0.26	0.30	MnO	0.49	0.52	0.53	0.50	MnO	0.82	0.59	0.57	0.59
MgO	16.73	15.71	15.69	MgO	13.11	11.60	11.47	15.98	MgO	6.74	7.88	8.95	8.10
SrO	0.16	0.19	0.14	SrO					SrO		0.10		0.10
CaO	25.36	24.96	25.08	CaO	23.84	22.33	22.22	25.05	CaO	13.59	17.47	17.73	17.78
Na ₂ O	0.17	0.26	0.27	Na ₂ O	0.66	1.34	1.29	0.27	Na ₂ O	6.31	4.15	3.95	3.69
TOTAL	100.35	99.78	100.52	TOTAL	96.76	96.46	95.65	97.23	TOTAL	99.26	99.97	99.60	99.18
Si	1.970	1.950	1.954	Si	1.890	1.914	1.894	1.951	Si	1.989	1.985	1.991	1.994
²⁷ Al	0.027	0.050	0.046	²⁷ Al	0.073	0.077	0.084	0.017	²⁷ Al	0.011	0.015	0.009	0.006
⁵⁴ Fe ³⁺	0.003			⁵⁴ Fe ³⁺	0.037	0.009	0.022	0.032	⁵⁴ Fe ³⁺				
²⁷ Al		0.008	0.007	²⁷ Al					²⁷ Al	0.014	0.011	0.012	0.021
⁵⁴ Fe ³⁺	0.035	0.046	0.046	⁵⁴ Fe ³⁺	0.132	0.153	0.163	0.060	⁵⁴ Fe ³⁺	0.452	0.294	0.277	0.243
Ti	0.004	0.007	0.006	Ti	0.015	0.017	0.020		Ti	0.005	0.008	0.005	0.009
Fe ²⁺	0.043	0.064	0.074	Fe ²⁺	0.054	0.114	0.099		Fe ²⁺	0.098	0.203	0.173	0.241
Mn	0.007	0.008	0.009	Mn	0.016	0.017	0.018	0.016	Mn	0.026	0.019	0.018	0.019
Mg	0.907	0.861	0.854	Mg	0.753	0.671	0.670	0.898	Mg	0.383	0.446	0.505	0.462
Sr	0.003	0.004	0.003	Sr					Sr		0.002		0.002
Ca	0.989	0.983	0.981	Ca	0.984	0.928	0.933	1.011	Ca	0.555	0.711	0.719	0.729
Na	0.012	0.019	0.019	Na	0.049	0.101	0.098	0.020	Na	0.466	0.306	0.290	0.274
TOTAL	4.000	4.000	4.000	TOTAL	4.000	4.000	4.000	4.004	TOTAL	4.000	4.000	4.000	4.000
O	6	6	6	O	6	6	6	6	O	6	6	6	6

C.190.x	ijolite		(contact)	fenite					NC.153.82	fenite						
	PT 20	PT 23		PT 6	PT 10	PT 12	PT 25	PT 26		PT 28	PT 63	PT 65	PT 68	PT 69	PT 70	PT 72
SiO ₂	50.97	50.31		49.64	50.15	50.06	49.66	49.93	51.79	SiO ₂	53.02	51.67	52.80	52.19	53.59	52.71
TiO ₂	0.50	0.94		0.86	0.69	0.45	0.72	0.65	0.48	TiO ₂	0.54	0.58	0.47	1.82	1.07	0.90
Al ₂ O ₃	1.35	1.90		0.99	0.74	0.78	0.97	0.99	1.14	Al ₂ O ₃	0.35	0.24	0.35	0.34	0.24	0.24
Fe ₂ O ₃	4.60	4.57		17.73	23.48	20.32	18.24	17.72	26.35	Fe ₂ O ₃	10.51	13.37	10.70	9.79	9.41	11.88
FeO	6.12	6.31		10.94	5.34	8.72	10.56	11.20	3.78	FeO	5.83	6.90	5.06	9.94	5.16	7.11
MnO	0.30	0.27		0.61	0.65	0.55	0.57	0.57	0.38	MnO	0.30	0.39	0.20	1.00	0.21	0.34
MgO	11.48	11.20		0.63	0.62	0.56	0.58	0.62	0.35	MgO	8.46	6.04	8.68	4.95	9.19	6.67
SrO	0.17	0.10		0.25	0.15	0.22	0.25	0.28		SrO	0.13	0.17	0.10		0.18	0.14
CaO	22.28	22.57		9.32	6.82	8.03	9.17	9.56	2.34	CaO	15.97	13.78	16.90	13.05	15.63	13.84
Na ₂ O	1.26	1.18		7.62	10.02	8.53	7.75	7.54	11.77	Na ₂ O	4.77	5.71	4.55	5.94	4.99	5.75
TOTAL	99.03	99.35		98.60	98.66	98.22	98.47	99.07	98.38	TOTAL	99.88	98.85	99.81	99.02	99.67	99.58
Si	1.936	1.909		1.979	1.967	1.993	1.982	1.983	2.018	Si	2.002	1.997	1.993	2.020	2.013	2.011
²⁷ Al	0.060	0.085		0.021	0.033	0.007	0.018	0.017		²⁷ Al		0.003	0.007			
⁵⁴ Fe ³⁺	0.004	0.006								⁵⁴ Fe ³⁺						
²⁷ Al				0.026	0.001	0.030	0.028	0.029	0.052	²⁷ Al	0.016	0.008	0.009	0.016	0.011	0.011
⁵⁴ Fe ³⁺	0.127	0.124		0.532	0.693	0.609	0.548	0.530	0.773	⁵⁴ Fe ³⁺	0.299	0.389	0.304	0.285	0.266	0.341
Ti	0.014	0.027		0.026	0.020	0.013	0.022	0.019	0.014	Ti	0.015	0.017	0.013	0.053	0.030	0.026
Fe ²⁺	0.195	0.200		0.365	0.175	0.290	0.352	0.372	0.123	Fe ²⁺	0.184	0.223	0.160	0.322	0.162	0.227
Mn	0.010	0.009		0.021	0.022	0.019	0.019	0.019	0.013	Mn	0.010	0.013	0.006	0.033	0.007	0.011
Mg	0.650	0.634		0.037	0.036	0.033	0.035	0.037	0.020	Mg	0.476	0.348	0.489	0.286	0.515	0.379
Sr	0.004	0.002		0.006	0.003	0.005	0.006	0.006		Sr	0.003	0.004	0.002		0.004	0.003
Ca	0.907	0.918		0.398	0.287	0.343	0.392	0.407	0.098	Ca	0.646	0.571	0.684	0.541	0.629	0.566
Na	0.093	0.087		0.589	0.762	0.658	0.600	0.581	0.889	Na	0.349	0.428	0.333	0.446	0.363	0.425
TOTAL	4.000	4.000		4.000	4.000	4.000	4.000	4.000	4.000	TOTAL	4.000	4.000	4.000	4.000	4.000	4.000
O	6	6		6	6	6	6	6	6	O	6	6	6	6	6	6

APPENDIX Jxix PYROXENOIDS

pecolite

E.133.30 mel.

	PT 61
SiO ₂	53.35
Al ₂ O ₃	0.91
Fe ₂ O ₃	0.14
MnO	0.07
SrO	0.13
CaO	33.34
Na ₂ O	8.70
TOTAL	96.64

Si	2.968
Al	0.060
Fe ³⁺	0.006
Mn	0.003
Sr	0.004
Ca	1.988
Na	0.939

TOTAL 5.968

O 8.5

S.78.15 mel.

	PT 27
SiO ₂	52.40
Al ₂ O ₃	0.10
Fe ₂ O ₃	
MnO	
SrO	
CaO	33.52
Na ₂ O	8.93
TOTAL	94.95

Si	2.978
Al	0.006
Fe ³⁺	
Mn	
Sr	
Ca	2.041
Na	0.984

TOTAL 6.011

O 8.5

ijolite-fenite samples

	PT 11	PT 65	PT 66	PT 2
SiO ₂	52.06	53.13	53.46	54.56
TiO ₂		0.09		
Fe ₂ O ₃	0.60	0.11		
MnO	2.50	0.19	0.08	0.20
SrO	0.73			
CaO	30.40	32.49	33.47	33.02
Na ₂ O	9.12	9.26	9.06	10.25
TOTAL	95.41	95.27	96.07	98.03

Si	2.975	3.002	2.832	3.001
Ti		0.004		
Fe ³⁺	0.026	0.004		
Mn	0.120	0.009	0.004	0.007
Sr	0.024			
Ca	1.862	1.966	1.900	2.114
Na	1.010	1.014	0.931	1.092

TOTAL 6.018 6.000 5.667 6.046

O 8.5 8.5 8.5 8.5

wollastonite

N.64.18c

	PT 49
SiO ₂	51.12
Al ₂ O ₃	
FeO	0.12
MnO	0.13
SrO	0.19
CaO	47.95
Na ₂ O	
TOTAL	99.51

Si	1.993
Al	
Fe ²⁺	0.004
Mn	0.004
Sr	0.004
Ca	2.002
Na	

TOTAL 4.007

O 6

N.64.18

	PT 97	PT 98
SiO ₂	52.41	51.95
Al ₂ O ₃	0.06	
FeO	0.11	0.09
MgO	0.06	
SrO	0.08	0.08
CaO	47.81	47.93
Na ₂ O	0.05	
TOTAL	100.58	100.05

Si	2.01	2.006
Al	0.003	
Fe ²⁺	0.004	0.003
Mg	0.003	
Sr	0.002	0.002
Ca	1.965	1.983
Na	0.004	

TOTAL 3.99 3.994

O 6 6

S.162.300

	PT 1	PT 4	PT 17
SiO ₂	51.28	51.52	51.75
Fe ₂ O ₃	0.50	0.52	
FeO	0.05		0.28
MnO	0.36	0.31	0.22
MgO	0.14	0.12	0.10
SrO	0.19	0.14	0.14
CaO	47.24	48.10	47.58
TOTAL	99.76	100.71	100.07

Si	1.993	1.986	2.002
Fe ³⁺	0.015	0.015	
Fe ²⁺	0.002		0.009
Mn	0.012	0.010	0.007
Mg	0.008	0.007	0.006
Sr	0.004	0.003	0.003
Ca	1.967	1.986	1.972

TOTAL 4.000 4.007 3.998

O 6 6 6

APPENDIX Jxx TITANITE

C.90.29 pyroxenite				C.AG.48A olivine melteigite				C.AG.122 ol.melt.		E.123.106 ijolite				
	PT 10	PT 14	PT 20		PT 30	PT 65	PT 83		PT 21		PT 7	PT 11	PT 12	PT 29
SiO ₂	30.45	30.31	30.49	SiO ₂	29.94	30.05	30.43	SiO ₂	36.99	SiO ₂	30.40	30.06	30.20	29.64
TiO ₂	36.34	35.32	36.93	TiO ₂	36.12	35.09	35.67	TiO ₂	25.43	TiO ₂	36.46	36.06	37.70	36.27
Al ₂ O ₃	0.74	0.75	0.63	Al ₂ O ₃	1.24	1.62	1.32	Al ₂ O ₃	0.72	Al ₂ O ₃	0.40	0.31	0.33	0.41
FeO	2.09	2.27	1.77	FeO	1.54	1.91	1.75	FeO	4.03	FeO	2.23	1.81	1.44	1.91
MnO				MnO				MnO	0.12	ZrO ₂	0.29	1.77	0.39	0.37
MgO				MgO	0.05	0.09		MgO	4.32	MgO				
CaO	28.26	27.80	27.74	CaO	27.86	27.60	27.81	CaO	26.07	CaO	27.73	27.38	27.47	27.09
SrO	0.14	0.16	0.09	SrO	0.22	0.35	0.51	SrO	0.18	SrO	0.39	0.53	0.46	0.46
BaO		0.18		BaO				BaO		BaO	0.14	0.18		
Na ₂ O	0.15	0.18	0.17	Na ₂ O	0.07	0.12	0.08	Na ₂ O	0.73	Na ₂ O	0.19	0.32	0.23	0.23
F		0.48	0.70	F				F	0.82	F			0.41	0.45
O=F		-0.20	-0.29	O=F				O=F	-0.35	O=F			-0.17	-0.19
TOTAL	98.17	97.25	98.23	TOTAL	97.04	96.77	97.57	TOTAL	99.06	TOTAL	98.23	98.42	98.46	96.64
Si	1.021	1.025	1.015	Si	1.013	1.022	1.025	Si	1.206	Si	1.022	1.014	1.008	1.010
Ti	0.916	0.898	0.925	Ti	0.919	0.898	0.903	Ti	0.084	Ti	0.922	0.915	0.946	0.929
Al	0.029	0.030	0.025	Al	0.049	0.055	0.053	Al	0.028	Al	0.016	0.012	0.013	0.016
Fe ²⁺	0.059	0.064	0.049	Fe ²⁺	0.043	0.054	0.049	Fe ²⁺	0.110	Fe ²⁺	0.063	0.051	0.040	0.054
Mn				Mn				Mn	0.003	Zr	0.005	0.029	0.006	0.006
Mg				Mg	0.003	0.004		Mg	0.252	Mg				
Ca	1.015	1.007	0.990	Ca	1.010	1.006	1.003	Ca	0.910	Ca	0.999	0.990	0.982	0.989
Sr	0.003	0.003	0.002	Sr	0.004	0.007	0.008	Sr	0.003	Sr	0.008	0.010	0.009	0.009
Ba		0.002		Ba				Ba		Ba	0.002	0.002		
Na	0.010	0.012	0.011	Na	0.005	0.008	0.005	Na	0.044	Na	0.012	0.021	0.015	0.015
TOTAL	3.053	3.042	3.016	TOTAL	3.046	3.057	3.048	TOTAL	3.138	TOTAL	3.049	3.046	3.019	3.030
O	5	5	5	O	5	5	5	O	5	O	5	5	5	5

TU 120A turjite					ijolite-fenite			C.190.x ijolite-fenite		
	PT 4	PT 6	PT 14	PT 15		PT 45	PT 57		PT 18	PT 30
SiO ₂	30.30	30.39	30.46	30.51	SiO ₂	29.84	30.48	SiO ₂	30.05	30.05
TiO ₂	37.77	37.14	38.01	38.38	TiO ₂	36.74	38.14	TiO ₂	35.29	36.52
Al ₂ O ₃	0.47	0.51	0.53	0.50	Al ₂ O ₃	0.55	0.30	Al ₂ O ₃	1.29	0.95
Fe ₂ O ₃	2.13	2.14	1.95	1.71	FeO	2.12	1.19	FeO	2.24	1.53
MnO					MnO			MnO		
MgO					MgO			MgO	0.06	
CaO	27.88	27.73	27.85	28.17	CaO	27.69	27.34	CaO	27.68	27.47
SrO					SrO	0.65	0.36	SrO	0.16	0.15
BaO					BaO	0.12	0.11	BaO	0.14	0.14
Na ₂ O					Na ₂ O	0.20	0.48	Na ₂ O	0.09	0.09
F					F	0.63	0.66	F	0.88	
O=F					O=F	-0.27	-0.28	O=F	-0.37	
TOTAL	98.55	97.91	98.80	99.27	TOTAL	98.27	98.78	TOTAL	97.51	96.90
Si	1.007	1.015	1.008	1.005	Si	1.002	1.010	Si	1.010	1.018
Ti	0.944	0.933	0.946	0.951	Ti	0.927	0.950	Ti	0.892	0.930
Al	0.019	0.020	0.021	0.019	Al	0.022	0.012	Al	0.051	0.038
Fe ³⁺	0.053	0.054	0.049	0.042	Fe ³⁺	0.060	0.033	Fe ²⁺	0.063	0.043
Mn					Mn			Mn		
Mg					Mg			Mg	0.003	
Ca	0.992	0.993	0.988	0.995	Ca	0.996	0.971	Ca	0.997	0.997
Sr					Sr	0.013	0.007	Sr	0.003	0.003
Ba					Ba	0.002	0.001	Ba	0.002	0.002
Na					Na	0.013	0.031	Na	0.006	0.006
TOTAL	3.014	3.015	3.011	3.013	TOTAL	3.033	3.015	TOTAL	3.028	3.036
O	5	5	5	5	O	5	5	O	5	5