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**Isotope geochemistry,
petrology, and source evaluation of
the Leucite Hills lamproites, Wyoming**

by
Hassan Mirnejad, B.Sc., M.Sc.

A thesis submitted to
the Faculty of Graduate Studies and Research
in partial fulfillment of
the requirements for the degree of
Doctor of Philosophy

Ottawa-Carleton Geoscience Centre
Department of Earth Sciences
Carleton University
Ottawa, Ontario
Canada
April, 2002

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of the Leucite Hill lamproites, Wyoming

Submitted by
Hassan Mirnejad, B.Sc., M.Sc.

in partial fulfilment of the requirements
for the degree of Doctor of Philosophy

Chair, Department of Earth Sciences

Thesis Supervisor

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ABSTRACT

Major, trace element, Sr, Nd and Pb isotopic data are reported for the Leucite Hills lamproites and their lower crustal xenoliths. The new chemical data indicate that the two main groups of lamproites, phlogopite lamproites and madupitic lamproites are derived from primary magmas. Analysis of separated mineral grains show that the transitional madupitic lamproites can be included with the madupitic lamproites while the olivine-bearing phlogopite lamproites are similar to the phlogopite lamproites. The data from both types of lamproites form two distinct grouping in many isotopic and trace element diagrams.

The texture, mineralogy, and whole rock compositions indicate that the lower crustal xenoliths from the Leucite Hills lamproites are meta-igneous granulites. Estimates of pressure and temperature suggest that they formed at 4-9 kb, and about 800 °C. The chemical and isotopic compositions of the xenoliths indicate that the protoliths of the granulites crystallized from a fractionated, mantle-derived melt. Based on Pb-Pb and Sm-Nd pseudoisochrons, and Nd model ages, the basic parental magma was probably either derived from a CHUR-like reservoir or a depleted mantle source metasomatised by subducted materials of continental crust origin.

The model favoured for the origin of the source of the Leucite Hills lamproites involves Archean depletion followed by a subsequent late-Archean enrichment event. Geochemical evidence such as Nb and Ta depletion and the low U/Pb ratio of the lamproites, as well as the correlation of the data from the lower crustal xenoliths and the lamproites on the Sr-Nd and Pb-Pb isotopic diagrams favor Archean metasomatism of the mantle source by subduction of continental crust materials. Some geochemical features of the Leucite Hills lamproites (e.g., high LILE and K₂O contents) are attributed to a recent metasomatic activity. This recent enrichment may be related to a plume activity which may also have induced melting in the source.

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CHAPTER 1

INTRODUCTION

1.1 Overview of ultrapotassic rocks

Potassium-rich igneous rocks cover a wide range of rock types that vary both in mineralogy and chemical composition. They can be classified into eight main groups: the shoshonite association, Roman-type ultrapotassic series, kamafugites, kimberlites, lamproites, leucitite - leucite basalts, potassic lamprophyres (e.g., minette, alnöite, alvikite, aillikite), and potassium-rich intrusive rocks (e.g., shonkinite, missourite, malignite and yakutite) (Mitchell and Bergman, 1991; Peccerillo, 1992). Among these rocks lamproites, although volumetrically the least significant, have attracted much attention among petrologists because of their unusual geochemistry and because they can contain diamonds.

1.2 Definition of lamproites

Mineralogically, lamproites can contain the following primary phases: titanian phlogopite, titanian richterite, forsteritic olivine, Al_2O_3 - and Na_2O -poor diopside, non-stoichiometric Fe-rich leucite, and Fe-rich sanidine. Some of the minor and accessory phases which may also be present include apatite, chromite, perovskite, magnetite, priderite, wadeite, and shcherbakovite (Mitchell and Bergman, 1991).

Geochemically, lamproites are defined as ultrapotassic Mg-rich igneous rocks characterized by high $\text{K}_2\text{O}/\text{Na}_2\text{O}$ ratio (>5) and extreme trace element concentrations (high Cr and Ni as well as Rb, Sr, Zr, Ba and the LREE). The high concentration of the incompatible trace elements (e.g., Ba > 2000 ppm, Sr > 1000 ppm, Zr > 500 ppm, La > 200 ppm) is one of the more important criteria for lamproite recognition (Scott Smith and Skinner, 1984a; Jaques et al., 1984; Foley et al., 1987; Mitchell and Bergman, 1991). The generally high MgO, Ni and Cr abundance of lamproites are attributed to their derivation from relatively depleted mantle sources (Vollmer et al., 1984). The very high incompatible element contents including LREE, suggest that their sources have been affected by metasomatic activity (Fraser et al., 1985). Lamproites are also characterized

by radiogenic Sr and unradiogenic Nd isotopic compositions and thus plot on the enriched part of the Nd-Sr isotopic diagram (Fig. 1.1). Excluding the Spanish samples, most lamproites have unradiogenic Pb isotopic compositions and normally plot to the left of geochron on the $^{207}\text{Pb}/^{204}\text{Pb}$ versus $^{206}\text{Pb}/^{204}\text{Pb}$ diagram (Fig. 1.2).

1.3 Why study the Leucite Hills lamproites?

The Leucite Hills lamproites are considered to be the type locality for lamproites (Mitchell and Bergman, 1991). They are composed of 22 volcanic occurrences that range in age from 3.1 to 0.8 Ma (Mcdowell, 1971; Carmichael et al., 1997). Because the Leucite Hills lamproites are young, they do not need to be back-corrected for the isotopic growth of radiogenic daughter products. As a result, radiogenic isotope compositions of the Leucite Hills lamproites can provide important information about their mantle source, assuming they have crystallized from primary magmas that have behaved as closed chemical systems.

Based on mineralogy and chemical composition, the Leucite Hills lamproites are divided into madupitic lamproites, transitional madupitic lamproites, phlogopite lamproites, and one sub-group; the olivine-bearing phlogopite lamproites (Mitchell and Bergman, 1991). Madupitic lamproites are more mafic and contain higher concentrations of incompatible trace elements than the phlogopite lamproites. In addition, the madupitic lamproites contain lower time-integrated Rb/Sr, Nd/Sm and U/Pb ratios relative to the phlogopite lamproites (Mitchell and Bergman, 1991). A number of melting experiments on lamproites ranging from SiO₂-poor olivine lamproites and madupitic lamproites to SiO₂-rich phlogopite lamproites have been carried out to determine the P-T conditions, the mineralogy of source, and the evolutionary trend of the lamproite magmas (e.g., Barton and Hamilton, 1979; Edgar and Mitchell, 1997).

The Leucite Hills lamproites have been extensively studied for their major, trace elements and isotopic compositions (e.g., Carmichael, 1967; Vollmer et al. 1984; Fraser, 1987). However, there are no studies that evaluate their alteration and the role that crustal contamination has played in their petrogenesis. In addition, the very important issue concerning the geochemical character of the primary melt has not yet been resolved.

Figure 1.1 Nd and Sr isotopic compositions of lamproites. LH madp lamp: Leucite Hills madupitic lamproites, LH phg lamproites: Leucite Hill phlogopite lamproites, WA ol lamp: Western Australia olivine lamproites, WA phg lamp: Western Australia phlogopite lamproites. After Mitchell and Bergman (1991).

Figure 1.2 Pb isotopic compositions of lamproites. Abbreviations as in Figure 1.1. After Mitchell and Bergman (1991).

None of the research on the Leucite Hills lamproites has helped to successfully and convincingly unravel the nature and the evolutionary history of the mantle source.

A large number of lower crustal xenoliths from the Wyoming Archean Craton have been entrained by the Leucite Hills lamproites. Detailed study of these xenoliths can provide the most direct access to the present-day composition of the Wyoming Archean lower crust and help in understanding the evolution of the underlying mantle.

1.4 Objectives

In spite of the relatively large number of chemical analyses already published, few mineral separates, their corresponding whole rocks and the crustal xenoliths entrained by the lamproites have been analyzed for their trace element and isotopic compositions. Chemical analyses were conducted on selected lamproites including both whole rock and mineral separates. In addition, sedimentary and lower crustal xenoliths were analyzed for their major and trace element contents and isotopic ratios. Whole rocks were analyzed by ICP and ICP-MS for both major and trace elements. Isotopic analyses of Sr, Nd and Pb were carried out using thermal ionization mass spectrometry for both whole rocks and mineral separates. The major and trace element contents of the mineral separates were measured using electron microprobe and laser ablation mass spectrometry.

The aims of the thesis are as follow:

1. Investigation of the lower crustal xenoliths:

Lower crustal xenoliths were collected from Hatcher Mesa during the summer of 1996. The petrography and chemical composition of these xenoliths were used to interpret the nature of the lower crust that lies below the Leucite Hills. Isotopic data including Sm, Nd and Pb are used to determine the age of the lower crustal rocks and to obtain some information about the timing of the major melt removal episode from the Wyoming sub-continental mantle.

2. Evaluation of the role of crustal contamination in the evolution of the Leucite Hills lamproites

The madupitic lamproites from the Leucite Hills contain abundant sedimentary xenoliths. This has led Ogden (1978) to propose that the madupitic lamproites attained their

peculiar mineral assemblage and chemical composition by reaction between the phlogopite lamproite melts and sedimentary xenoliths. In order to see whether these two types of lamproites can be linked by the assimilation of crustal rocks, mixing calculations were used to evaluate the relationships between the lamproites and the crustal rocks. In addition, leaching tests and analysis of individual mineral grains helped ascertain the state of isotopic equilibrium among the Leucite Hills lamproites.

3. The geochemical features of the primary lamproite magmas

There has been an ongoing controversy over the choice of the primary lamproitic melt and the petrogenetic relationships among the different lamproite types. Among different varieties of lamproite types, the madupitic lamproites and olivine lamproites are the most controversial ones. While Jaques et al. (1986) and Foley (1993) consider the Mg-rich magmas such as olivine lamproites as a primary magma, Mitchell and Bergman (1991) and Mitchell (1995) favor the more silica-rich phlogopite lamproite as the parental lamproitic melt. On the other hand, Ogden (1979) and Kuehner (1980) suggest that each type of lamproite in the Leucite Hills originates from a distinct primary melt. In other words, different degrees of partial melting from a common phlogopite-bearing, garnet-free peridotitic source could produce the compositional variations observed in the Leucite Hills suite. Mitchell and Bergman (1991) do not consider the madupitic lamproites as being derived from primary melts and consider them as evolved hybrids with an unsolved petrogenesis. These authors favor the more silica-rich phlogopite lamproites as the primary lamproite magma in the Leucite Hills. Careful examination of the geochemistry of different lamproite types from Leucite Hills has been undertaken to investigate their state of primary characteristics.

4. Characteristics of the mantle source of the Leucite Hills lamproites

Lamproites occur in continental intraplate settings (e.g., Wyoming and Western Australia) and in rare instances in post-collisional environments (e.g., southeastern Spain and central Italy). The cratonic lithospheric mantle tends to be strongly depleted and heterogeneous. The nature of the mantle source of the Leucite Hills lamproites is evaluated in an attempt to understand the extent of chemical (e.g., depletion, metasomatism) and

physical (e.g., subduction, convection) processes that have contributed to mantle heterogeneity. This can be done by a detailed evaluation of the geochemical characteristics of the Leucite Hills lamproites and other contemporaneous volcanic rocks erupted on the Wyoming Archean Craton.

CHAPTER 2

GEOLOGY

The Leucite Hills lamproites are located at 41°47'N, 109°00'W, northeast of Rock Springs in Wyoming, USA (Fig. 2.1). They cut through Upper Cretaceous to Eocene clastic sedimentary rocks and overly a thick cratonic basement known as the Wyoming Archean Province.

A comprehensive account of the geology of Wyoming Province is lacking and very few areas have been studied in detail. Condie (in Peterman, 1979) estimates that about 60% of the Precambrian rocks of Wyoming are gneisses, 30% granites and 10% metavolcanic and metasedimentary rocks. Ages of 2.7 Ga or greater have been obtained from gneisses, including metamorphosed spuracrustals in the Beartooth, Bighorn and Granite Mountains (Muller and Wooden, 1988). Archean metamorphic activity in the Wyoming Province is characterized by amphibolite and granulite grade metamorphism (Ernst, 1988).

The Wyoming Archean Province is surrounded by the Archean Hearne Province to the north, the Trans-Hudson Proterozoic Province to the east, and the Yavapai-Mazatzal Proterozoic Province to the south. The western margin is not well defined due to the lack of outcrop. The northern edge of the Wyoming Province coincides with a geophysical anomaly in Alberta that has been interpreted as a possible suture between the Wyoming Archean Province and the Hearne Province of Canada (Houston, 1993). The eastern boundary of the Wyoming Archean Province is not well exposed, but the geochronologic data show that the ages of the basement have been largely reset during the Proterozoic by the Trans-Hudsonian orogeny (Goldish et al., 1966). The Trans-Hudson Belt between the Wyoming and Superior Provinces is considered the result of a collision that welded the two cratons together (Hoffman, 1989). The southern boundary continues from northern Colorado through northern Utah and into the Basin and Range Province of Nevada. A crustal boundary known as the Cheyenne Belt separates the Wyoming craton from the southern Yavapai-Mazatzal Proterozoic Province (Fig. 2.2).

The main mechanism for the growth of the continental crust surrounding the Archean nucleus in Wyoming occurred principally during: (1) the early and mid-

Figure. 2.1 The Leucite Hills lamproites (0.8 - 3.1 Ma) are located north of Rock Springs in Wyoming, USA. All, except Pilot Butte, lie on the northern flank of the Cretaceous Rock Springs Uplift and on nearby Tertiary sedimentary rocks.

Proterozoic, when most of the sialic basement was generated progressively southward, and (2) the Paleozoic and especially Mesozoic, when the continent grew westward (Ernst, 1988).

The early to mid-Proterozoic Cheyenne Belt, marked by a several kilometer thick, mylonite zone trends from southeastern Wyoming to northeastern Nevada, and represents a continental arc collision zone between the Wyoming Archean Province that once dipped southeastward beneath the Colorado Proterozoic Province (Hills and Houston, 1979; Smithson and Boyd, 1989). The convergent margin is indicated by the intrusion of numerous plutons of gabbroic and calcalkaline affinity. In the western United States, the continental subduction is considered to be the main process responsible for the accretion of the crust during mid-Mesozoic to mid-Cenozoic time (Ernst, 1988). This is reflected by the Nevadan arc that was established from California to British Columbia beginning in late Jurassic. The arc reflects normal-angle subduction of the Farallon beneath North America.

The Wyoming Archean Province is composed mainly of supracrustal rocks and granitic gneiss terrains. The Wyoming craton was established by 2.7 Ga, during which time it was affected by strong compressional deformation and plutonism (Hoffman, 1989). Other tectonic events such as folding, faulting and intrusion also affected the Wyoming Archean Province during the Proterozoic which left behind a highly fractured, sheared, anisotropic and heterogeneous craton (Blackstone, 1983). The fractures were re-activated during the Phanerozoic, but major deformation of the Wyoming basement did not occur until late Cretaceous with the onset of the Laramide Orogeny. The Laramide Belt (Fig. 2.3), extending along the length of the Rocky Mountains, is likely the result of subduction of the Farallon plate under the Wyoming craton. Such subduction may be responsible for the Laramide magmatism from Cretaceous to Eocene (Eggler, 1988). Hamilton (1981), however, attributes the origin of the Laramide Orogeny to the clockwise rotation of the Colorado Plateau plate relative to the interior of the North America continent. This rotation absorbed a very small amount of the convergence between the North America continent and the Farallon plate.

It is generally believed that Laramide deformation uplifted and thrust the Precambrian basement blocks and subsided adjacent basins by as much as 10 km. As a

Fig. 2.2. Cheyenne Belt, located at the southern margin of the Wyoming Archean Province, represents the site of a collisional zone (modified after Hausel, 1996)

Fig. 2.3. Map showing the extent of area affected by Laramide Orogeny (gray shaded area) during late Cretaceous to Eocene. The formation of four large structural units: Green River Basin, Washakie Basin, Great Divide Basin and Rock Springs Uplift formed in the Wyoming basin as a result of Laramide deformation (modified after Stewart, 1978; Blackstone, 1983).

result, four large structural units were formed in the Leucite Hills and nearby areas: the Green River Basin, the Great Divide Basin, the Washakie Basin and the Rock Springs Uplift (Bradley, 1964) (Fig. 2.3).

During the early Tertiary, sediments of the Green River and the Washakie basins were deposited in a widespread lake known as Lake Gosiute located in southwestern Wyoming. Laramide deformation caused downwarping and subsidence of the Green River and the Washakie Basins during which they formed thick sedimentary sequences that make up the Green River and Washakie Formations (Bradley, 1964). The Green River and Washakie Formations are composed of lacustrine, fluvial, and evaporitic Tertiary sedimentary sequences. A pronounced unconformity observed between the Upper Cretaceous and the Lower Tertiary Formations of the Green River and Washakie Basins is attributed to the Laramide Orogeny (Roehler, 1961). The Rock Springs Uplift, a doubly-plunging symmetrical anticline with a northwest-southeast axis, separates the Green River Basin in the west from the Washakie Basin in the east (Fig. 2.3) (Roehler, 1961). By the end of the Cretaceous or Early Paleocene, the large anticline arched up into a dome that allowed the sedimentary rocks from the upper parts of Cretaceous Formations to be eroded (Bradley, 1964). The Rock Springs Uplift reached its present structural relief late in the Tertiary, after the deposition of rocks of Bridger Formation.

The Leucite Hills lamproites cut through the northern flank of the Rock Springs Uplift and rocks belonging to the adjacent Fort Union, Wasatch and Green River Formations (Fig. 2.1). The Upper Cretaceous and Tertiary sedimentary sequences that underlie the lamproites are summarized in Fig. 2.1. Detailed description on the lithology and stratigraphy of the sedimentary sequences can be found in Johnston (1959), Smithson (1959) and Bradley (1964).

The Leucite Hills lamproites are composed of 22 volcanic occurrences that range in age from 0.8 to 3.1 Ma (Mcdowell, 1971; Carmichael et al., 1997). Lamproites are oriented NW-SE and parallel some of the structural features such as the Farson Lineament and a Maastrichtian thrust fault, produced during the Laramide Orogeny (Blackstone, 1983).

The Leucite Hills lamproites crop out as small groups of volcanic cones, lava flows, plugs and dykes. Highly vesicular scoria and cinders are the dominant materials

that make up the tephra cones, whereas flow units and some pyroclastic deposits commonly form composite cones. From field observations, it appears that pyroclastics either post-dated or formed simultaneously with the lava flows. The lava flows vary considerably in thickness from <1 m on the sides of the cones to > 20 m downlope.

Ogden (1979) has classified the flows into simple and compound. The simple flows are greater than 4.5 m thick and exhibit a basal rubble zone, a thin intermediate platy zone and a massive zone. The massive zone contains cooling joints, and shows vesicular texture and flow layering. The outcrops at Deer Butte, Hatcher Mesa and North Table Mountain are examples of simple flows in the Leucite Hills area. The compound lava flows are characterized by successive flows with significant time lapse between them. The time intervals between the flows are long enough to solidify the previous flow and short enough to prevent weathering, erosion or deposition of sediments. Compound flows can be observed on the west end of Zirkel Mesa.

Volcanic necks are common in the Leucite Hills and are the erosional remnants of volcanic vents. The necks show vertical flow layering marked by phlogopite and are composed of pyroclastic rocks and massive lamproite. Dykes normally fill fissures and fractures, trend in a NW direction, and vary in width from 1 m to more than 10 m. The length varies from < 2 m to > 300 m. Table 2.1 presents a list of lamproite occurrences in the Leucite Hills along with their dominant volcanic features.

The remaining part of this section contains a description of localities visited during the summer of 1996. The descriptions are based on the author's observations and those of Kemp (1897), Kemp and Knight (1903), Schultz and Cross (1912), Johnston (1959), Smithson (1959), Ogden (1979) and Mitchell and Bergman (1991).

Badgers Teeth

Located 22 km NE of Rock Springs (Fig. 2.1), the Badgers Teeth lamproite cut the Upper Cretaceous Blair sandstone. In this area, five tooth-like bodies are oriented east-west and talus covers the base of the neck. Kemp (1897) considers the Badgers Teeth outcrop to be part of an elongate volcanic neck built up in a short fissure. The teeth are composed of both agglomerate and massive lamproite (Fig. 2.4). In the largest "tooth" on the east side of outcrop, the agglomerate is overlain by massive

Table 2.1. Summary of the dominant volcanic features observed in the Leucite Hills (Kemp and Knight, 1903; Schultz and Cross, 1921; Ogden, 1979; Keuhner, 1980; Mitchell and Bergman, 1991).

Outcrop Name	Volcanic Features
Badgers Teeth*	Volcanic neck
Black Rock	Volcanic flow without cones
Boars Tusk*	Volcanic neck
Cabin Butte	Volcanic flow without cones
Deer Butte*	Volcanic flow with cones
Emmons Mesa	Volcanic flow with cones
Endlich Hill	(Covered with talus)
Hague Hill	(Covered with talus)
Hallock Butte	Volcanic neck
Hatcher Mesa*	Volcanic flow without cone
Iddings Butte	Volcanic neck, dyke
Matthew Hill*	Volcanic neck
Middle Table Mountain*	Volcanic flow without cones
North Table Mountain*	Volcanic flow without cones
Pilot Butte*	Volcanic flow without cones
South Table Mountain*	Volcanic flow without cones
Spring Butte	Volcanic flow with cones
Steamboat Mountain	Volcanic flow with cones
Weed Butte	Volcanic neck
Wortman Dyke	Dyke
Zirkel Mesa*	Volcanic flow with cones

*** Localities visited and sampled during the course of this study.**

Fig.2.4. Part of the outcrop at Badgers Teeth consisting of agglomerate (top right) and massive (bottom left) lamproites. The dashed red line separates the two types.

lamproite. The lamproite shows elongate vertical vesicles near its margins with the agglomerate. Both agglomerate and massive lamproite contain abundant enclaves of shale and sandstone. On the eastern side of the second largest “tooth”, a flat-lying tuff is sandwiched between overlying massive lamproite and underlying agglomerate lamproite. The tuff is composed of centimeter-sized lamproite fragments contained within a carbonate cement.

Middle Table Mountain

Middle Table Mountain is the smallest, flat-topped mesa in the Leucite Hills and is located between North and South Table Mountain, northeast of Rock Springs (Fig. 2.1). Vertical flow layering in some parts may suggest that Middle Table Mountain is a volcanic neck (Kemp, 1897; Schultz and Cross, 1912; Johnston, 1959). However, the presence of auto-brecciated lava at the base of the south side, flow layering of mica phenocrysts, and columnar jointing capped by blocky lamproite may be evidence for the presence of a lava flow (Ogden, 1979; Mitchell and Bergman, 1991). The lamproites from the Middle Table Mountain and the neighboring North and South Table Mountain all cut through rocks of the Lewis, Almond and Fort Union Formations. Sedimentary xenoliths are abundant in the Middle Table Mountain lamproites. In some cases, partial digestion of the sedimentary xenoliths has produced a striking pattern that reflects intermingling of the sedimentary rocks and the lamproitic melt (Fig. 2.5).

Pilot Butte

The Pilot Butte flow, located 18 km west of Rock Springs (Fig. 2.1), is the only lamproite in the Leucite Hills that has intruded through rocks of the Tertiary Green River Formation. The sides of the flat-topped butte are steep but at the base it changes to a smooth slope covered with talus. Sedimentary enclaves from rocks of the Green River Formation are scattered throughout the lamproite. Some enclaves are rimmed by brownish haloes, the result of interaction with the melt. The mingling pattern of sedimentary rocks and lamproites, similar to that observed at Middle Table Mountain, may indicate partial digestion of the sedimentary xenoliths.

South Table Mountain

South Table Mountain is the highest flat-topped mesa in the Leucite Hills area, and forms an elongate mesa. The lamproite flow overlies rocks of the Upper Cretaceous Lewis and Fort Union Formations. However, the contact between lamproite and country rocks is obscured due to the large talus blocks covering the base. From lower to higher elevations, the lamproite varies from gray to pink and vesicular to non-vesicular. This may indicate the extrusion of two different lavas. Centimeter-sized olivine xenocrysts occur at this locality.

Hatcher Mesa

The Hatcher Mesa flow covers rocks of the Upper Cretaceous Eriscon and Almond Formations. Hatcher Mesa is another example of a volcanic flow in the Leucite Hills. The Hatcher Mesa lamproites are characterized by a distinctive green color, high porosity, and abundant xenoliths. Rounded to angular, lower crustal granulite and upper crustal granite xenoliths are commonly found at this locality.

Boars Tusk

The volcanic neck at Boars Tusk outcrops 55 km NW of Rock Springs (Fig. 2.1). It cuts rocks of the Wasatch Formation and the contact with the country rock is covered with talus blocks. The outcrop at Boars Tusk forms two spires (Fig. 2.6). The dominant volcanic products at Boars Tusk consist of agglomerate, breccia tuff, and massive non-vesicular lamproite. Fragments of cognate lamproites and sedimentary enclaves such as shale, siltstone and sandstone cemented by tuff are abundant in the agglomerates. The massive lamproite is part of a dyke that cross cuts the main body of in a north-south direction.

Deer Butte

Deer Butte is formed by an elongated flow located 28 km northeast of Rock Springs (Fig. 2.1). In addition, two cones are exposed on the northwest and southeast parts of the outcrop. The variation in color, mineralogy and texture may suggest the

Fig. 2.5 Sedimentary enclaves surrounded by reaction rims.
The light-coloured patterns in the lamproite host are partially digested
sedimentary material. False colour image.

Fig. 2.6. Boars Tusk volcanic neck.

presence of multiple flows. The lamproites contain medium to large (few mm to several cm) felsic igneous and sedimentary xenoliths that are more abundant close to the contacts with the sedimentary rocks. The cinder cones are composed of pumice and vesicular lamproite and both contain small xenoliths of igneous and sedimentary rocks.

Zirkel Mesa

Zirkel Mesa, the largest lamproite exposure in the Leucite Hills, is located 40 km northeast of Rock Springs (Fig. 2.1). The Zirkel Mesa lamproite cuts through rocks of the Eriscon and Almond Formations. The contacts between the lamproite and sedimentary rocks are well exposed and in parts, the lava has baked the underlying sedimentary rocks. In addition, a discontinuous basal rubble zone, overlying the sedimentary rocks, grades upwards into massive lava. The Zirkel Mesa lamproites are part of a lava flow that contains five cones aligned in a northwest-southeast direction. Four cones are composed of cinders and one cone is mainly composed of lava overlain with cinders. Both cinders and lava contain igneous and sedimentary xenoliths.

North Table Mountain

The North Table Mountain lava flow outcrops a few hundred meters north of Middle Table Mountain, approximately 60 km northeast of Rock Springs. Lamproite from the North Table Mountain intruded through rocks of the Tertiary Fort Union Formation to the north and Upper Cretaceous Lance Formation to the south. Gray vesicular flows alternate with less vesicular, pink lamproite flows.

CHAPTER 3

LOWER CRUSTAL XENOLITHS

3.1 Introduction

The Leucite Hills lamproites are deposited on Upper Cretaceous to Eocene clastic sedimentary rocks and overlie the thick cratonic basement of the Wyoming Archean Province. The Wyoming Province contains the main exposure of Archean rocks in the western United States. The Archean exposures, mainly confined to crustal blocks uplifted during Laramide time (see Chapter 2, Geology), are scattered widely throughout Wyoming, southwestern Montana, and parts of Idaho, Utah and South Dakota (Fig. 2.2). The lack of continuous exposure of the Wyoming Archean rocks implies that exposed areas are not necessarily representative of the province as a whole.

Information on the lithology, chemical composition and evolutionary history of the Archean basement, particularly that of the lower crust, can be obtained through three independent methods: exposed granulite-facies terrains, entrained xenoliths in volcanic rocks, and geophysical surveys. Each method has its own advantages and limitations. Granulite terrains can provide direct observations of the metamorphic gradations, contact relationships, lithological abundances and temporal variations. However, granulite terrains have generally experienced intensive erosion and uplift during their history and hence may no longer fully reflect the physical and chemical characteristics of the lower crust.

Xenoliths of lower crustal materials, carried to the surface by mantle-derived magmas, provide the most direct information about the lithology and composition of the deep crust, even though the spatial relationships between samples might not be clear. Lower crustal xenoliths are relatively rare from Archean cratons and only a few examples have been reported. Included among these are Bearpaw Mountains, Montana (Collerson et al., 1988), Elovu island, Kola (Kempton et al., 1995), Lace Kimberlites, South Africa (Dawson and Smith, 1987), Loch Roag, Scotland (Halliday et al., 1993), Mand Shield, Liberia (Toft et al., 1989), and the Snake River Plains, Idaho (Leeman et al., 1985). In the Leucite Hills area, a large number of lower crustal xenoliths from the Wyoming Archean

Craton have been entrained by the lamproites at Hatcher Mesa. These xenoliths provide the most direct access to the present day composition of the Wyoming Archean crust.

Crustal xenoliths were collected from Hatcher Mesa during the summer of 1996 and studied for their petrography, mineral composition, whole rock major and trace element chemical composition and isotopic composition. These data are used to interpret the nature of the lower crust overlying the mantle source of the Leucite Hills lamproites, and to estimate the pressure and temperature conditions that existed during the formation of the lower crust beneath the Leucite Hills. Because major crustal formation is assumed to be the result of partial melting in the upper mantle, by determining the age of the lower crustal xenoliths, information can be obtained about the melt removal episode from the Wyoming sub-continental mantle. The mechanisms for melt production (i.e., subduction versus plume activity) and crystallization (i.e., basaltic underplating versus restite) are also of importance.

3.2 Previous studies

The crustal xenoliths from the Leucite Hills lamproites were first observed and mentioned by Cross (1897). Smithson (1959) reported the presence of granite, norite and mica schist inclusions from the Hatcher Mesa outcrop. It was not until 1978 that detailed studies were conducted by Kay et al. (1978) who determined the mineral compositions, trace element geochemistry and P-T conditions of a number of crustal xenoliths from Hatcher Mesa and Cabin Butte. These authors identified some of the xenoliths as meta-igneous granulites equilibrated at temperatures of about 875°C and pressures <10 kb (~30 km), consistent with an origin in the lower crust. Barton and Bergen (1981) presented a detailed study of the composition of clinopyroxenes from the Hatcher Mesa lavas and crustal xenoliths and concluded that clinopyroxene megacrysts in the Leucite Hills lamproites originated from the disaggregation of clinopyroxene-rich xenoliths. Barton and Bergen (1981) also suggested that such xenoliths are either fragments of metasomatised upper mantle or fragments of a crystallized magma of unknown composition. Sperr (1985) collected a variety of xenoliths from the Leucite Hills lamproites ranging from lower crustal granulites to upper crustal sedimentary and granitic rocks and reconstructed the stratigraphy of the crust underlying the Leucite Hills volcanic

rocks. He concluded that the lower crust consists of anorthosites at the deepest level overlain by granulites. The composition of mineral phases from eleven xenoliths from Hatcher Mesa was examined by Perrota (1997). The chemical compositions of minerals show linear trends on bivariate diagrams which suggested fractional crystallization during the evolution of the parental magma. Lester and Framer (1998) presented a comprehensive literature review of data from the lower crustal and upper mantle xenoliths in the Archean Wyoming and Proterozoic Colorado provinces. They concluded that beneath both provinces the lower continental crust is dominantly basic in composition and that the upper mantle is a heterogeneous combination of fertile and enriched peridotite.

3.3 Sample description and petrography

Hatcher Mesa is known for its abundant igneous xenoliths (Barton and Bergen 1981; Sperr 1985). These range from granitic upper crustal to ultramafic/mafic and felsic lower crustal rocks. The xenoliths vary in size from 1 cm to 50 cm. The upper crustal samples are not discussed here because they are pervasively altered. The lower crustal xenoliths are abundant, and dominantly fresh.

Granulite is a metamorphic term for plagioclase-bearing rocks that may also contain pyroxene(s) and garnet and have been equilibrated to granulite facies conditions (O'Reilly et al., 1989). The term granulite is used here to refer to the mafic-ultramafic xenoliths that show evidence of granulite facies conditions and originated from the lower crust. High-grade metamorphism of the igneous rocks is usually reflected by the development of straight to gently-curving grain boundaries, triple junctions with 120° angles, deformation of polysynthetic twinning in plagioclase, broken grains, and overall elimination of igneous microstructure (Vernon, 1970; Wysoczanski et al., 1995). The polygonal interlocking grains with 120° triple junctions, the production of granoblastic texture, and the granulation of primary plagioclase are normally considered typical of granulite facies textures (Joswiak 1992). The polygonal granoblastic textures are also considered typical of rocks that have annealed over a long period of time at high pressures (Kempton et al., 1997).

The lower crustal xenoliths from Hatcher Mesa exhibit granoblastic texture, deformed twins in plagioclases, and evidence of subsolidus high pressure – temperature recrystallization such as polygonal and interlocking grains with 120° triple junctions (Fig.3.1). They are coarse grained and vary from dark and light gray to chalky white in colour. The lower crustal xenoliths tend to be small and rounded (6 cm) or subrounded to flattened (5 cm $\times$ 2 cm). It seems that the lower crustal xenoliths have not reacted with the lamproite host as is evidenced by their sharp contacts (Fig.3.2) and the absence of reaction rims. The lower crustal xenoliths are usually equigranular and some samples exhibit bands of mafic and felsic minerals. Banding seen in some of the Leucite Hills lower crustal xenoliths is defined by distinctive layers made up of plagioclase and pyroxene.

On the basis of their mineralogy, xenoliths from the Leucite Hills indicates can be divided into ultramafic and mafic granulites. The dominant minerals found in the lower crustal xenoliths from the Leucite Hills are plagioclase, pyroxene and amphibole. No garnet-bearing xenoliths have been found. However, orthopyroxene is a common mineral

Fig. 3.1 Interlocking plagioclase with granoblastic texture (Sample 157 - 30). Field of view 50 mm.

Figure 3.2 An outcrop of Hatcher Mesa lamproite with lower crustal xenoliths. The xenoliths are easily removed from the weathered host lamproite. Note also the sharp contact between the xenolith near the hammer and the host volcanic rock.

phase in the xenoliths and it is commonly used to indicate the initiation of the granulite grade metamorphism (Miyashiro, 1994). The lower crustal xenoliths with textural evidence of high-grade metamorphism and the presence of orthopyroxene, clinopyroxene and plagioclase are normally considered two-pyroxene granulites (Irving, 1974a).

Fifteen mafic-ultramafic xenoliths, in addition to two anorthosite and two felsic xenoliths, were selected for detailed petrographic study. The estimated modal percentages of the xenoliths are presented in Table 3.1. The lower crustal xenoliths have been classified on the basis of the modal proportions of clinopyroxene, orthopyroxene and plagioclase (Fig. 3.3). The ultramafic xenoliths are classified as orthopyroxenite (156-2 HAM), clinopyroxenite (156-2b HAM), and websterite (157-25 HAM). Olivine is absent as a primary mineral phase in the ultramafic granulites, but occurs in reaction rims around pyroxene. The olivine becomes coarser grained where orthopyroxene is in the vicinity of, or in contact with, the lamproite host. Orthopyroxene forms subhedral to anhedral (0.4 mm - 1mm), equant grains with abundant 120° triple junctions. Clinopyroxene is subhedral to anhedral (1 mm in size) and some grains show exsolution lamella of either ilmenite or orthopyroxene. Amphibole can also be present. It is normally

fresh and forms equant, euhedral (1.2 mm) and anhedral (0.2 mm) grains. The anhedral crystals occur as interstitial phases between the orthopyroxene (Fig. 3.4). Mica forms euhedral to subhedral laths (0.1 - 0.2 mm). Oxide minerals (e.g., ilmenite and magnetite) and apatite are commonly found as interstitial grains in the ultramafic xenoliths.

Mafic granulites are the most abundant of the xenoliths found in the Hatcher Mesa area. They display a wide variation in modal mineralogy. On the basis of the nomenclature proposed by Irving (1974a), they can be considered gabbroic granulites and two-pyroxene granulites (Fig. 3.3). Two samples are also classified as anorthosites. The gabbroic granulite is composed of clinopyroxene and plagioclase in equal amounts. The two-pyroxene granulites, the most abundant of the mafic xenoliths, contain both orthopyroxene and clinopyroxene in addition to plagioclase and/or amphibole. Plagioclase and to a lesser extent pyroxene and oxides make up the mineralogy of the anorthosite xenoliths.

The mafic xenoliths display textures that vary from allotriomorphic granular to granoblastic. Plagioclase commonly occurs as either large crystals (0.8 - 1 mm) accompanied by smaller granulated grains (0.2 mm) or medium-sized (0.4 - 0.6 mm) interlocking crystals (Fig. 3.2). Most grains can be considered as xenoblastic and commonly exhibit deformation twins and undulatory extinction. Plagioclase also shows polysynthetic twinning and perthitic intergrowths. In rare cases, the plagioclase has been transformed to a light-brown coloured fibrous assemblage or to dark opaque material. Orthopyroxene is subhedral to anhedral (0.4 - 0.6 mm), equant or elongated, and pleochroic (pale pink to pink). Orthopyroxene can also occur in clusters. The straight to gently curving grain boundaries can be rimmed by fine-grained olivine grains. Orthopyroxene is usually subordinate to the clinopyroxene. Clinopyroxene, 0.4 - 0.6 mm in size, is light green in colour and anhedral. Dark-brown ilmenite occurs locally as exsolution lamella in clinopyroxene grains.

Amphibole occurs in a number of mafic xenoliths and it can be fresh (157-4, -9, -28 HAM), partly altered (157-8 HAM) or completely altered (157-3, -5, -6, -30, -36, -38 HAM). Fresh amphibole is anhedral (0.2 - 0.8 mm) with light brown to dark

Table 3.1 Modal mineralogy of selected lower crustal xenoliths from Hatcher Mesa, Leucite Hills.

Sample #	opxenite			cpxenite	websterite	one px granulite			anorthosite			two px granulite		
	156-2	156-2b	157-25			157-31	157-7	157-29	157-23	157-2	157-4	157-5	157-6	
Pl	0	0	0	72	42	94	90	30	42	55	33			
Cpx	0	75	48	21	53	0	6	15	9	4	7			
Opx	74	0	39	0	0	0	1	18	20	2	5			
Amp	25	0	5	0	0	2	0	35	28	31	50			
Bt	0	21	6	0	0	0	0	0	0	0	0			
Qtz	0	0	0	0	0	0	0	0	0	0	0			
Kfs	0	0	0	0	0	0	0	0	0	0	0			
Oxide	1	1.5	1	5	4	2	2	1	1	2	2			
Ap	0	0	0	0	0	0	0	0	0	0	0			
Accessory	0	0.5	1	1	0	0	0	0	0	0	1			
Other	0	2	0	2	1	2	1	1	0	6	2			
Total	100	100	100	101	100	100	100	100	100	100	100			

Amp = Amphibole, Ap = Apatite, Bt = Biotite, Cpx = Clinopyroxene, Kfs = K-feldspar, Opx = Orthopyroxene, Qtz = Quartz
(Mineral abbreviations after Kretz, 1983).

opxenite = orthopyroxenite, cpxenite = clinopyroxenite, one px granulite = one pyroxene granulite,
two px granulite = two pyroxene granulite,

Table 3.1 (continued)

Sample #	two px granulite			felsic xenolith		
	157-8	157-9	157-12	157-22	157-24	157-28
Pl	45	48	55	50	53	30
Cpx	12	16	20	9	15	22
Opx	10	10	12	23	20	16
Amp	25	21	8	14	6	28
Bt	0	0	0	0	0	1
Qtz	0	0	0	0	0	0
Kfs	0	0	0	0	0	0
Oxide	0	2	2	1	4	1
Ap	0	0	0	0	0	0
Accessory	3	tr	1	0	0	0
Other	5	3	2	3	2	2
Total	100	100	100	100	100	100

Figure 3.3 Classification scheme of lower crustal xenoliths based on the modal abundances of plagioclase, orthopyroxene and clinopyroxene (Nomenclatures from Irving, 1974a).

Figure 3.4 Amphibole (orange to brown) occurs as interstitial phase filling the spaces between orthopyroxene(colourless to creamy) crystals. Field of view 20 mm.

brown pleochroism. The altered amphibole appears opaque, probably due to the presence of numerous magnetite inclusions. In some samples (e.g., 157-6), amphibole can form up to 50% of the xenoliths, surrounding other minerals such as plagioclase and pyroxene. Amphibole normally occurs as large equant grains. Ilmenite is a ubiquitous accessory mineral in all of the mafic xenoliths and is particularly concentrated around the pyroxene and amphibole grains.

Gabbroic granulite (157-7 HAM), the rarest of the mafic xenoliths, contains no amphibole or orthopyroxene. It consists of fresh pale green anhedral clinopyroxene (0.2 - 0.8 mm) and subhedral to anhedral plagioclase (0.4 - 0.8 mm). The clinopyroxene exhibits exsolved, opaque, iron oxide. The polygonal texture with triple junctions is common in the gabbroic granulite.

The mafic and ultramafic lower crustal xenoliths from the Leucite Hills also exhibit relict cumulate textures. The dominantly large tabular crystals are considered cumulate in origin and the interstitial phases are interpreted as intercumulus. In the mafic granulites, pyroxene is the major cumulus phase whereas plagioclase occurs as both a cumulus and intercumulus phase (Fig. 3.5). Other intercumulus phases include oxide minerals and amphibole. In the ultramafic xenoliths, pyroxene is a dominantly cumulus phase and amphibole occurs as both cumulus and intercumulus phases. Overall, the variation in rock types (i.e., from ultramafic to mafic) and the presence of anorthosite resembles the range found in layered intrusions.

The felsic xenoliths (157-27 HAM, 157-41 HAM) contain high abundances of felsic minerals such as quartz and feldspar. Feldspar, consisting of equant interlocking plagioclase and orthoclase, is the most abundant mineral and can crystallize either as large (1 - 2 mm) euhedral to subhedral grains or as small anhedral crystals in the groundmass (0.1 - 0.3 mm). Feldspar commonly shows perthitic or antiperthitic intergrowths. Quartz occurs as elongate, anhedral, fractured porphyroclasts (0.5mm) and shows curved grain boundaries. Orthopyroxene is the dominant mafic mineral present in the felsic xenoliths and forms subhedral to anhedral xenoblastic crystal (0.2 - 2 mm). Ti-rich magnetite and minor apatite are also present in the felsic xenoliths. Lower crustal xenoliths with similar mineral assemblages to the felsic xenoliths have been reported from the northern part of Wyoming Craton (Joswiak, 1992) and the Snake River Plain

volcanic field (Leeman et al., 1985).

Fig.3.5 An anhedral intercumulus plagioclase surrounded by cumulus clinopyroxene in a mafic xenolith (Sample 157-4). Field of view 100 mm.

3.4 Secondary alteration

Although the lower crustal xenoliths from Hatcher Mesa are dominantly fresh, secondary alteration can be seen in a number of xenoliths, particularly at their margins. Most pyroxenes at the margins of the xenoliths have been partly transformed to a fine-grained assemblage of olivine, clinopyroxene, orthopyroxene, magnetite, apatite and barite set in a groundmass of amphibole, leucite and K-feldspar (Fig.3.6). Plagioclase, at the margins of the xenoliths, has been transformed to an unidentified light to dark brown fibrous material. Based on textural observations, it seems likely that the margins of some xenoliths have interacted with the lamproite melt.

Fig.3.6 Replacement of clinopyroxene by a fine-grained mineral assemblage. Field of view 50 mm.

3.5. Mineral chemistry

Minerals from the lower crustal xenoliths were analyzed for major element concentrations using the electron microprobe at Carleton University. The analytical techniques are described in Appendix B. Tables 3.2 – 3.9 list the chemical composition and unit formula of mineral phases from the xenoliths and their lamproite host.

Feldspars

The chemical compositions of feldspar from the lower crustal xenoliths are presented in Table 3.2. Most of the feldspar is plagioclase. Potassium feldspar is found only in the felsic xenoliths and in the secondary fine-grained aggregates that replace pyroxene at the margins with the lamproite host. Most plagioclase is unzoned and chemically homogeneous (Fig.3.7). The anorthite content of plagioclase ranges from oligoclase-andesine (An_{22-29}) in the felsic xenoliths to andesine-labradorite (An_{40-55}) in the mafic xenoliths and labradorite-bytownite (An_{65-68}) in the anorthosite xenolith. The composition of the K-feldspar in the felsic xenoliths is mainly orthoclase (Or_{80-87}). Two of the feldspar data points from the felsic xenolith plot out of the compositional range

Table 3.2 Chemical composition of feldspar from the lower crustal xenoliths, Leucite Hills.

Sample#	157-4	157-5	157-6	157-7	157-8	157-9	157-12
SiO ₂	54.93	55.91	55.47	55.68	55.45	55.97	55.44
Al ₂ O ₃	28.88	27.48	28.16	27.45	28.45	28.97	28.06
FeO _T	0.12	0.12	0.16	0.06	0.00	0.00	0.00
BaO	0.00	0.00	0.00	0.00	0.00	0.00	0.00
CaO	10.69	9.61	10.12	9.85	10.63	10.53	9.82
Na ₂ O	5.41	5.41	5.82	4.93	5.71	6.05	5.83
K ₂ O	0.32	1.26	0.54	1.61	0.37	0.28	0.51
Total	100.35	99.79	100.27	99.58	100.61	101.80	99.66
Si	9.9	10.1	10.0	10.1	9.9	9.9	10.0
Al	6.1	5.9	6.0	5.9	6.0	6.0	6.0
Fe ²	0.0	0.0	0.0	0.0	0.0	0.0	0.0
Ba	0.0	0.0	0.0	0.0	0.0	0.0	0.0
Ca	2.1	1.9	2.0	1.9	2.0	2.0	1.9
Na	1.9	1.9	2.0	1.7	2.0	2.1	2.0
K	0.1	0.3	0.1	0.4	0.1	0.1	0.1
Ab	46.9	46.8	49.5	43.1	48.3	50.2	50.3
An	51.2	46.0	47.5	47.6	49.7	48.3	46.8
Or	1.8	7.2	3.0	9.3	2.1	1.5	2.9

Ab = Albite, An = Anorthite, Or = Orthoclase

Table 3.2 (continued)

Sample#	157-22	157-29	157-36	157-38
SiO ₂	55.77	51.14	52.77	57.72
Al ₂ O ₃	27.57	31.49	29.33	26.43
FeO _T	0.00	0.13	0.50	0.05
BaO	0.00	0.00	0.00	0.00
CaO	9.47	13.71	11.87	8.15
Na ₂ O	5.98	4.09	4.44	6.52
K ₂ O	0.36	0.00	0.63	0.50
Total	99.15	100.56	99.54	99.37
Si	10.1	9.3	9.6	10.4
Al	5.9	6.7	6.3	5.6
Fe ₂	0.0	0.0	0.1	0.0
Ba	0.0	0.0	0.0	0.0
Ca	1.8	2.7	2.3	1.6
Na	2.1	1.4	1.6	2.3
K	0.1	0.0	0.1	0.1
Ab	52.2	35.1	38.9	57.4
An	45.7	64.9	57.5	39.7
Or	2.1	0.0	3.6	2.9

Table 3.2 (continued)

Sample#	157-27-1	157-27-2	157-27-3	157-27-4	157-27-5	157-27-6
SiO ₂	65.80	64.76	61.96	63.17	60.57	63.50
Al ₂ O ₃	18.77	18.65	23.14	23.54	24.36	23.40
FeO _T	0.06	0.04	0.03	0.03	0.02	0.02
BaO	0.99	0.88	0.16	0.02	0.00	0.02
CaO	0.03	0.04	4.64	4.86	6.07	4.76
Na ₂ O	1.41	1.26	4.25	4.21	7.27	8.55
K ₂ O	14.57	14.89	7.04	7.25	1.02	0.97
Total	101.63	100.52	101.22	103.08	99.31	101.22
Si	12.0	11.9	11.1	11.1	10.9	11.1
Al	4.0	4.0	4.9	4.9	5.1	4.8
Fe ₂	0.0	0.0	0.0	0.0	0.0	0.0
Ba	0.1	0.1	0.0	0.0	0.0	0.0
Ca	0.0	0.0	0.9	0.9	1.2	0.9
Na	0.5	0.4	1.5	1.4	2.5	2.9
K	3.4	3.8	1.6	1.6	0.2	0.2
Ab	12.8	12.9	37.1	36.1	64.4	72.3
An	0.2	0.1	22.4	23.0	29.7	22.3
Or	87.0	80.0	40.5	40.9	5.9	5.4

Figure 3.7 Microprobe traverses along margin (m) – core (c) – margin (m) of plagioclase from lower crustal xenoliths. (a) sample 157-6, (b) sample 157-38. Considering the analytical uncertainty, no zonation is observed in plagioclase feldspar.

Figure 3.8 Feldspar classification for the lower crustal xenoliths from the Leucite Hills. The composition of plagioclase from the mafic xenoliths is andesine to labradorite. The plagioclase from the felsic xenolith is more Na-rich than that from the mafic xenoliths. Two feldspars from the felsic xenoliths plot outside of the compositional range established for feldspar classification. Mafic xenoliths ○, felsic xenoliths ●, fine-grained assemblage □. Classification diagram after Deer et al. (1978).

exhibited by igneous rocks and fall almost in the center of the Or-Ab-An ternary diagram (Fig. 3.8).

The feldspar found in the fine-grained secondary assemblage replacing clinopyroxene is K-feldspar. Because the fine-grained secondary feldspar is found at the margins of several xenoliths, it probably formed as the result of interaction between the xenolith and the melt. This is supported by the fact that when plagioclase occurs close to the contact with the lamproite host, it is usually rimmed by a light brown K – rich fibrous material.

Amphibole

Amphibole in the lower crustal xenoliths is usually pleochroic (light brown to dark brown). In some of the samples, the amphibole has been replaced by magnetite and some unknown silicate minerals. The pseudomorphed amphibole appears opaque under the petrographic microscope. Amphibole can also occur as rounded, small grains that has replaced pyroxene at the margins of xenoliths.

The chemical composition, unit formula and end-member composition of the amphibole are reported in Table 3.3. In general, the fine-grained amphibole and lamproite amphibole are more alkali-rich than amphibole from the granulite (Fig. 3.9a). On the Mg/Mg + Fe²⁺ versus Si classification diagram (Fig. 3.9b), most of the amphiboles from the lower crustal xenoliths are classified as edenite hornblende to Fe-rich pargasitic hornblende. In contrast, amphibole from the lamproite host, originally named magnophorite by Prider (1939), is the K-rich variety of richterite (Mitchell and Bergman, 1991) and is classified as titanian potassium richterite and titanian potassium magnesio-katophorite (Fig. 3.9c). The secondary fine-grained amphibole is enriched in K and Na and depleted in Ca compared to the two other amphibole types and can be classified as K-rich richterite and eckermannite (Fig. 3.9c & d).

The fresh and pseudomorphed amphiboles from the lower crustal xenoliths have similar major element compositions. However, these amphiboles are generally more enriched in Al₂O₃ compared with the amphibole from host lamproite and the secondary fine-grained amphibole at the margins of the xenoliths (Fig. 3.10a). The higher Al content

Table 3.3 Chemical composition of amphibole from the lower crustal xenoliths and the Leucite Hills lamproite.

Sample#	156-2	157-22	157-25	157-28	157-3	157-5	157-6	157-8	157-9	156-2b-1-1	157-8-3-2	157-12-8
SiO ₂	41.62	42.18	43.65	43.73	43.47	42.67	42.96	42.33	42.7	54.85	55.23	57.08
TiO ₂	0.99	2.31	2.09	1.7	1.39	2.01	1.64	2.21	2.02	0.54	0	0.29
Al ₂ O ₃	13.83	12.3	10.94	11.15	11.26	11.68	11.47	12.7	12.14	0.45	0.56	0.52
FeO _T	11.67	13.11	10.85	10.19	10.28	11.62	11.21	12.14	11.22	9.15	6.46	8.87
Cr ₂ O ₃	0.35	0.37	0.52	0.42	0	0.15	0.15	0.12	0.16	0	0	0
MnO	0	0	0.1	0	0	0.21	0	0.09	0.17	0.15	0.21	0.18
MgO	13.25	12.73	13.59	13.82	13.8	12.89	13.95	12.83	13.45	17.39	19.25	19.61
CaO	11.97	12.11	11.53	11.89	13.74	12.86	11.77	11.86	13.89	2.5	4.6	2.59
Na ₂ O	2.73	1.63	1.8	1.48	2.83	1.98	2.55	1.89	1.5	6.9	6.71	7.97
K ₂ O	1.38	1.42	1.52	1.45	1.34	1.23	1.6	1.6	1.51	3.34	3.35	3.37
Total	97.79	98.16	96.59	95.83	98.11	97.3	97.3	97.77	98.76	95.27	96.37	100.48
TSi	8.187	6.410	6.657	6.512	6.563	6.340	6.370	6.261	6.256	8.060	7.974	7.958
TAI	0.000	1.590	1.343	1.488	1.437	1.660	1.630	1.739	1.744	0.000	0.026	0.042
CAI	3.204	0.612	0.622	0.467	0.565	0.384	0.373	0.473	0.351	0.078	0.069	0.043
CCr	0.054	0.044	0.063	0.049	0.000	0.018	0.018	0.014	0.019	0.000	0.000	0.000
CFe3	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000
CTi	0.146	0.264	0.240	0.190	0.158	0.225	0.183	0.246	0.223	0.060	0.000	0.030
CMg	1.595	2.884	3.090	3.068	3.106	2.855	3.083	2.829	2.938	3.810	4.143	4.076
CFe2	0.000	1.196	0.985	1.225	1.171	1.444	1.344	1.438	1.375	1.053	0.780	0.851
CMn	0.000	0.000	0.000	0.000	0.000	0.026	0.000	0.000	0.021	0.000	0.007	0.000
CCa	0.000	0.000	0.000	0.000	0.000	0.048	0.000	0.000	0.074	0.000	0.000	0.000
BMg	2.290	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000
BFe2	1.920	0.471	0.398	0.044	0.127	0.000	0.046	0.064	0.000	0.072	0.000	0.184
BMn	0.000	0.000	0.013	0.000	0.000	0.000	0.000	0.011	0.000	0.019	0.019	0.021
BCa	0.000	1.529	1.589	1.897	1.873	1.999	1.870	1.880	2.000	0.394	0.712	0.387
BNa	0.000	0.000	0.000	0.059	0.000	0.001	0.084	0.045	0.000	1.516	1.270	1.408
ACa	2.523	0.443	0.295	0.000	0.349	0.000	0.000	0.000	0.107	0.000	0.000	0.000
ANa	1.041	0.480	0.532	0.368	0.828	0.570	0.649	0.497	0.426	0.450	0.609	0.746
AK	0.346	0.275	0.296	0.275	0.258	0.233	0.303	0.302	0.282	0.626	0.617	0.599

Samples 156-2b-1-1, 157-8-3-2, and 157-12-8 are from the fine-grained assemblages.

Table 3.3 (continued)

	lamproite 156-2-h	lamproite 157-4-h	lamproite 157-8-h	lamproite 157-9-h
Analysis				
SiO ₂	53.12	52.85	52.27	51.52
TiO ₂	5.68	7.85	6.35	8.48
Al ₂ O ₃	0.79	0.55	0.93	0.97
FeO _T	2.71	4.57	3.14	4.38
Cr ₂ O ₃	0.09	0.07	0.01	0.09
MnO	0.1	0.13	0.03	0.08
MgO	20.01	17.14	19.74	17.96
CaO	6.14	5.48	6.37	5.82
Na ₂ O	4.01	4.57	4.32	4.79
K ₂ O	4.61	4.73	4.29	4.16
Total	97.26	97.94	97.45	98.25
TSi	7.547	7.533	7.44	7.33
TAI	0.132	0.092	0.156	0.163
CAI	0	0	0	0
CCr	0.01	0.008	0.001	0.01
CFe3	0	0	0	0
CTi	0	0	0	0
CMg	4.238	3.642	4.189	3.809
CFe2	0.322	0.545	0.374	0.521
CMn	0.012	0.016	0.004	0.01
CCa	0.417	0.79	0.433	0.65
BMg	0	0	0	0
BFe2	0	0	0	0
BMn	0	0	0	0
BCa	0.517	0.047	0.539	0.237
BNa	1.105	1.263	1.192	1.321
ACa	0	0	0	0
ANa	0	0	0	0
AK	0.836	0.86	0.779	0.755

Figure 3.9 Amphibole classification schemes. Lower crustal xenoliths ○, secondary fine-grained assemblage □, and lamproite host ◆. (a) amphibole from the lower crustal xenoliths is more Ca-rich than that of the fine-grained assemblage and lamproite host, (b) amphibole from the lower crustal xenoliths is classified as edenite hornblende and Fe-rich pargasitic hornblende, (c) amphibole from the Leucite Hills lamproites is classified as richterite and magnesio-kataphorite, and (d) the fine-grained amphibole that has replaced pyroxene at the edge of the lower crustal xenoliths is classified as eckermannite. Nomenclature of amphiboles after Leake (1985).

Figure 3.10 (a) FeO versus MgO , and (b) TiO_2 versus Al_2O_3 variation diagrams for amphibole from the Leucite Hills lower crustal xenoliths, secondary fine-grained assemblage, and lamproite host. Amphibole from the lower crustal xenoliths is noticeably more enriched in FeO and Al_2O_3 than that of the secondary fine-grained assemblage and the Leucite Hills lamproites.

in the amphibole from the granulite xenoliths may result from elevated pressures at the time of amphibole crystallization in the lower crust. The TiO_2 content of the granulite and secondary fine-grained amphibole is three orders of magnitude lower than that from the lamproite host. Amphibole from the lower crustal xenoliths displays a slightly negative correlation on the FeO versus MgO plot (Fig. 3.10b) and show higher FeO and lower MgO abundances than amphiboles from the fine-grained assemblage and the lamproite.

Olivine

Olivine does not occur as a primary phase in the mafic to ultramafic xenoliths. This may indicate that olivine, if present, probably became unstable during an increase in pressure involving the reaction $\text{An} + \text{Fo} \rightarrow \text{En} + \text{Di}$ (Kushiro and Yoder, 1966). The mafic to ultramafic granulites contain secondary olivine that crystallizes either as rims enclosing orthopyroxene or as rounded grains accompanied by a fine-grained mineral assemblage consisting of pyroxene, amphibole, Fe-oxide and/or apatite after primary pyroxene.

At the margin of the xenoliths and close to the lamproite host, the olivine that forms the rims around pyroxene becomes coarser and the secondary fine-grained olivine becomes more abundant. The chemical composition of olivine is reported in Table 3.4. It appears that the globular olivine from the secondary fine-grained assemblage (Fo_{50-75}) is possibly more Mg-rich than the olivine that rims the orthopyroxene (Fo_{40-70}). In the mafic to ultramafic xenoliths, no correlation is observed between the xenolith type and the composition of the fine-grained, secondary olivine. However, olivine shows a wide range in composition and a strong negative correlation exists between MgO and FeO (Fig. 3.11).

Mica

Mica has been found in only two samples from the lower crustal xenoliths (156-2b-1, 157-25). Mica is typically associated with amphibole and occurs along grain boundaries. The deep red to brown pleochroism displayed by the mica can be attributed to high concentrations of Ti. The chemical composition and the unit formula of mica for the lower crustal xenoliths are reported in Table 3.5. The chemical composition of mica

Table 3.4 Chemical composition of secondary olivine.

Sample #	156-2	157-6	157-23	156-2b	157-5	157-8	157-12	157-36
SiO ₂	36.21	34.21	33.87	37.38	36.18	37.29	38.36	35.28
FeO	32.68	44.42	48.52	31.41	34.03	26.46	23.67	41.24
MnO	0.55	0.84	0.77	0.80	1.01	0.77	0.67	0.94
MgO	29.97	19.63	17.18	31.77	28.38	34.48	36.70	22.80
CaO	0.15	0.32	0.58	0.19	0.18	0.12	0.43	0.15
NiO	0.00	0.00	0.00	0.00	0.00	0.00	0.39	0.00
Total	99.56	99.42	100.92	101.55	99.78	99.12	100.22	100.41
Si	0.999	1.007	1.002	1.002	1.004	1.002	1.006	1.008
Fe ²	0.754	1.094	1.200	0.704	0.790	0.594	0.519	0.985
Mn	0.013	0.021	0.019	0.018	0.024	0.018	0.015	0.023
Mg	1.232	0.861	0.758	1.269	1.174	1.381	1.434	0.971
Ca	0.004	0.010	0.018	0.005	0.005	0.003	0.012	0.005
Ni	0.000	0.000	0.000	0.000	0.000	0.000	0.008	0.000
Cations	3.002	2.993	2.997	2.998	2.997	2.998	2.994	2.992
Fe	0.380	0.560	0.610	0.360	0.400	0.300	0.270	0.500
Mg	0.620	0.440	0.390	0.640	0.600	0.700	0.730	0.500

Figure 3.11 FeO versus MgO variation diagram for olivine that rims pyroxene and for fine-grained, secondary olivine from the mafic, lower crustal xenoliths, Leucite Hills.

from the lamproite host is listed in Table 4.5. Mica from the granulite xenoliths is enriched in Fe while that from the lamproite is Mg-rich, thus the former one is biotite.

The granulite biotite also shows higher concentrations of TiO_2 and Al_2O_3 than the phlogopite from the enclosing lamproite (Fig. 3.12). The higher Al_2O_3 concentrations of biotite from the lower crustal xenoliths reflect its crystallization under elevated pressures.

Leucite

Leucite does not crystallize as a primary mineral phase in the lower crustal xenoliths but occurs as a secondary phase along with the secondary fine-grained assemblage that replaces pyroxene at the margins of the xenoliths. Chemically, this secondary leucite is different from that found in the lamproites (Table 3.6). The latter has higher Al_2O_3 and lower FeO contents than the former (Fig. 3.13).

Oxide Minerals

Table 3.7 lists the composition and unit formula of oxide minerals from the lower crustal xenoliths. The oxide mineral phases identified in the lower crustal xenoliths are ilmenite, titanian magnetite ($\text{TiO}_2 > 10\%$), magnetite ($\text{FeO}_T > 90\%$) and hematite. They occur as inclusions in the silicates or as interstitial grains. Oxide minerals can also be found in the fine-grained, secondary aggregates occurring at the margins of the xenoliths. The iron oxides in the fine-grained aggregates are normally more enriched in Mg (titanian ilmenite) than those occurring as single or exsolved crystals within the granulite xenoliths.

Commonly, ilmenite and magnetite crystallize as well-developed, micro-intergrowths that can be rimmed by a thin layer of hematite. The ilmenite micro-intergrowths have been documented from a variety of igneous and metamorphic rocks. According to Lattard (1995), the magnetite-ilmenite intergrowth is not only the product of redox reaction within magnetite, but could also be due to the vacancy relaxation in magnetite causing exsolution.

Table 3.5 Chemical composition of mica from the lower crustal xenoliths, Leucite Hills.

Sample #	156-2b-1	157-25
SiO ₂	37.53	37.11
TiO ₂	5.09	5.16
Al ₂ O ₃	14.05	13.57
Cr ₂ O ₃	0.15	0.53
FeO _T	12.35	11.79
MgO	16.14	15.81
BaO	0.00	0.63
K ₂ O	10.11	9.72
Total	95.42	94.32
Si	5.557	5.574
Ti	0.567	0.583
Al	2.450	2.400
Fe ²	1.529	1.481
Cr	0.018	0.063
Mg	3.563	3.540
Ba	0.000	0.037
K	1.910	1.862
Cations	15.594	15.540
O	24.000	24.000
Fe	0.300	0.290
Mg	0.700	0.710

Table 3.6 Chemical composition of leucite from the lower crustal xenoliths, Leucite Hills.

Sample #	157-11-3	157-7-3	157-28-1	157-30-1
SiO ₂	55.74	55.56	54.15	54.72
Al ₂ O ₃	20.78	20.28	22.97	23.39
Fe ₂ O ₃ T	1.65	2.24	0.56	0.21
K ₂ O	20.66	20.64	21.02	21.3
Total	98.83	98.72	98.7	99.62
Si	2.055	2.058	1.996	1.995
Al	0.902	0.884	0.997	1.004
Fe ²	0.051	0.069	0.017	0.006
K	0.971	0.975	0.988	0.991

Table 3.7 Chemical composition of oxide minerals from the lower crustal xenoliths, Leucite Hills.

Sample #	157-24-1-1	157-38-13	15738-1	15738-11
SiO ₂	0.22	0	0.21	0.07
TiO ₂	53.41	45.58	46.55	6.4
Al ₂ O ₃	0.31	0	0.28	0.34
Fe ₂ O ₃	3.63	3.56	52.63	53.07
FeO	37.15	39.15	0	33.37
MnO	0.68	0.93	0.36	0.45
MgO	5.78	5.95	1.5	0.85
CaO	0	0	0.23	0.35
Total	101.18	95.17	101.76	94.9

Figure 3.12 Al_2O_3 versus TiO_2 variation diagram for mica from the lower crustal xenolith and lamproite host. The high Al_2O_3 content of mica from the lower crustal xenoliths indicates crystallization at elevated pressures in the lower crust.

Figure 3.13 FeO versus Al_2O_3 variation diagram shows the compositional distinctions between leucite from the Leucite Hills lamproites and the fine-grained secondary assemblage in the lower crustal xenoliths.

Clinopyroxene

Data for clinopyroxene from the lower crustal xenoliths, the fine-grained secondary clinopyroxene that has replaced primary pyroxene in the xenoliths and the lamproite host, are listed in Table 3.8 and plotted on the pyroxene classification diagram in Fig. 3.14.

Following the classification scheme of Morimoto et al. (1988), clinopyroxene from the lower crustal xenoliths and the host lamproite is diopside, whereas the fine-grained secondary clinopyroxene is augite. Clinopyroxene from the lower crustal xenoliths is, however, more Fe-rich relative to that of the lamproite host. The data from the clinopyroxene show a negative correlation on a plot of MgO versus FeO (Fig. 3.15a). This could be attributed to the compositional control by a fractionating magma. Compared to the clinopyroxene from the Leucite Hills lamproites and that from the secondary assemblages, clinopyroxene from the granulite xenoliths contain higher concentrations of Al_2O_3 (Fig. 3.15b) which may reflect crystallization under higher pressures. The Na_2O concentrations of clinopyroxene from the granulite xenoliths and the lamproite host are similar (0.0-0.5 wt%), but lower than that of the fine-grained, secondary clinopyroxene (1.2-2.1 wt%). The MnO concentration of clinopyroxene in the granulite xenoliths (0.1-0.4 wt%) and the fine-grained aggregates (0.3-0.6 wt%) are higher than that of the host lamproite (< 0.13 wt%) (Fig. 3.15c). Clinopyroxene from the lamproite host also has higher abundances of TiO_2 (0.7-1.8 wt%) than that of the granulite xenoliths (0.1-0.5 wt%) and fine-grained secondary aggregates (0.0-0.3 wt%). The clinopyroxene from the lower crustal xenolith shows no sign of zonation (Fig. 3.16).

The relative contents of the jadeite and Tschermak molecule have been utilized to compare different granulite types (e.g., garnet granulite versus two-pyroxene granulite) and to help in identifying granulites from igneous rocks. Pyroxenes in two-pyroxene granulites are known to be more enriched in the Ca-Tschermak molecule and less enriched in the jadeite component relative to garnet granulites (Kay and Kay, 1983). This is because the Tschermak components are rejected from the pyroxene to produce a pyrope-rich garnet in garnet granulites. It was also noted by Kushiro (1965) that increasing the pressure at constant temperatures changes the gabbroic clinopyroxene from Ca-Tschermak-rich, jadeite-poor to Ca-Tschermak-poor, jadeite-rich compositions.

Table 3.8 Chemical composition of clinopyroxene from the lower crustal xenoliths, secondary clinopyroxene from the fine-grained assemblage that replaces the clinopyroxene at the margin of the xenoliths, and the Leucite Hills lamproites.

Sample #	155	156-2b	157-4	157-5	157-6	157-7	157-8	157-9	157-22	157-23	157-28
SiO ₂	52.08	52.58	50.97	51.80	52.34	51.25	50.77	52.10	52.74	50.63	52
TiO ₂	0.26	0.19	0.21	0.33	0.09	0.46	0.46	0.31	0.34	0.35	0.38
Al ₂ O ₃	2.30	2.34	3.60	2.37	2.51	3.52	3.63	2.71	2.63	3.15	2.23
Cr ₂ O ₃	0.00	0.14	0.11	0.06	0.12	0.07	0.09	0.19	0.44	0	0.22
FeO _T	9.26	7.73	10.34	9.39	8.83	8.34	8.12	7.68	7.47	9.8	5.79
MnO	0.26	0.23	0.23	0.27	0.34	0.38	0.21	0.21	0.13	0.33	0.13
MgO	12.40	13.71	13.03	13.01	13.91	13.12	13.68	14.16	14.37	11.85	14.42
CaO	22.43	22.70	21.32	22.55	22.37	22.53	22.09	22.81	22.81	22.19	23.1
Na ₂ O	0.80	0.53	0.95	0.66	0.09	0.73	0.11	0.00	0.56	0	0.17
Total	99.79	100.15	100.76	100.44	100.60	100.40	99.16	100.17	101.49	98.3	98.44
Si	1.946	1.947	1.881	1.922	1.938	1.894	1.903	1.932	1.922	1.931	1.948
Ti	0.007	0.005	0.006	0.009	0.003	0.013	0.013	0.009	0.009	0.010	0.011
Aliv	0.054	0.053	0.119	0.078	0.062	0.106	0.097	0.068	0.078	0.069	0.052
Fe ³ (iv)	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000
Alvi	0.048	0.049	0.037	0.025	0.048	0.048	0.063	0.050	0.035	0.073	0.047
Cr	0.000	0.004	0.003	0.002	0.004	0.002	0.003	0.006	0.013	0.000	0.007
Fe ³ (vi)	0.049	0.028	0.136	0.080	0.012	0.083	0.013	0.000	0.051	0.000	0.000
Fe ²	0.240	0.211	0.184	0.211	0.261	0.175	0.242	0.238	0.177	0.313	0.181
Mn	0.008	0.007	0.007	0.008	0.011	0.012	0.007	0.007	0.004	0.011	0.004
Mg	0.691	0.757	0.717	0.720	0.768	0.723	0.765	0.783	0.781	0.674	0.806
Ca	0.898	0.901	0.843	0.897	0.888	0.892	0.887	0.906	0.891	0.907	0.927
Na	0.058	0.038	0.068	0.047	0.006	0.052	0.008	0.000	0.040	0.000	0.012
Total	4.000	4.000	4.000	4.000	4.000	4.000	4.000	3.998	4.000	3.988	3.995

Samples 156-2b-1, 157-5-16, 157-8-33, 157-36-3, 157-38 are from the fine-grained assemblage.

Table 3.8 (continued)

Sample #	Lamproite 157-4-h	Lamproite 157-7-h	Lamproite 157-23-h	Lamproite 157-38-h
SiO ₂	54.87	52.59	54.82	55.30
TiO ₂	0.75	2.11	0.73	0.69
Al ₂ O ₃	0.46	0.67	0.61	0.38
Cr ₂ O ₃	0.00	0.00	0.00	0.00
FeO	3.20	5.11	3.93	4.06
MnO	0.00	0.11	0.13	0.12
MgO	17.41	15.43	16.80	16.03
CaO	24.38	22.23	24.47	24.34
Na ₂ O	0	1.37	0	0.31
Total	101.070	99.620	101.490	101.230
Si	1.98	1.93	1.98	2.00
Ti	0.02	0.06	0.02	0.02
Aliv	0.02	0.03	0.02	0.00
Fe ³ (iv)	0.00	0.04	0.00	0.00
Alvi	0.00	0.00	0.00	0.01
Cr	0.00	0.00	0.00	0.00
Fe ³ (vi)	0.00	0.05	0.00	0.00
Fe ²	0.10	0.07	0.12	0.12
Mn	0.00	0.00	0.00	0.00
Mg	0.94	0.85	0.90	0.86
Ca	0.94	0.88	0.95	0.94
Na	0.00	0.10	0.00	0.02
Total	3.99	4.00	3.99	3.99

Figure 3.14 Ca-Mg-Fe classification diagram for clinopyroxene in lower crustal xenoliths ○, secondary clinopyroxene □, and clinopyroxene in the lamproite host ◆. Classification scheme after Morimoto et al. (1988).

Figure 3.15 (a) FeO versus MgO, (b) Al₂O₃ versus Na₂O, and (c) TiO₂ versus MnO variation diagrams for clinopyroxene from lower crustal xenoliths, secondary fine-grained assemblage, and lamproite host. Clinopyroxene from the lower crustal xenoliths contain higher Al₂O₃, FeO, MnO and lower MgO and TiO₂ than those of the lamproite host.

Figure 3.16 Microprobe traverses along margin (m) – core (c) – margin (m) of clinopyroxene from lower crustal xenoliths. Data from (a) sample 157-6, (b) sample 157-8, (c) sample 157-25, (d) sample 157-38. Data from clinopyroxene do not show significant zonation in MgO, CaO, FeO, and Al₂O₃ contents from core to the

Therefore, the jadeite-Tschermak relationships can indicate the pressures at which the xenoliths equilibrated.

Microprobe analyses are always reported as Fe_{total} , and the uncertainty inherited in the Fe^{3+} recalculation can introduce errors to the calculations of the clinopyroxene end-member contents, particularly those of acmite and jadeite. Some methods (e.g., Kushiro, 1965; White, 1964) calculate acmite before jadeite, whereas others (e.g., Cawthorn and Collerson, 1974) compute the end members in the reverse order. O'Brien (1983) and Bradely (1985) have applied the method of Cawthorn and Collerson (1974) and considered all Fe as Fe^{2+} . The problem inherent in this method is that the lack of discrimination between ferric and ferrous Fe (i.e., having all Fe as Fe^{2+}) for determining the clinopyroxene end-members results in calculating unrealistically lower acmite and higher jadeite components. To overcome the dilemma of the Fe^{3+} recalculation, Droop (1987) proposed an equation to estimate the Fe^{3+} from Fe_{total} in ferromagnesian silicates based on stoichiometry. In an Excel program written by Rukhlov (1999), the jadeite is calculated first and the acmite molecules are then calculated on the basis of available Na. The advantage to this approach is that the acmite content is not overestimated due to the presence of Fe^{3+} . More importantly, the remaining Fe^{3+} can be used to calculate the Fe-Tschermak, hedenbergite and ferrosilite molecules and these components can thus be approximated accurately. Clinopyroxene from the Leucite Hills granulites has been calculated for their end-member contents using the approach taken by Rukhlov (1999) (Table 3.8). The jadeite/Tschermak ratios for the Leucite Hills granulite xenoliths are given in Table 3.9.

The jadeite component of clinopyroxene from the Leucite Hills granulites ranges from 0 to 5.2 mole % and the Tschermak molecule ranges from 4.4 – 9.0 mole %. The jadeite/Tschermak ratio thus varies between 0.0 to 0.6. Worldwide, two-pyroxene granulites contain a lower jadeite/Tschermak ratio (0.0 - 0.7) than garnet granulites (0.6 – 2.0). The fact that the jadeite and Tschermak contents of the Leucite Hills granulite clinopyroxene are lower than samples from other localities show that the clinopyroxene has either lost their jadeite component to secondary mineral phases during the slow rise to the surface or they formed under relatively low pressure conditions.

Table 3.9 Calculated clinopyroxene components from the lower crustal xenoliths, Leucite Hills.

Sample	Jd	Aeg	CATs	FATs	TiTs	Di	Hd	En	Fs	Jd/Ts
155	2.86	2.94	1.92	1.98	0.73	62.67	22.54	3.22	1.16	0.62
156-2b	2.49	1.33	2.40	1.50	0.52	66.60	19.22	4.62	1.33	0.56
157-4	1.56	5.25	2.15	8.33	0.58	57.94	15.42	6.93	1.84	0.14
157-6	0.52	0.12	4.26	1.09	0.25	61.53	21.80	7.70	2.73	0.09
157-7	1.97	3.27	2.81	5.01	1.28	63.74	16.49	4.32	1.12	0.22
157-8	0.67	0.13	5.68	1.17	1.30	60.92	19.78	7.82	2.54	0.08
157-9	0.00	0.00	5.02	0.00	0.87	64.82	20.27	6.88	2.15	0.00
157-23	0.00	0.00	7.36	0.00	1.01	56.03	26.88	5.90	2.83	0.00
157-25	1.92	2.06	1.61	3.08	0.94	68.24	15.78	5.17	1.20	0.34
157-28	1.24	0.00	3.45	0.00	1.08	72.14	16.62	4.44	1.02	0.27
157-30	0.00	0.00	5.88	0.00	0.40	57.58	23.19	9.24	3.72	0.00
157-31	2.97	1.38	2.50	1.23	0.70	62.29	20.85	6.05	2.03	0.67
157-36	0.95	1.68	1.98	3.88	0.77	58.78	23.72	5.87	2.37	0.14
157-38	1.65	1.87	1.72	2.61	0.92	62.43	21.11	5.75	1.95	0.31

Orthopyroxene

Most orthopyroxene in the lower crustal xenoliths from the Leucite Hills coexists with clinopyroxene, except for samples 156-2 (an amphibole-orthopyroxenite) and 157-27 (the felsic xenolith).

The chemical composition and unit formula of the orthopyroxenes are listed in Table 3.10. Figure 3.17 shows the compositional variations of orthopyroxene on a Ca-Mg-Fe diagram. The data from the fine-grained, secondary orthopyroxene which occurs at the margins of the xenoliths are also plotted on Figure 3.17.

The orthopyroxene is enstatite and ranges in composition from moderately Mg-rich (En_{68}) to Fe-rich (En_{53}) varieties. On the MgO versus FeO diagram (Fig. 3.18a) the data from orthopyroxene exhibits a strong negative correlation extending from pyroxene with low FeO, high MgO contents (the ultramafic xenoliths) to that with high FeO and low MgO contents (the felsic and mafic xenoliths).

An inverse correlation also exists between the Al_2O_3 and FeO for orthopyroxene (Fig. 3.18b). The data from orthopyroxene from the ultramafic xenolith plot at a much higher Al_2O_3 concentration than those from the rest of the xenoliths. However, the felsic and mafic xenoliths have comparable Al_2O_3 concentrations. The observed negative correlation between Al_2O_3 and FeO is apparently common in granulite facies rocks (Griffin and Heier, 1969). However, the orthopyroxene from the Leucite Hills granulite xenoliths shows lower Al_2O_3 (0.8 – 1.9 wt%, ultramafic sample excluded) and higher FeO (19 – 28 wt%) concentrations than many of the two-pyroxene granulites documented by Rudnick and Taylor (1987) and Mengel (1990) from other areas. The only exception is the orthopyroxene from the Highwood Mountains (Joswiak, 1992) that contains comparable Al_2O_3 and FeO contents to those of the Leucite Hills granulite xenoliths. Compositionally, no noticeable differences exist between the orthopyroxene from the mafic xenoliths and the secondary, fine-grained orthopyroxene occurring at the xenolith margins (Fig. 3.18a, b).

Fe-Mg distribution coefficients between orthopyroxene and clinopyroxene have been used as criteria to separate igneous rocks from their metamorphic equivalents (Kretz, 1963; Saxena, 1968; Aoki and Shiba, 1973). This approach assumes that the distribution of elements among coexisting phases is dependent on pressure, temperature,

Table 3.10 Chemical composition of orthopyroxene from the lower crustal xenoliths and the fine-grained assemblage that replaces the clinopyroxene at the margins of the xenoliths.

Sample #	155	156-2	157-4	157-5	157-6	157-8	157-9	157-12	157-22	157-23	157-24
SiO ₂	52.35	52.63	52.23	51.22	52.87	52.74	52.49	52.03	52.62	51.74	51.68
TiO ₂	0.00	0.00	0.03	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.11
Al ₂ O ₃	1.24	2.95	1.87	0.96	1.02	1.63	1.43	1.56	1.37	1.10	1.16
Cr ₂ O ₃	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
FeO _T	24.84	19.06	21.50	26.61	23.42	21.09	21.90	22.61	21.28	27.52	28.58
MnO	0.58	0.55	0.58	0.77	0.94	0.55	0.65	0.54	0.45	0.86	0.84
MgO	20.73	24.55	23.07	19.06	22.16	23.58	23.01	21.60	22.58	18.64	18.55
CaO	0.38	0.29	0.31	0.48	0.50	0.44	0.29	0.85	0.35	0.36	0.48
Na ₂ O	0.00	0.00	0.19	0.00	0.00	0.00	0.00	0.00	0.00	0.00	0.00
Total	100.12	100.03	99.78	99.10	100.91	100.03	99.77	99.19	98.65	100.22	101.40
Si	1.968	1.924	1.934	1.966	1.958	1.946	1.950	1.957	1.975	1.972	1.952
Ti	0.000	0.000	0.001	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.003
Aliv	0.032	0.076	0.066	0.034	0.042	0.054	0.050	0.043	0.025	0.028	0.048
Fe ³ (iv)	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000
Alvi	0.023	0.051	0.016	0.010	0.002	0.017	0.013	0.026	0.035	0.021	0.003
Cr	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000
Fe ³ (vi)	0.009	0.025	0.063	0.024	0.040	0.037	0.037	0.017	0.000	0.007	0.039
Fe ²	0.772	0.557	0.603	0.831	0.685	0.614	0.644	0.694	0.668	0.870	0.864
Mn	0.018	0.017	0.018	0.025	0.029	0.017	0.020	0.017	0.014	0.028	0.027
Mg	1.162	1.338	1.274	1.091	1.223	1.297	1.275	1.211	1.263	1.059	1.044
Ca	0.015	0.011	0.012	0.020	0.020	0.017	0.012	0.034	0.014	0.015	0.019
Na	0.000	0.000	0.014	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000
Total	4	4	4	4	4	4	4	4	3.995	4	4

Table 3.10 (continued)

Sample #	157-27	157-28	157-36	157-38	156-2b-3-1	157-5-15	157-36-8
SiO ₂	52.04	54.44	51.03	51.91	52.99	51.58	52.13
TiO ₂	0.00	0.13	0.00	0.04	0.00	0.00	0.04
Al ₂ O ₃	0.77	1.19	0.94	1.24	0.87	1.09	1.02
Cr ₂ O ₃	0.00	0.12	0.00	0.00	0.00	0.00	0.02
FeO _T	23.72	22.22	28.87	25.94	23.38	25.28	28.77
MnO	0.88	0.42	0.84	0.62	0.69	0.83	0.79
MgO	20.77	24.49	18.18	20.53	21.61	18.84	17.61
CaO	0.84	0.32	0.36	0.48	0.31	0.33	0.37
Na ₂ O	0.00	0.13	0.00	0.00	0.00	0.00	0.00
Total	99.02	103.46	100.21	100.76	99.85	97.95	100.75
Si	1.975	1.943	1.954	1.945	1.985	1.992	1.985
Ti	0.000	0.003	0.000	0.001	0.000	0.000	0.001
Aliv	0.025	0.050	0.042	0.055	0.015	0.008	0.015
Fe ³ (iv)	0.000	0.007	0.004	0.000	0.000	0.000	0.000
Alvi	0.009	0.000	0.000	0.000	0.023	0.041	0.031
Cr	0.000	0.003	0.000	0.000	0.000	0.000	0.001
Fe ³ (vi)	0.016	0.056	0.046	0.053	0.000	0.000	0.000
Fe ²	0.737	0.600	0.874	0.760	0.732	0.817	0.917
Mn	0.028	0.013	0.027	0.020	0.022	0.027	0.025
Mg	1.175	1.303	1.038	1.147	1.207	1.085	1.000
Ca	0.034	0.012	0.015	0.019	0.012	0.014	0.015
Na	0.000	0.009	0.000	0.000	0.000	0.000	0.000
Total	4	4	4	4	3.996	3.983	3.990

Samples 156-2b-3-1, 157-5-15, and 157-36-8 are from the fine-grained assemblage.

Fig. 3.17 Classification of orthopyroxene from the lower crustal xenoliths. The secondary orthopyroxene (squares) lies closer to the ferrosilite field. Symbols as in Fig. 3.14.

Figure 3.18 (a) FeO versus MgO and (b) FeO versus Al_2O_3 variation diagrams for orthopyroxene from the lower crustal xenoliths. Ultramafic xenolith $\blacktriangle$, mafic xenolith $\circ$, felsic xenolith $\bullet$, and orthopyroxene in the fine-grained assemblage $\square$.

and the chemical composition of the whole rock. The distribution coefficients with reference to the Mg-Fe exchange for orthopyroxene- clinopyroxene pairs can be used to estimate the conditions of equilibration and is calculated as:

$$K_D^{(Mg-Fe)} = (X_{Mg}^{Opx}/1 - X_{Mg}^{Opx}) * (1 - X_{Mg}^{Cpx}/X_{Mg}^{Cp})$$

where $X_{Mg}^{Opx} = Mg/(Mg+Fe^{2+})$ in orthopyroxene and $X_{Mg}^{Cpx} = Mg/(Mg+Fe^{2+})$ in clinopyroxene. Based on numerous analysis of known mafic igneous and metamorphic outcrops, the distribution coefficient ($K_D^{(Mg-Fe)}$) for the igneous rocks and granulites have been reported to be 0.73 and 0.59, respectively (Saxena, 1968). The calculated $K_D^{(Mg-Fe)}$ for most of the Leucite Hills lower crustal mafic xenoliths is less than 0.59 with the exception of three samples having $K_D^{(Mg-Fe)} = 0.60-0.67$ (Table 3.11). This could indicate that even though the lower crustal xenoliths originally crystallized from a mafic magma, they have also experienced granulite metamorphic recrystallization at some point in their history.

Table 3.11 Calculated Kd values for pyroxenes from the lower crustal xenoliths, Leucite Hills.

Sample	X_{Mg}^{Cpx}	X_{Mg}^{Opx}	Kd
155	0.74	0.6	0.52
157-3	0.76	0.56	0.4
157-4	0.8	0.68	0.54
157-6	0.77	0.57	0.39
157-7	0.75	0.64	0.61
157-8	0.76	0.68	0.67
157-9	0.77	0.66	0.6
157-23	0.68	0.55	0.56
157-25	0.82	0.68	0.49

3.6. Whole Rock Geochemistry

The whole rock chemistry of the xenoliths can yield information about the rock types and the process by which they formed. A number of lower crustal xenoliths from the Leucite Hills were analyzed for their major and trace element contents by Activation Laboratories Ltd using ICP and ICP-MS. A full description of the sample preparation and analytical procedures are presented in Appendices A and B. The analyzed samples are considered representative of xenoliths from Hatcher Mesa. It should be noted that samples that were geochemically analyzed for their major and trace element and isotopic ratios were carefully examined to avoid any veins, attached host lamproite, and fine-grained secondary alteration at the margins of the xenoliths.

3.6.1 Major elements

Table 3.12 gives the major element abundances, the Mg# and the CIPW norms of the whole rocks of the lower crustal xenoliths. The formulas for calculating the Fe_2O_3 values are adopted from Middlemost (1989). The LOI was not determined for some of the samples because of difficulty in obtaining sufficient material.

The lower crustal xenoliths are basic to ultrabasic in composition (SiO_2 44 – 54 %), except for sample 157-27 which is a felsic xenolith ($\text{SiO}_2 = 70\%$). The $\text{Na}_2\text{O} + \text{K}_2\text{O}$ versus SiO_2 diagram (Fig. 3.19) is used to classify the xenoliths according to the classification scheme of Le Maitre et al. (1989). Sample 156-2, which has the lowest SiO_2 content, is depleted in its total alkali content and is classified on this diagram as a microbasalt. The ultrabasic composition of this sample is also reflected in its high abundance of mafic minerals (i.e., high modal abundance of orthopyroxene and amphibole). At the other extreme, sample 157-27 has the highest silica and alkali content and the data from it plot in the rhyolite field. This sample is characterized by modal alkali feldspar and quartz. The remainder of the lower crustal xenoliths are classified as either basalts or basaltic andesites. The abundance of mafic xenoliths suggests that the lower crust underlying the Leucite Hills lamproites is dominantly basic in composition.

The lower crustal xenoliths are olivine and nepheline normative, except for sample 157-29 (anorthosite), 157-27 (felsic), and 157-38 (leucogabbro). The felsic and

Table 3.12 Whole rock major element and CIPW normative composition of the lower crustal xenoliths from the Leucite Hills.

Sample #	156-2	157-4	157-6	157-7	157-27	157-29	157-38
SiO ₂	44.24	46.84	45.06	50.54	70.27	49.55	54.10
TiO ₂	0.545	0.817	0.835	0.703	0.06	0.162	0.940
Al ₂ O ₃	8.96	14.57	15.37	14.97	14.24	28.25	17.88
MgO	17.98	9.36	8.69	6.62	0.13	0.63	4.73
CaO	7.24	10.79	11.16	14.71	2.93	13.21	9.21
MnO	0.257	0.166	0.17	0.193	0.004	0.02	0.13
Fe ₂ O ₃ T	17.32	11.43	12.47	6.94	0.65	1.81	7.68
Na ₂ O	1.25	2.42	2.46	2.57	2.04	3.52	4.42
K ₂ O	1.46	1.03	2.06	1.41	6.8	1.04	0.89
P ₂ O ₅	0.13	0.52	0.26	0.38	0.65	0.22	0.45
LOI	n.d	n.d	0.22	n.d	n.d	0.5	0.23
Total	99.38	97.95	98.74	99.05	97.78	97.61	100.65
Mg#	0.65	0.59	0.55	0.63	0.26	0.38	0.52

Mg# = (Mg/Mg + 0.899Fe⁺²)

	Qtz	0	0	0	0	28.04	0	0
An	14.54	26.28	25	25.28	9.78	59.6	26.04	0
Di	16.74	20.27	23.95	37.2	0.74	4.65	13.46	0
Ol	45.77	22.38	20.74	3	0	0	4.69	0
Ne	1.09	1.16	9.33	8.35	0	3.04	0	0
Or	8.59	6.12	12.23	8.35	40.93	6.23	5.18	0
Ab	8.55	18.66	3.76	16.37	17.67	24.83	37.03	0
Hy	0	0	0	0	0.84	0	8.9	0
Mt	3.34	2.28	2.5	1.5	0.19	0.1	1.6	0
Ilm	1.05	1.6	1.62	1.35	0.11	0.32	1.77	0
Ap	0.28	1.12	0.56	0.82	1.4	0.47	0.97	0

Figure 3.29 Classification of meta-igneous granulite xenoliths based on alkaline and silicic contents (Le Maitre et al., 1989).

◆ 156-2, ■ 157-4, ● 157-6, + 157-7, ▲ 157-27, ● 157-29, × 157-38

leucogabbro xenoliths (157-38) are hypersthene normative. Using the classification scheme of Yoder and Tilley (1962), the protolith of the leucogabbro granulite is an olivine tholeiite, whereas protoliths for the remainder are alkali-olivine basalts. The Mg# of the mafic samples lie between 0.65 and 0.38, lower than that observed in primary, mantle-derived melts.

Because of the broad range in MgO content, this oxide was used to illustrate the variability of major element oxides in the Leucite Hills lower crustal xenoliths (Fig. 3.20). Many oxides show strong to moderate correlation with MgO. In the MgO versus SiO₂ diagram, the highest SiO₂ and lowest MgO contents belong to the felsic xenolith while the amphibole orthopyroxenite xenolith contains the minimum and maximum amounts of MgO and SiO₂, respectively. A strong correlation also exists on the MgO versus FeO diagram. The amphibole-bearing samples (156-2, 157-4, 157-6) show the highest abundances of FeO and MgO of any of the lower crustal xenoliths. The K₂O content of the xenoliths is 2.0 wt% except for the felsic xenoliths. The mafic xenoliths have higher abundances of Na₂O and CaO than the felsic xenolith. The anorthosite and the orthopyroxenite xenoliths have higher and lower Al₂O₃ contents, respectively, than the other lower crustal xenoliths.

The variations in the major element compositions of the mafic to ultramafic lower crustal xenoliths suggest that their protoliths crystallized from a fractionating magma.

3.6.2 Trace elements

The trace element concentrations of the lower crustal xenoliths are given in Table 3.13. There is a great deal of variation in trace element compositions reflecting the considerable range of mineralogy of the lower crustal xenoliths.

The Ni and Cr contents of the xenoliths vary from 20 to 380 ppm and 120 to 1400 ppm, respectively. Excluding sample 157-6 (amphibole orthopyroxenite) whose composition is closest to that expected for a primary mafic liquid, the Ni and Cr contents, as well as the Mg#, of the granulite xenoliths are generally lower than those expected for primary mafic magmas (i.e., Ni > 400 ppm, Cr > 1000 ppm, Mg# > 0.7; Wilson 1989). This suggests that the granulites are probably the metamorphosed equivalents of a fractionated, mantle-derived magma. Furthermore, the strong positive correlation exists

Figure 3.20 Whole rock major element plots of lower crustal granulite xenoliths from the Leucite Hills. Correlation between the major element contents of the lower crustal sample in most of the diagrams shows that the samples are probably related by magmatic fractionation.

between the Ni, Cr, and Co contents of the xenoliths (Fig. 3.21) consistent with fractional crystallization.

The REE concentrations of the lower crustal xenoliths from the Leucite Hills lamproites have been normalized to chondrites and plotted in Figure 3.22. A marked variation in the REE profiles is apparent among the lower crustal xenoliths that seems to correlate, in large part, with the variations in their modal mineralogy, normative composition and major element abundances. The anorthosite and amphibole orthopyroxenite have, respectively, the lowest and highest total REE content of any of the other lower crustal xenoliths from the Leucite Hills. The lower crustal xenoliths also show varying degrees of LREE enrichment over HREE ranging from $(La/Yb)_N = 265$ in the felsic xenolith (sample 157-27) to $(La/Yb)_N = 2$ in the amphibole - bearing mafic xenolith (sample 157-6). Granulites with fresh amphibole (156-2, 157-4) display a higher LREE enrichment than those with altered amphibole (157-6). The REE patterns of the two anhydrous gabbroic granulites (157-7, 157-38) overlap and both have HREE contents similar to the amphibole-bearing xenoliths.

The strong LREE enrichment and HREE depletion in the felsic xenolith are unlike any of the patterns observed in the lower crustal xenolith collected from Hatcher Mesa and hence require some discussion. The felsic xenolith has HREE contents lower than any of the studied xenoliths. In addition, the LREE enrichment pattern in the felsic xenolith is steeper than that of the other xenoliths. REE profiles similar to those of the felsic xenolith have been reported elsewhere from other granulite xenoliths (Rogers and Hawkesworth, 1982; Leeman et al., 1985; Joswiak, 1992) and granulite terrains (Weaver and Tarney, 1981). Tindle and Pearce (1983) calculated the REE abundances for partial melts of a quartz diorite source and showed that these melts have REE patterns similar to those of felsic xenoliths. They therefore concluded that felsic xenoliths may be the result of partial melting of pre-existing mafic to intermediate crustal rocks. The lower crustal felsic xenolith from Leucite Hills could not have been produced by magmatic fractionation of the melt that generated mafic to ultramafic xenoliths. This is reinforced by the fact that the total REE abundance of the felsic xenolith is low compared to the mafic and ultramafic xenoliths. In addition, the strongly positive Eu anomaly in the felsic

Table 3.13 Whole rock trace element composition (in ppm) of lower crustal xenoliths from the Leucite Hills.

Sample #	156-2	157-4	157-6	157-7	157-27	157-29	157-38
V	202	220	251	152	10	38	175
Cr	1400	446	377	206	95	118	287
Co	74	53	40	26	*	2	26
Ni	378	220	123	85	*	20	76
Cu	-10	231	31	60	17	42	50
Zn	127	94	*	98	*	*	77
Ga	14	16	14	15	15	22	19
Ge	1.2	1.6	0.8	1.4	0.5	0.7	1
Mo	17	*	2	17	16	16	*
Cs	*	*	0.1	0.1	*	*	*
Ba	3290	3710	1320	1210	3870	1120	1090
Rb	21	8	16	22	33	12	3
Th	1.03	1.10	1.34	1.66	0.47	1.40	0.97
U	1.04	1.63	0.43	1.26	0.51	0.70	1.31
Ta	0.2	0.2	0.2	0.2	0.0	0.3	0.6
Nb	5.5	5.3	2.8	4.7	0.9	3.8	4.7
Sr	587	926	346	1100	1680	585	531
Zr	58	112	54	57	76	38	135
Hf	1.9	2.7	1.4	1.5	1.8	0.8	3.4
Y	41.1	20.2	19.0	18.4	0.6	7.5	15.1
W	1.2	0.4	0.7	1.5	1.0	2.8	*
Tl	0.05	0.05	0.02	0.11	0.00	0.08	*
Pb	33	52	*	41	91	46	13
Bi	*	0.09	*	*	0.26	*	*
Pb	33	52	*	41	91	46	13
La	80.8	27.2	6.47	16.2	26.5	5.87	14.5
Ce	247	61.6	14.2	32.5	30.8	11.5	34
Pr	36.9	8.02	1.69	3.60	2.40	1.27	3.91
Nd	139	30.6	7.80	13.6	6.19	4.99	15.7
Sm	20.2	5.29	2.31	3.05	0.52	1.19	3.33
Eu	4.35	1.45	0.641	1.14	1.69	0.614	1.2
Gd	14.0	4.86	2.90	3.26	0.33	1.25	3.07
Tb	1.42	0.65	0.54	0.56	0.03	0.22	0.51
Dy	6.82	3.61	3.37	3.33	0.11	1.29	2.8
Ho	1.21	0.71	0.73	0.70	0.02	0.26	0.6
Er	3.32	2.04	2.15	1.92	0.06	0.74	1.65
Tm	0.449	0.290	0.321	0.271	0.012	0.111	0.206
Yb	3.06	1.99	2.17	1.82	0.07	0.75	1.38
Lu	0.437	0.292	0.316	0.255	0.011	0.103	0.208

* Below detection level

Figure 3.21 Variation of Ni, Cr and Co in lower crustal granulite xenoliths from the Leucite Hills. A positive correlation between the data indicates that crystal fractionation was probably an important mechanism in the petrogenesis of the lower crustal rocks. Symbols as in Fig.3.20.

Figure 3.22 Chondrite-normalized REE patterns of the lower crustal xenoliths, Leucite Hills. The amphibole-bearing orthopyroxenite (156-2) contains the highest concentration of REE. The felsic xenolith (157-27) shows a strong positive Eu anomaly and an extreme depletion in HREE.

xenoliths could not have been generated by fractional crystallization from the mafic magma. It seems more likely that the felsic xenolith is the result of partial melting of the mafic granulites.

The amphibole-bearing, ultramafic xenolith (156-2) has a significantly elevated REE pattern and is also enriched in LREE. Sample 157-4, a mafic xenolith which also contains fresh amphibole, follows a REE pattern similar to sample 156-2 but has a lower total REE. Amphibole-bearing xenoliths from central Montana show LREE enrichment over HREE similar to that found in the Leucite Hills xenoliths. Joswiak (1992) suggests that the amphibole-bearing granulite xenoliths from central Montana crystallized directly from a hydrous alkaline mafic melt.

A slightly negative Eu anomaly ($\text{Eu}/\text{Eu}^* = 0.7$) is seen in the data from the amphibole-bearing mafic xenoliths (samples 157-4, 157-6). By contrast, the anorthosite xenolith displays a moderately positive Eu anomaly ($\text{Eu}/\text{Eu}^* = 1.7$) whereas the felsic xenolith displays a strongly positive Eu anomaly ($\text{Eu}/\text{Eu}^* = 15$). A negative Eu anomaly indicates either the crystallization and removal of plagioclase from a melt or the presence of a plagioclase-bearing mantle in which plagioclase is a residual phase after melting of the source. The presence of amphibole also could have been a contributing factor because the negative anomaly only occurs in the amphibole-bearing samples. A positive Eu anomaly may reflect either melting from a restite which contains plagioclase or that plagioclase has accumulated during magmatic fractionation. The accumulation of plagioclase to produce the positive Eu anomaly in gabbroic (157-7, 157-38) and anorthositic (157-29) xenoliths is supported by petrographic studies. The strikingly strong positive Eu anomaly in the felsic xenolith cannot, however, be attributed solely to the plagioclase accumulation. The petrography, SEM studies and geochemical analyses (low CaO and high K_2O abundances) confirm that the feldspars in the felsic xenolith are dominantly K-feldspar and not plagioclase. It would seem more reasonable to assume that the protolith of the felsic xenolith originates from partial melting of a plagioclase-bearing source.

Trace element data from the lower crustal xenoliths from Leucite Hills have been normalized to primitive mantle values and plotted on a conventional spider diagram (Fig. 3.23) in order of decreasing chemical incompatibility from left to right. The spider

diagram patterns for the xenoliths have similar configurations in terms of enrichment in many incompatible trace elements, negative Cs anomaly, a strong positive Sr anomaly and a pronounced Nb-Ta trough.

Even though the degrees of incompatibility of Ba, Rb and Cs are very similar, the granulite xenoliths show different degrees of enrichment for these elements. For example, the leucogabbro and the felsic xenolith display the lowest and the highest levels of Ba and Rb enrichment, respectively (Fig. 3.23). Ba and Rb substitute for K and therefore their abundances should be controlled by the presence or absence of mineral phases such as K-feldspar, hornblende and mica. The leucogabbro contains none of these minerals and therefore shows the lowest levels of Rb and Ba enrichment. In contrast, the felsic xenolith is characterized by a high modal abundance of K-feldspar and thus contains the highest concentrations of Rb and Ba.

The concentrations of U and Th in the lower crustal xenoliths, range from 0.5 to 1 ppm, and fall well below those of the average upper crust (2.8 and 10.8 ppm, respectively, Taylor and McLennan, 1985). It is likely that the conditions that existed at the lower crustal levels have depleted the granulites in both U and Th. The fractionation of U and Th is also reflected in the Pb isotopic composition of the xenoliths (see next section).

The Nb and Ta content of the granulite xenoliths range from 1 to 5.5 ppm and 0.02 to 0.6 ppm, respectively. Nb and Ta show significant negative anomalies relative to the neighboring elements in the hygromagmatophile element diagram (Fig. 3.23). The depletion in Nb and Ta has been usually attributed to either subduction process or the presence of a residual titanate phase in the source region (Ryerson and Watson, 1987). The significance of Nb and Ta depletion in evaluating the presence of arc-like geochemical feature in the lower crustal xenoliths is discussed in section 3.8.2.

On the trace element spider diagrams, the data from the lower crustal xenoliths (except that from the amphibole orthopyroxenite) show a positive Sr anomaly. Sr usually substitutes for Ca in plagioclase and clinopyroxene and for K in K-feldspar. The amphibole orthopyroxenite, however, lacks these minerals and thus exhibits a negative Sr anomaly.

Figure 3.23 Primitive mantle-normalized trace element patterns for the lower crustal granulite xenoliths, Leucite Hills.

The Zr and Hf concentrations in the xenoliths range from 40 to 130 ppm and 1 to 3 ppm, respectively. On the extended incompatible element diagram (Fig. 3.23), Zr and Hf follow a similar trend relative to the adjacent elements with similar incompatibility. The only exception to such behavior is that of the amphibole orthopyroxenite xenolith (156-2) which displays a trough in Zr and Hf. This is consistent with the fact that Zr and Hf usually substitute for Ti and that the amphibole orthopyroxenite xenolith contains a lower TiO_2 abundance than that of the mafic xenoliths.

3.6.3 Isotope Geochemistry

The Rb-Sr, Sm-Nd and Pb-Pb isotope data may help determine the age of the xenoliths and evaluate the isotope geochemistry of their source. The age of the lower crustal components can also help unravel the evolution of the Wyoming Craton through which the Leucite Hills lamproites have intruded. Because crustal rocks ultimately originate from melting of the underlying mantle, determining the age of the lower crustal xenoliths might help constrain the time of upper mantle depletion. The analytical procedures including the sample preparations, column exchange and mass spectrometry operation are given in appendices A and B.

3.6.3.1 Rb –Sr isotopes

The Rb-Sr isotope data obtained for the lower crustal xenoliths are given in Table 3.14 and are plotted in Figure 3.24. The Rb and Sr concentrations have been determined either by ICP-MS or by isotope dilution. The measured $^{87}\text{Sr}/^{86}\text{Sr}$ ratios in the granulites vary from 0.70269 to 0.70581. The $^{87}\text{Sr}/^{86}\text{Sr}$ ratios show no correlation with the observed $^{87}\text{Rb}/^{86}\text{Sr}$ ratios and hence, it was not possible to obtain an isochron for this suite of samples. This finding indicates that the lower crust experienced an open-system behavior and/or redistribution of either Rb or Sr or both during the granulite-facies metamorphism. Even though such disturbances prevent any meaningful isochron and/or model age to be drawn from the Rb-Sr isotopic system, the measured $^{87}\text{Sr}/^{86}\text{Sr}$ ratios can still be used for comparative purposes. In general, the amphibole-bearing mafic xenoliths and the felsic xenolith have higher $^{87}\text{Sr}/^{86}\text{Sr}$ ratios than the rest of the granulites.

Table 3.14 Rb and Sr isotopic composition and concentration data for the lower crustal xenoliths from the Leucite Hills lamproites.

Sample #	Rb	Sr	$^{87}\text{Rb}/^{86}\text{Sr}$	$^{87}\text{Sr}/^{86}\text{Sr}$
155*	229	3293	0.20110	0.702693
156-2*	21	587	0.10408	0.705813
157-3*	16.9	611	0.08001	0.706055
157-4	8.29	989	0.02424	0.704373
157-5	73	1508	0.14000	0.703620
157-6	17.1	350	0.14133	0.705757
157-8	9	794	0.03276	0.704236
157-27*	33	1,680	0.05766	0.705168
157-29*	12	585	0.05842	0.703868
157-36	5.77	1145	0.01458	0.704666
157-38*	3	531	0.01483	0.703091

Sr isotopic ratios have been normalized to $^{86}\text{Sr}/^{88}\text{Sr} = 8.3721$.

NBS 987 Sr standard = 0.71025 ± 0.00002 . Uncertainty is given at the 2σ level.

Rb and Sr concentrations were measured by isotope dilution, except those samples marked with * that were analysed using ICP-MS at ACTLABS laboratory.

Uncertainties of Rb and Sr abundances, determined by isotope dilution, are $\pm 0.5\%$ (2σ level).

Figure 3.24 $^{87}\text{Sr}/^{86}\text{Sr}$ versus $^{87}\text{Rb}/^{86}\text{Sr}$ isotopic diagram for the lower crustal xenoliths from the Leucite Hills. Only the isotope dilution data have been plotted. No isochron age can be obtained due to the possible redistribution of Rb during the high-grade metamorphism.

3.6.3.2 Sm-Nd isotopes

The Sm-Nd isotope data obtained for the lower crustal xenoliths are presented in Table 3.15. The Sm and Nd abundances have been determined by either isotope dilution or by ICP-MS. The present-day $^{143}\text{Nd}/^{144}\text{Nd}$ display values from 0.51023 to 0.51261. $^{147}\text{Sm}/^{144}\text{Nd}$ ratios range between 0.067 to 0.198 with an average of 0.136.

Using the ISOPLOT program (Ludwig, 1988), an age of 2839 ± 300 Ma is calculated for the data from the lower crustal xenoliths. Considering the relatively high M.S.W.D.* (15), the distribution of data on the Sm-Nd isotopic diagram (Fig. 3.25) should be considered a pseudo-isochron. However, the 2.8 Ga “age” determined by the Sm-Nd pseudo-isochron age is close to the date obtained by the Pb-Pb secondary isochron method (see next section). In addition, the Sm-Nd pseudo-isochron age for the lower crustal xenoliths from the Leucite Hills is similar to a Sm-Nd isochron age (2.76 ± 0.22 Ga) obtained from lower crustal xenoliths from the Snake River Plain volcanic field (Leeman et al., 1985).

Because of relative geochemical coherence of Sm and Nd, the Sm/Nd ratios do not seem to have been affected significantly by high-grade metamorphism (DePaolo et al., 1982; Whitehouse, 1990; Barovich and Patchett, 1992). Therefore, the Sm-Nd pseudo-isochron age from the Leucite Hills lower crustal xenoliths can be used to reflect the time of the widespread crustal extraction from the mantle source underlying the Wyoming basement. The Sm – Nd pseudo-isochron for the lower crustal xenoliths define an initial $^{143}\text{Nd}/^{144}\text{Nd}$ ratio of 0.509319, corresponding to an ϵNd value of -0.1 using a present day $(^{143}\text{Nd}/^{144}\text{Nd})_{\text{CHUR}} = 0.512636$ and $(^{147}\text{Sm}/^{144}\text{Sm})_{\text{CHUR}} = 0.1967$. The ϵNd value close to zero may indicate that at 2.8 Ga the basic magma that formed the lower crustal rocks in Wyoming probably originated from a CHUR-like, undifferentiated mantle source. However, the trace element patterns of the granulites show enrichment in incompatible elements, depletion in Nb and Ta and positive Pb anomaly that point to the

*M.S.W.D. = Mean Square of Weighted Deviates is a measure of the ratio of the observed scatter of the points (from the best-fit line) to the expected scatter (from the assigned errors and error correlation).

Table 3.15 Sm and Nd isotopic composition and concentration data for the lower crustal xenoliths from the Leucite Hills lamproites.

Sample #	Nd	Sm	$^{147}\text{Sm}/^{144}\text{Nd}$	$^{143}\text{Nd}/^{144}\text{Nd}$	T_{CHUR} (Ga)
155*	22.39	4.35	0.1174	0.511359	
156-2*	139.76	20.27	0.0877	0.510823	
157-4	30.77	5.40	0.1060	0.510941	2.8
157-5	27.36	5.75	0.1272	0.511299	2.9
157-6	7.78	2.38	0.1852	0.512478	2.1
157-7	13.18	2.92	0.1340	0.511461	2.8
157-27*	5.71	0.48	0.0509	0.510230	
157-29*	4.09	0.98	0.1444	0.511779	
157-36	8.89	2.92	0.1984	0.512608	2.5
157-38*	15.70	3.33	0.1283	0.511554	

Nd isotopic ratios have been normalized to $^{143}\text{Nd}/^{144}\text{Nd} = 0.7219$.

La Jolla $^{143}\text{Nd}/^{144}\text{Nd} = 0.511870 \pm 0.00003$. Uncertainty is given at the 2σ level.

Sm and Nd concentrations were measured by isotope dilution method, except those samples marked with * that were analysed using ICP-MS.

Uncertainties of Sm and Nd abundances, determined by isotope dilution, is $\pm 0.5\%$ (2σ level).

Figure 3.25 $^{143}\text{Nd}/^{144}\text{Nd}$ versus $^{147}\text{Sm}/^{144}\text{Nd}$ diagram for the lower crustal xenoliths from the Leucite Hills. Only the isotope dilution data have been plotted.

presence of crustal component in the mantle source of the granulites. This observation may indicate that an ϵNd close to 0 was obtained as a result of mixing between depleted MORB-like mantle reservoir and metasomatic fluids emanating from the subducted oceanic pelagic sediments or continental detritus.

The linear array on the $^{143}\text{Nd}/^{144}\text{Nd}$ versus $^{147}\text{Sm}/^{144}\text{Nd}$ diagram cannot be the result of the contamination of the lower crustal xenoliths by the Leucite Hills lamproite host. It can be seen from Fig. 3.26 that the data from the Leucite Hills lamproites plot well away from those of the lower crustal xenoliths and no correlation is observed between the two. An important conclusion that can be drawn from this diagram is that the mantle source of the Leucite Hills lamproites has a present-day $^{143}\text{Nd}/^{144}\text{Nd}$ ratio that is different to that of the lower crustal xenoliths.

The “age” determined by Sm-Nd pseudo-isochron is based on a limited number of data points. In order to obtain independent information on the mantle-extraction ages, it seems reasonable to calculate the model Nd ages for the lower crustal samples. The calculated Nd model age for the mafic granulite xenoliths range from 2.1 to 2.9 Ga for T_{CHUR} (Table 3.15). Because the Sm and Nd concentrations were determined by isotope dilution, the variation in the calculated Nd model ages cannot be attributed to analytical uncertainty. In addition, the Nd model ages are in reasonable agreement with the Sm-Nd and Pb-Pb whole rock “ages” indicating that the Wyoming sub-continental mantle experienced an episode of melt removal during the Archean. Andesitic amphibolites from the Beartooth Mountains of Montana that originated from partial melting of Wyoming Archean mantle provide Nd chondritic model ages of 2.88 – 2.95 Ga. Based on such observation, Mueller et al. (1983) propose that the parental magma to the andesitic amphibolites was generated in a metasomatised mantle in late Archean.

3.6.3.3 Pb isotopes

The Pb isotopic ratios of the lower crustal xenoliths are listed in Table 3.16 and plotted in Figures 3.27 and 3.28. The Pb isotopic ratios in the granulite xenoliths encompass a relatively large range. $^{206}\text{Pb}/^{204}\text{Pb}$ ratios range from 14.77 to 17.06, $^{207}\text{Pb}/^{204}\text{Pb}$ from 15.05 to 15.42 and $^{208}\text{Pb}/^{204}\text{Pb}$ from 24.95 to 37.02. On the $^{207}\text{Pb}/^{204}\text{Pb}$ versus $^{206}\text{Pb}/^{204}\text{Pb}$ diagram (Fig. 3.27), the granulite xenoliths plot to the left of geochron

Figure 3.26 The $^{143}\text{Nd}/^{144}\text{Nd}$ versus $^{147}\text{Sm}/^{144}\text{Nd}$ isotopic diagram. Data from the Leucite Hills lamproites (LH lamp) plot well away from those of the lower crustal xenoliths (LC xlth) indicating different sources.

Table 3.16 Pb isotopic data for the lower crustal xenoliths from the Leucite Hills lamproites.

Sample #	$^{206}\text{Pb}/^{204}\text{Pb}$	$^{207}\text{Pb}/^{204}\text{Pb}$	$^{208}\text{Pb}/^{204}\text{Pb}$
156-2	16.57	15.37	36.33
157-3	15.07	15.14	35.11
157-4	16.48	15.35	36.58
157-5	17.06	15.42	37.02
157-6	14.77	15.05	34.98
157-7	16.53	15.35	36.56
157-8	16.08	15.31	36.25
157-27	16.67	15.36	36.66
157-29	16.35	15.35	36.23
157-36	15.97	15.26	35.85

An average fractionation correction of 0.1% per mass unit was applied to all measured Pb isotopic ratios based on analysis of NBS 981. The reproducibility of Pb isotopic ratios is 0.1% at the 2σ level.

Figure 3.27 Data from the lower crustal xenoliths from the Leucite Hills lamproites plot to the left of the geochron on the $^{207}\text{Pb}/^{204}\text{Pb}$ versus $^{206}\text{Pb}/^{204}\text{Pb}$ indicating an old U/Pb fractionation event. Numbers indicate ages on the Stacey - Kramers growth line.

indicating a U/Pb differentiation event that occurred during granulite metamorphism in the lower crust. Whatever the origin of the U depletion, it has a minimal effect on the daughter isotopes and thus the Pb-Pb isochron can be safely used to date the xenoliths.

Pb-Pb isochron ages have already been shown to be effective in dating granulites, particularly when combined with other methods of dating (Hamilton et al., 1979; Leeman et al., 1985). In addition, the immunity of Pb isotopic compositions to the effect of rehomogenization can offer ages older than the time of the high-grade metamorphism (Jacobson, 1978; Rudnick, 1992). Therefore, dating the granulites by the Pb-Pb isotopic method may be related to the time of the crystallization of the original melt that formed the granulites' igneous protoliths.

The Pb isotopic data from the lower crustal xenoliths form a secondary isochron on the $^{207}\text{Pb}/^{204}\text{Pb}$ versus $^{206}\text{Pb}/^{204}\text{Pb}$ diagram (Figs.3.28). Using the ISOPLOT program (Ludwig, 1988), the Pb-Pb isochron yields an age of 2477 ± 130 Ma (M.S.W.D. = 9.4) for the xenoliths. The linear array observed on the Pb-Pb isotopic diagram, however, is not a mixing line since the age determined by the Pb - Pb isochron method is similar to the Sm-Nd pseudo-isochron "age" of about 2.8 Ga and to the Nd model age that range from 2.1 to 2.9 Ga. In addition, the linear array cannot be the result of mixing Pb from the xenoliths with either the laboratory blank or the lamproite host. The laboratory blank would be most significant for those samples with the lowest Pb concentrations (i.e., sample 156-2). For the Pb procedure, normally 100 mg of the sample is digested which provides 5 μg Pb for sample 156-2. In order to significantly change the Pb isotopic ratios of this sample, the laboratory blank has to have been around 0.05 μg . However, a number of tests showed blanks of < 1 ng.

The mixing between radiogenic Pb from the lower crustal rocks and radiogenic Pb from the lamproite host can also be evaluated. First, there is no petrographic evidence to support a pervasive alteration in the xenoliths by fluids or melts from the lamproite, except for some of the amphibole-bearing (altered) mafic granulites. However, the altered amphibole-bearing granulites have lower Pb concentrations than the fresh amphibole-bearing mafic/ultramafic granulites. The alteration by the lamproite, if any, should have increased the Pb concentrations to be effective in changing the Pb isotopic ratios. Secondly, the lower crustal xenoliths and the lamproite host do not form any mixing

Figure 3.28 Plots of (a) $^{207}\text{Pb}/^{204}\text{Pb}$ versus $^{206}\text{Pb}/^{204}\text{Pb}$ and (b) $^{208}\text{Pb}/^{204}\text{Pb}$ versus $^{206}\text{Pb}/^{204}\text{Pb}$ for the lower crustal xenoliths from the Leucite Hills. The isochron age of 2477 +/- 200 Ma is similar to the age obtained from the Sm-Nd pseudo-isochron.

patterns on a $^{206}\text{Pb}/^{204}\text{Pb}$ versus Pb diagram (Fig. 3.29).

It is concluded that the whole rock Pb-Pb trend reflects an event of late Archean age. A compilation of the available Pb isotopic data for worldwide lower crustal xenoliths show that most granulite xenoliths have radiogenic Pb isotopic compositions and plot near the fields of MORB, OIB and upper crustal rocks. Rudnick (1992) proposes that the xenoliths plotting near the fields of MORB, OIB and upper crustal rocks became granulites (and underwent U-depletion) only recently. In the case of the granulite xenoliths from the Leucite Hills lamproites, the unradiogenic Pb signatures must have developed in regions that have remained stable for a very long period of time.

3.7 Discussion

3.7.1 Pressure-temperature estimates

A variety of geothermobarometers have been calibrated and refined over the last twenty years to provide pressure and temperature estimates of both igneous and metamorphic rocks. In the case of granulites, it is important to select the appropriate thermometers and barometers to be able to record peak metamorphic conditions. The geothermobarometers of pyroxene and garnet usually satisfy such conditions since the composition of pyroxene and garnet tend to be fixed at, or near, peak temperatures due to their low diffusion rates (Hewins, 1975). The geobarometers of Raheim and Green (1974), Ellis and Green (1979), Ganguly (1979) and Wells (1979) for garnet-clinopyroxene and that of Wood (1974), MacGregor (1974) and Akella (1976), Harley and Green (1982) and Harley (1984) for garnet-orthopyroxene pairs have been commonly used to calculate the pressures of equilibration for granulite xenoliths. To measure temperatures of equilibration, however, the orthopyroxene-clinopyroxene geothermometers of Wood and Banno (1973), Lindsley and Dixon (1976), Wells (1977) and Herzberg (1978), Lindsley (1983) and Kretz (1982) are widely used. A comprehensive review on geothermobarometry techniques applied to granulites and the detailed discussions of each method are reported in Kay and Kay (1981), Essene (1982) and Frost and Chacko (1989). Garnet has not been reported in any of the Leucite Hills lower crustal xenoliths (Kay et al., 1978; Barton and Bergen, 1988; Sperr, 1985; Lester and

Figure 3.29 Plot of $^{206}\text{Pb}/^{204}\text{Pb}$ versus Pb for the lower crustal xenoliths (LC xlth) and lamproite host (LH lamp) from the Leucite Hills. The absence of any correlation rule out the possibility that the lamproites have affected the Pb isotopic composition of the lower crustal xenoliths.

Framer, 1998) thus any estimate for the pressure of equilibration among the xenoliths is limited. Orthopyroxene and clinopyroxene, on the other hand, are present in some of the xenoliths and can be used to estimate the temperatures of granulite formation.

The state of equilibration among the mineral phases is usually indicated by the lack of compositional zonation. However, it is common that the chemical compositions of similar parts of the mineral pairs (i.e., core and core or rim and rim parts of the adjacent grains) are selected for geothermobarometry calculations (Harley, 1984; Loock et al. 1990; Schaaf et al., 1994).

The temperature of equilibration for the two-pyroxene granulites from Hatcher Mesa was calculated using combined two-pyroxene geothermometers of Wood and Bannos (1973) and Wells (1977). Ferric iron was calculated following the equation proposed by Droop (1987) in which $\text{Fe}_2\text{O}_3 = 1.1113 * \text{FeO}_T * \text{Fe}^{3+} / (\text{Fe}^{2+} + \text{Fe}^{3+})$. The results of the temperature calculations for the Leucite Hills two-pyroxene granulites are listed in Table 3.17. T1r and T2r are the calculated temperatures from the rims using the calibrations of Wood and Bannos (1973), and Wells (1977), respectively. T1c and T2c are similar calculation but based on the analysis of the cores. The temperature ranges for the rims (T1r = 764-876°C, T2r = 774-902°C) are slightly lower than those calculated for the cores (T1c = 850-883°C, T2c = 861-903°C). Considering the uncertainty inherent in each method ($\sim \pm 50^\circ\text{C}$), the results are in reasonable agreement and they point to an equilibration temperature of around 800 - 850°C. This temperature estimate confirms the observation of Kay et al. (1978) that the Leucite Hills granulites equilibrated at around 875°C. In general, temperatures of ca. 800°C are normally considered as the typical blocking temperature for equilibration in granulite xenoliths (Frost and Chacko, 1989; Loock et al., 1990; Carswell and O'Brien, 1993; Kempton et al. 1995). A compilation of pressure and temperature data shows that the majority of regional granulite terrains also exhibit peak metamorphic temperatures of $800 \pm 50^\circ\text{C}$ (Mezger, 1992).

As was mentioned earlier, the equilibrium pressures could not be calculated for the xenoliths due to the absence of suitable mineral barometers such as garnet. However, the equilibration pressure ranges may be inferred by comparing the granulite mineral assemblages to the phase stability fields determined for the gabbro to eclogite transitions under increasing pressure and temperature. Such an approach, undertaken by Ringwood

and Green (1966), Green and Ringwood (1967, 1971), Irving (1974b) and more recently by Wood (1987), encompasses the entire range of basaltic compositions including quartz tholeiite, olivine tholeiite, and alkali olivine basalt. The stability field shows that the appearance of garnet in mafic rocks is a function of not only pressure but also the chemical composition of whole rock. As a result, garnet appears at higher pressures in quartz tholeiite than it does in olivine tholeiite and alkali olivine basalt. The Leucite Hills two-pyroxene granulites are predominantly alkali-olivine basalt in composition (Section 3.7.1) and based on the garnet-in reaction curve (Fig. 3.30), a maximum pressure of 9 kb can be proposed for these rocks.

Because the two-pyroxene xenoliths do not contain any primary olivine, the olivine-out reaction curve can also be used to obtain the lower limit of the pressure range for the lower crustal xenoliths. The olivine-out reaction is not presently available for alkaline olivine basalt but Irving (1974b) examined such a reaction on a two-pyroxene granulite with the normative composition of olivine tholeiite. Irving's experiment (1974b) is deemed here to be the most appropriate for estimating the minimum pressure for the formation of the Leucite Hills granulites. Fig. 3.30 shows that the pressures of equilibration for the two-pyroxene granulite from the Leucite Hills lower crust can range from 4 to 9 kb (~10 - 30 km depth) at temperatures of around 800°C.

3.7.2 The origin of the Leucite Hills granulites

The characteristics of the lower crustal xenoliths discussed above have an important bearing on determining the origin of the igneous protoliths. The lower crust is generally considered to be either the result of crystallization of a mantle-derived magma or a mafic residue that remained after intracrustal partial melting (Rogers and Hawkesworth, 1982; Bradley, 1985). In the restite model, it is assumed that partial melting of an intermediate lower crust generates acidic melts to create the upper crust, leaving behind a residue that produces the mafic lower crust. Such residuum is expected to be characterized by lower K_2O and REE in the more basic samples and to show a limited range in the concentration of compatible elements in addition to a variable range in the abundances of incompatible elements (Rudnick and Taylor, 1987). The

Table 3.17 Temperature estimates for the Leucite Hills lower crustal xenoliths.

Sample #	T _{1r} (°C)	T _{2r} (°C)	T _{1c} (°C)	T _{2c} (°C)
157-6	876	902	850	863
157-8	844	835	883	885
157-25	848	846	883	885
157-36	764	774	853	903
157-38	804	815	838	861

Figure 3.30 P-T estimates for the Leucite Hills granulites (dotted rectangular).

Reaction curve: (1) garnet-in, quartz tholeiite, Green and Ringwood (1966), (2) garnet-in, alkali-olivine basalt, Green and Ringwood (1966), (3) olivine-in, olivine tholeiite, Irving (1974b). The position of data from the lower crustal xenoliths shows that the lower crustal granulites underlying the Leucite Hills lamproites equilibrated at temperature of $\sim 800^{\circ}\text{C}$ and pressures between 4 and 9 kb.

granulite xenoliths from Leucite Hills do not display any of the features pointing to a restite origin, but they do exhibit textures (e.g., cumulate and interstitial phases) indicative of underplating of a basic magma that has undergone crystal fractionation. Lower crustal xenoliths from Archean cratons, excluding those of the Snake River Plains and the Lace kimberlites, are predominantly mafic in composition and have also been interpreted as the crystallization products of mafic magma, resulting from basaltic underplating (Rudnick, 1992).

The overall gradational change in abundance of the whole rock major elements (section 3.7.1) and corresponding mineral compositions (section 3.5) suggest that the lower crustal xenoliths are samples of a differentiated igneous complex. In addition, the isotopic data from the lower crustal xenoliths that form near linear arrays on the Sm-Nd and Pb-Pb isotopic diagrams support the idea that the lower crustal xenoliths represent a cogenetic suite. In other words, the variety of lower crustal xenoliths from the Leucite Hills represent igneous rocks that have formed by crystal fractionation of a common magma.

The geothermobarometry calculations and Sm-Nd and Pb-Pb isochron ages on lower crustal xenoliths from the Leucite Hills lamproites indicate that granulites were probably formed in the lower parts of the Wyoming continental craton sometime during the late Archean (sections 3.4 and 3.5). There appears to be no petrogenetic relationship between the host lamproite and the xenolith suite.

An important geochemical finding from the granulite xenoliths is their pronounced depletion in Nb and Ta reflecting the possible involvement of subduction in their origin. The Beartooth Mountains contain a suite of compositionally diverse late Archean (2.74 – 2.79 Ga) andesites (Muller et al., 1983) for which the trace element patterns, particularly the Nb and Ta depletion, resemble those associated with subduction-related volcanism in modern arc environments. Based on these data and other geological observations, Mueller and Wooden (1988) and Wooden and Mueller (1988) suggest that the Wyoming Province may have been the site of a Late Archean convergent continental margin. Rocks similar in age and composition to the andesitic amphibolites of the Beartooth Mountains have been reported from the Bighorn Mountains, 200 km to the east (Barker et al., 1979) supporting widespread subduction in the Wyoming Archean

mantle. However, the subducted materials must have originated from a crustal source older than late Archean. Henry et al. (1982) have reported Rb-Sr and U-Pb zircon data from a lithologically heterogeneous assemblage of rocks that is 3.2 to 3.5 Ga old, located in the eastern Beartooth Mountains. All the mentioned evidence points to Archean subduction in the Wyoming Province that probably helped in the formation of the lower crustal rocks and accretion of the continental crust.

3.8 Summary

Several lines of evidence including texture, composition of pyroxenes and estimates of pressure-temperature equilibration indicate that the xenoliths originated from lower crustal depths. However, a meta-igneous rather than meta-sedimentary origin for the lower crustal xenoliths from the Leucite Hills lamproites is indicated by their petrographic and geochemical characteristics. The mineralogy and chemical composition of the xenoliths suggest that the bulk composition of the lower crust xenoliths is typical of mafic granulite xenoliths observed worldwide. Summarizing the evidence from thermobarometric calculations and experimental constraints, it is concluded that the mafic granulites experienced peak metamorphic temperature of $\sim 800^{\circ}\text{C}$ and a maximum pressure of 9 kb. The Sm-Nd and Pb-Pb pseudo-isochron ages suggest that the lower crustal granulites are late-Archean in age (ca 2.5 Ga). The lower crustal xenoliths are genetically unrelated to the young Leucite Hills lamproite magmatic activity but are related to an old basaltic phase of igneous activity. The magma that formed the protoliths of the lower crustal xenoliths is a fractionated mantle-derived melt that underplated at lower crustal levels. The Nd model ages show that the original magma was probably derived from a MORB-like depleted source that was fluxed with enriched metasomatic fluids emanating from subducted oceanic pelagic sediments. The noticeable arc-like character of the lower crustal xenoliths from the Leucite Hills favors a convergent plate margin. The lower crust of the Wyoming craton can thus be considered an example of continental growth by subduction. From the isotopic point of view, the most notable geochemical feature is the U-depleted and LREE-enriched nature of the lower crustal xenoliths from the Leucite Hills lamproites.

CHAPTER 4

PETROGRAPHY AND CHEMICAL COMPOSITION OF THE LEUCITE HILLS LAMPROITES

4.1 Introduction

The Leucite Hills lamproites have been extensively studied since the turn of the nineteenth century (e.g., Zirkel, 1867; Cross, 1897; Carmichael, 1967; Fraser, 1987; Mitchell, 1995a). However, a detailed and comprehensive study of their isotope and trace element compositions is lacking. In this chapter, the role of alteration and contamination in the genesis of the Leucite Hills samples are evaluated in the light of new trace element and isotopic data.

4.2 Petrography

Because the Leucite Hills lamproites have been described petrographically by Cross (1897), Kemp and Knight (1903), Kuehner (1980), Carmichael (1967), Fraser (1987) and Mitchell and Bergman (1991), only short petrographic descriptions of the studied samples are given here. More emphasis is directed towards the lamproites from the Pilot Butte, Middle Table Mountain, Badgers Teeth and South Table Mountain because the origin of the madupitic lamproites and olivine-bearing sanidine/leucite phlogopite lamproites is the focus of this and the following chapter.

The Leucite Hills lamproites consist mainly of a combination of phlogopite, diopside, sanidine, leucite, apatite, perovskite and minor K-rich richterite $[\text{Na}(\text{Na,Ca})\text{Mg}_5\text{Si}_8\text{O}_{22}(\text{OH})_2]$, wadeite $[\text{K}_2\text{ZrSi}_3\text{O}_9]$, priderite $[(\text{K,Ba,Cs})_{1-2}(\text{Fe,Cr,Al,V,Ce})_{1-2}(\text{Ti,Nb})_{7-6}\text{O}_{16}]$ and sherbakovite $[(\text{BaK})(\text{K,Na})\text{Na}(\text{Ti,Nb,Zr,Fe})_2\text{Si}_4\text{O}_{14}]$. Of these minerals, phlogopite, clinopyroxene, and apatite can occur as either phenocrysts and microphenocrysts or groundmass crystals, whereas the remainder of the minerals usually crystallize as groundmass phases. Olivine is also found in lamproites, but its occurrence is normally limited to large macrocrysts which have been interpreted as xenocrysts (Mitchell and Bergman, 1991), and phenocrysts (Foley, 1993).

Based on the old terminology, lamproites from the Leucite Hills are classified as wyomingites, orendites and madupites (Cross, 1897). Wyomingite contains phlogopite

and diopside phenocrysts/microphenocrysts residing in a groundmass of leucite, clinopyroxene, apatite and glass. Orendite has the same mineralogy as wyomingite but sanidine, in addition to leucite, is found in the groundmass. Madupite is characterized by a distinctive abundant phenocryst/microphenocryst assemblage of diopside and poikilitic phlogopite set in a groundmass of diopside, apatite and glass. Perovskite is normally more abundant in madupite than wyomingite and orendite.

Because lamproites from different parts of the world exhibit different and distinct mineralogies and textures, a number of diverse names have been used to classify them. To avoid the use of unusual and complicated names, Scott Smith and Skinner (1984a,b), Mitchell (1985), and Woolley et al. (1996) have revised the old terminology of lamproites and proposed a straightforward approach to the classification. These authors still use the term lamproite and the essential mineral phases are used as a modifying prefix, in order of increasing abundance, to describe the modal mineralogy of the lamproites. As such, the wyomingites and orendites are termed diopside leucite phlogopite lamproites and diopside sanidine phlogopite lamproites respectively. Mitchell (1985) also proposes that the term madupites be replaced with the term madupitic lamproites to indicate its textural difference from the other types of lamproites. Mitchell and Bergman (1991) suggest that those lamproites containing inclusion-free phlogopite phenocrysts mantled by poikilitic phlogopite be named transitional madupitic lamproites.

The terminologies recommended by Scott Smith and Skinner (1984a,b), Mitchell (1985), and Woolley et al. (1996) follow a logical approach, prevent a listing of complicated and unnecessary locality names, and at the same time clearly indicate the major mineral constituents of the lamproites. Thus, the names proposed by these authors are used here to classify the three main lamproite types in the Leucite Hills. It should be noted that for the most of this chapter, the diopside leucite phlogopite lamproite and diopside sanidine phlogopite lamproite from the Leucite Hills are simply called phlogopite lamproites to prevent long and repetitive usage of names. Moreover, as will be discussed later, these two lamproites are genetically related and no outstanding differences are observed in their chemical and isotopic compositions. Table 4.1 outlines the mineralogy of the different types of lamproite in the Leucite Hills and the locality where each lamproite type can be found.

Table 4.1. Minerals commonly found in the Leucite Hills lamproites, with increases in abundance from left to right (after Smith and Skinner, 1984a, b; Mitchell 1985, Woolley et al. 1996). The locality for each lamproite type is also given.

Rock Type (Old Terminology)	Oreudite	Wyomingite
Rock Type (Revised Terminology)	diopside sanidine leucite phlogopite lamproite	diopside leucite phlogopite lamproite
Principal Minerals	Di - Sa - (Lct) - Phl	Di - Lct - Phl
Accessory Minerals	Rct - Prd - Wad - Ap	Rct - Prd - Wad - Ap
Locality	North Table Mountain South Table Mountain Zirkel Mesa	Deer Butte Hatcher Mesa Zirkel Mesa

Rock Type (Old Terminology)	Madupite	Transitional madupite
Rock Type (Revised Terminology)	madupitic lamproite	transitional madupitic lamproite
Principal Minerals	Phl - Di	Phl - Lct - Di
Accessory Minerals	Rct - Prd - Wad - Prv	Rct - Prd - Wad - Prv
Locality	Pilot Butte	Middle Table Mountain Badgers Teeth

Abbreviations after Kretz (1983):

Ap: apatite, Di: diopside, Phl: phlogopite, Prd: priderite, Prv: perovskite, Rct: richterite, Sa: sanidine, Wad: wadeite, Lct: leucite.

The classification and description of the samples are discussed on a locality-to-locality basis. The sample numbers quoted below refer to those samples that were analyzed for their major element, trace element and isotopic compositions.

(1) The Pilot Butte samples are classified as madupitic lamproites. They consist of poikilitic phlogopite (0.5×1.2 mm to 0.6×0.6 mm) that contains abundant inclusions of clinopyroxene (0.1×0.5 mm to 0.25×0.25 mm) and perovskite ($8 \mu\text{m}$). Phlogopite is anhedral (Fig. 4.1) and diopside grains exhibit a distinct trachytic texture that can be traced into the phlogopite (Fig. 4.2). The groundmass consists of clinopyroxene, perovskite and yellowish glass. Leucite and sanidine are not found in the Pilot Butte madupitic lamproites. Whereas samples 133PLB, 135PLB, 138PLB and 140PLB contain abundant poikilitic phlogopite, samples 136PLB and 137PLB do not. These samples consist of phlogopite phenocrysts scattered in an extremely fine-grained matrix. Sample 137PLB is the only sample from Pilot Butte that shows evidence for the incorporation of sedimentary xenoliths. These xenoliths are shales from the Green River Formation and they vary in size reaching up to 20 cm. The sedimentary xenoliths show reaction rims around their margins. In extreme cases, some greenish stringers composed of broken sedimentary xenoliths and lamproite minerals are observed within the lamproite host. The clinopyroxene laths in the greenish stringers are surrounded by fine-grained material, and the optical characteristic of the clinopyroxene is obscured. Alteration of phlogopite in the greenish stringers has produced isotropic minerals.

(2) Middle Table Mountain contains the greatest volume of transitional madupitic lamproite in the Leucite Hills (Mitchell and Bergman, 1991). Phlogopite phenocrysts form two populations based on size (2×0.8 mm, 0.5×0.2 mm) and all have inclusion-free cores. At their margins they contain abundant inclusions of diopside, leucite, perovskite and apatite (Fig. 4.3). Small grains of perovskite (0.05 mm) are concentrated at the contact between the inclusion-free phlogopite phenocrysts and the outer poikilitic rim. The inclusion-free phlogopite shows light yellow to colorless pleochroism whereas the outer poikilitic phlogopite shows reddish brown to yellow pleochroism. In rare examples, the phenocrysts enclose either small olivine or diopside grains. Euhedral to subhedral diopside laths (0.3 mm), occurring as microphenocrysts, define a trachytic texture. Diopside also occurs as hexagonal-shaped microphenocrysts, 0.1×0.2 mm in

size. Twinning is common in diopside. The groundmass is composed of euhedral leucite (5 μm), perovskite, phlogopite, wadeite and glass. Unlike perovskite from Pilot Butte that is inclusion free, the perovskite from Middle Table Mountain is commonly poikilitic and contains inclusions of clinopyroxene, apatite and leucite (Fig. 4.4).

(3) The lamproite at Badgers Teeth is another example of transitional madupitic lamproite. Most phlogopite phenocryst cores are rimmed by a thin layer of poikilitic phlogopite containing inclusions of leucite, clinopyroxene, apatite and perovskite. In some of the Badgers Teeth samples (e.g., sample 112 BGT), phlogopite is slightly corroded at the edges and is surrounded by a thin dark halo, probably made up of Fe-oxides (Fig. 4.5). In other samples (e.g., sample 113BGT), the degree of resorption can be so great that the phlogopite phenocrysts become rounded at the edges. Carbonate and barite fill spaces between the phlogopite cleavage in some samples. Phlogopite phenocrysts typically show strong optical zonation. Clinopyroxene (0.25 $\times$ 0.4 mm) and apatite (0.4 $\times$ 0.8 mm) occur as euhedral to subhedral microphenocrysts and phenocrysts. Lamproites from Badgers Teeth have the highest content of apatite of any of the lamproites from Leucite Hills (Fig. 4.6). Some apatite grains (114BGT) are embayed (Fig. 4.7). The groundmass of the Badgers Teeth lamproite is composed of phlogopite (0.05 mm), leucite (0.05 - 0.1 mm), clinopyroxene (0.05 mm), apatite (0.02 mm) and perovskite (5 μm). Leucite is either fresh or pseudomorphosed and can contain inclusions of acicular diopside and apatite. Radiating euhedral carbonate patches, inferred to be of secondary origin, occur in the matrix of some samples.

Xenoliths of sedimentary rocks from the Blair Formation can also be found in some of the Badgers Teeth samples. The sedimentary xenoliths show a thin reaction rim around their margins but no intensive digestion by the lamproite melt appears to have taken place. Even though lamproites from both the Badgers Teeth and the Middle Table Mountain contain inclusion-free cores surrounded by the poikilitic rims, the poikilitic phlogopite rims from the Middle Table Mountain are more extensive. Groundmass perovskite is much finer grained in the Badger Teeth lamproite and is not poikilitic, unlike the perovskite in the Middle Table Mountain samples.

Fig. 4.1. Groundmass poikilitic phlogopite (Phl) in the madupitic lamproites containing abundant diopside (Di) and perovskite (Prv), Pilot Butte. Field of view 0.5 mm.

Fig. 4.2 Diopside in the madupitic lamproites (Pilot Butte) shows a trachytic texture that extends into the phlogopite grains. Field of view 0.5 mm.

Fig. 4.3 A typical phlogopite from the transitional madupitic lamproites, Middle Table Mountain, in which a euhedral inclusion free core is surrounded by anhedral phlogopite containing abundant diopside, leucite, perovskite and apatite. Field of view 0.5 mm.

Fig. 4.4. Perovskite (Prv) from the Middle Table Mountain characteristically contains micro inclusions of diopside (Di), apatite (Ap) and leucite (Lct). Size of grain 60 microns. Back scattered image.

Fig. 4.5. A corroded phlogopite rimmed by a thin layer of Fe-oxides. Badgers Teeth lamproite (Sample 112BGT). Field of view 0.5 mm.

Fig. 4.6. The Badgers Teeth lamproite has the highest modal abundance of euhedral to subhedral apatite. Field of view 1 mm.

(4) Lamproite from the South Table Mountain contains phenocrysts of phlogopite and microphenocrysts of phlogopite and clinopyroxene set in a groundmass of fine-grained leucite and/or sanidine, clinopyroxene and phlogopite. Leucite-rich and sanidine-rich layers form irregular layers (Fig. 4.8). Sanidine ranges from 0.05 mm to 0.4 mm in size. The larger sanidine crystals always contain inclusions of clinopyroxene and apatite. Lamproites from South Table Mountain contain large (1 mm to few cm), anhedral xenocrysts of olivine that show kink banding, undulatory extinction and irregular fracturing. Olivine is invariably rimmed by phlogopite laths (Fig. 4.9), the result of reaction between olivine and the lamproite melt (Carmichael, 1967). Because the olivine is xenocrystic and also forms less than 10% of the rock, the name olivine-bearing phlogopite lamproite should be used to describe lamproites from the South Table Mountain.

(5) The lamproites from Boars Tusk contain phlogopite and clinopyroxene microphenocrysts/phenocrysts set in a fine-grained matrix composed of clinopyroxene, leucite and phlogopite. Sanidine is not found among the mineral assemblages. In the groundmass, layers of oxidized and fresh leucite are intermingled. Some clinopyroxene grains contain green sieved cores (Figs. 4.10). Barton and van Bergen (1981) consider the green clinopyroxene to be the result of the disaggregation of lower crustal xenoliths. The Boars Tusk lamproite also contains abundant sedimentary and granitic xenoliths. Sample 106BT is characterized by having the greatest proportion of the sedimentary xenoliths (>25%).

(6) The phlogopite lamproites from Zirkel Mesa are very diverse in both their mineralogy and texture. They basically consist of elongated phenocrysts of phlogopite and microphenocrysts of phlogopite and diopside (rare) set in a groundmass of brownish glass, diopside, leucite and sanidine. The groundmass varies from cryptocrystalline (e.g., sample 119ZM) (Fig. 4.11) to slightly crystalline (e.g., sample 120ZM).

(7) Lamproites from Deer Butte contain phlogopite phenocrysts and diopside microphenocrysts set in a matrix of diopside, leucite and sanidine. Sanidine can also contain abundant needle-like diopside inclusions. The Deer Butte lamproites always contain upper crustal felsic xenoliths. The xenoliths are usually rounded, and show no sign of metamorphism or strain. The xenocrysts of quartz, orthoclase and microcline in

Fig. 4.7 Embayed apatite crystal in lamproite. Badgers Teeth lamproite. (Sample 114BGT). Field of view 0.2mm.

Fig. 4.8 In the diopside sanidine leucite phlogopite lamproites, the leucite-rich layers usually form irregular intervals or batches (dark) within the sandine-rich layers (light), South Table Mountain. Field of view 0.2 mm.

Fig. 4.9. Phlogopite laths surrounding the olivine xenocryst, South Table Mountain. Field of view 0.5mm.

Figs. 4.10. Green clinopyroxene phenocrysts with resorbed cores are most likely the result of disaggregation of lower crustal xenoliths. Boars Tusk. Field of view 0.2 mm.

Fig. 4.11. Cryptocrystalline groundmass of Zirkel Mesa lamproites. Field of view 0.5 mm.

Figs. 4.12. Quartz xenocrysts, Deer Butte. Field of view 0.5 mm.

quartz, orthoclase and microcline in the lamproites (Figs. 4.12) are probably the product of the disaggregation of upper crustal xenoliths. Some of the disaggregated felsic minerals are so fine-grained that they are not readily detectable under the microscope. It is therefore probable that the upper crustal xenocrysts were included during the preparation of the whole rock sample (See section 4.4.3 for more details).

(8) The mineralogy of the phlogopite lamproites from the North Table Mountain is very similar to those from the Zirkel Mesa. Layers of leucite-rich and sanidine-rich lamproites are common.

4.3 Mineral chemistry

Cross (1897), Carmichael (1967), Kuehner (1980) and Mitchell and Bergman (1991) have extensively studied the major element compositions of minerals from the Leucite Hills lamproites. Very limited trace element data for individual mineral phases are available. This probably results from the analytical difficulty and/or the difficulties encountered in separating mineral phases from such fine-grained rocks.

There have been some attempts, in the past, to measure the concentration of REE in the mineral phases from lamproites (Carmichael 1967; Kuehner et al. 1981; Mitchell and Steele 1992) and kimberlites (Jones and Wyllie, 1984; Mitchell and Reed 1988) but these studies were largely confined to the LREE. The HREE from these minerals, as well as some other trace elements, were not determined because of the low detection limits using an electron microprobe. The scanning electron microscopy and electron microprobe may cause serious analytical errors such as overestimation of HREE and underestimation of Sm (Jones and Wyllie, 1984). The Laser Ablation Microprobe-Inductively Coupled Plasma-Mass Spectrometer (LAM-ICP-MS), on the other hand, is capable of determining the concentrations of the REE down to ppm and/or ppb levels (Jackson et al., 1991). Several trace elements from apatite, perovskite, clinopyroxene and wadeite from the Leucite Hills and West Kimberley lamproites have been measured using the LAM-ICP-MS method (Mirnejad and Bell, 2000).

Mitchell and Bergman (1991) have presented major element compositions of minerals from the Leucite Hills lamproites. To prevent repetition, only a brief review of some of the outstanding chemical features is discussed here. This is followed by a

detailed description of the trace element compositions of diopside, perovskite and apatite from the Leucite Hills madupitic and transitional madupitic lamproites analyzed by LAM-ICP-MS. These minerals were selected because they are the principal hosts for important trace elements such as REE, Ba, Sr, Nb, Ta, Zr, Th, U and Pb commonly utilized in petrogenetic interpretations of lamproites. Apatite, clinopyroxene and perovskite are also the most abundant phases, in addition to phlogopite, in the Leucite Hills lamproites, and especially in the coarser-grained madupitic and transitional madupitic lamproites. Because one of the prime objectives of this thesis is to investigate the petrogenesis of the madupitic and transitional madupitic lamproites from the Leucite Hills, the trace element concentrations of the madupitic and transitional madupitic lamproites are discussed in detail. Trace element contents as well as major element abundances of apatite, perovskite and diopside are given in Tables 4.2 and 4.3 and 4.4, respectively. The analytical techniques for electron microprobe and LAM-ICP-MS are described in Appendix B.

4.3.1 Apatite

Apatite [$\text{Ca}_5(\text{PO}_4)_3(\text{OH}, \text{F}, \text{Cl})$] occurs as a minor mineral phase in most igneous rocks and many trace elements, particularly REE, can be incorporated into its crystal structure. Concentrations of up to 11.14 wt% REE as oxides have been reported from some apatites (Rae et al., 1996) and thus apatite plays a significant role in controlling the REE contents of the host rocks. The REE^{3+} substitution for Ca^{2+} in the apatite is usually coupled by replacement of Si^{4+} for P^{5+} or Na^+ for Ca^{2+} to maintain the charge balance.

The P_2O_5 (~ 41 wt%) and CaO (~ 53 wt%) contents of apatite from the Badgers Teeth (Table 4.2) are similar to those from the other occurrences in the Leucite Hills. Even though Cl and F contents of apatite could not be determined by electron microprobe, higher values of F (>1%) relative to Cl (<1%) have been reported for apatites from lamproites (Edgar 1989; Mitchell and Bergman 1991) and it is therefore likely that the analyzed apatites are fluor-apatite.

The REE concentrations of apatites have been normalized to those of chondrites (Sun and McDonough, 1989) and the ratios are plotted in Fig. 4.13a. The total REE content of apatite from the Badgers Teeth can reach 7660 ppm. The Badger Teeth apatites are clearly

Table. 4.2 Major element (wt%) and trace element (ppm) concentrations in apatite from the Badgers Teeth lamproites, Leucite Hills.

Sample	112 BGT	114BGT	116 BGT
CaO	53.19	53.13	53.12
P ₂ O ₅	41.56	41.57	41.47
MgO	0.46	0.00	0.00
MnO	0.00	0.00	0.00
SrO	1.72	1.69	1.50
Total	97.89	96.97	96.50
Rb	1.4 *		1.0
Sr	11300	11800	12200
Y	64	70	73
Zr	23	24	28
Nb	0.2	0.2	0.3
Cs	*	*	*
Ba	552	527	508
La	1500	1800	1800
Ce	3200	3600	3600
Pr	440	460	470
Nd	1450	1730	1750
Sm	180	200	200
Eu	40	41	40
Gd	63	77	79
Tb	6	6	6
Dy	19	21	22
Ho	2	3	3
Er	4	1	5
Tm	0.49	0.37	0.40
Yb	1.7	2.1	1.4
Lu	0.20	0.22	0.22
Hf	0.07	0.00	0.01
Ta	0.01	0.00 *	
Pb	2.3	3.1	4.4
Th	32	37	40
U	3	2	3

* below detectoin limit

Figure 4.13 (a) Chondrite-normalized REE diagram, and (b) Primitive mantle-normalized hygromagmatophile element diagram of apatite from the Badgers Teeth lamproite.

enriched in LREE relative to HREE and $(La/Yb)_N$ ranges from 640 to 942. The three analyzed apatites from the Badgers Teeth are similar to one another in composition. The abundances of some other trace elements such as high field strength elements (HFSE) and large ion lithophile elements (LILE), have been normalized to primitive mantle values (McDonough, 1998 in <http://www.earthref.org/GERM>) and plotted on the extended trace element diagram (Fig. 4.13.b). The Badgers Teeth apatites are enriched in Ba, Th, U and Y and depleted in Nb and Hf and Zr relative to the neighboring elements on the spider diagram. Y follows a trend similar to the neighboring REE.

4.3.2 Perovskite

Perovskite, $(CaTiO_3)$, is a common accessory phase in many lamproites. Table 4.3 shows the major and trace element composition of perovskite from the Pilot Butte and Middle Table Mountain lamproites. The Pilot Butte perovskite consistently shows higher concentrations of CaO (32-35 wt %) and TiO_2 (53-54 wt %) than that from Middle Table Mountain (CaO ~24 wt % and TiO_2 ~50 wt %). It seems that the lower concentrations of major elements in the Middle Table Mountain perovskite is compensated by higher REE and Nb_2O_5 contents than those of the Pilot Butte perovskite. For example, a negative correlation between REE and CaO reflects REE^{3+} substitution for Ca^{2+} .

The REE concentrations in perovskite have been normalized to chondrite values (Fig. 4.14.a). Perovskite is enriched in LREE relative to HREE and $(La/Yb)_N$ varies from 6500 in Middle Table Mountain to 2800 in the Pilot Butte perovskite. Perovskite from the Middle Table Mountain poikilitically encloses small grains of diopside, apatite and leucite. Inclusions in perovskite from Middle Table Mountain could not have contributed to the high LREE enrichment in perovskite because apatite and leucite normally have lower LREE concentrations than perovskite. Moreover, apatite and leucite inclusions were avoided during the laser ablation analysis, and hence the high levels of LREE in the Middle Table Mountain perovskite reflect solely the composition of the perovskite.

Data from perovskite show a slight depletion in U, Nb and Ta and a strong depletion in Hf and Zr (Fig. 4. 14.b). The concentrations of HFSE seems to be slightly higher in the Middle Table Mountain than those from the Pilot Butte.

Table 4.3 Major element (wt%) and trace element (ppm) composition of perovskite from Pilot Butte and Middle Table Mountain lamproites, Leucite Hills.

Sample	133PLB	135PLB	138PLB	140PLB	147MTM	149MTM
SiO ₂	0.39	0.46	0.31	0.95	0.82	0.65
TiO ₂	53.61	54.32	54.65	52.86	50.93	49.90
Nb ₂ O ₅	0.82	0.72	0.71	0.84	1.81	1.67
FeO _T	1.17	1.16	0.91	2.10	0.64	0.43
CaO	32.09	32.15	31.71	35.06	24.51	23.66
SrO	3.13	2.99	4.43	2.32	6.34	7.23
Na ₂ O	1.18	1.47	1.04	1.07	2.73	3.05
Ce ₂ O ₃	3.75	3.37	3.63	2.63	7.04	8.25
La ₂ O ₃	1.79	1.63	2.15	1.35	5.67	4.86
	97.93	98.27	99.54	99.18	100.49	99.70
Sr	19000	26000	30000	28000	76000	64000
Y	130	150	230	180	110	125
Zr	370	460	610	830	540	305
Nb	2500	3000	3800	4700	6800	6200
Cs	0.1	0.9	1.5	0.7	1.6	1.2
Ba	340	490	630	290	2200	2200
La	13600	17900	11400	12000	5300	3800
Ce	37000	35000	22000	24000	102000	72000
Pr	3200	3800	2300	2700	10000	7900
Nd	10000	13000	7700	7900	34000	24000
Sm	1000	1300	1000	900	2500	1900
Eu	200	240	230	160	390	370
Gd	290	320	350	230	480	504
Tb	26	30	39	20	38	31
Dy	78	73	110	49	91	76
Ho	9	9	10	9	9	6
Er	22	18	23	15	8	11
Tm	1.1	1.6	1.9	1.2	1.2	2.2
Yb	3.3	3.6	7.3	6.4	1.2	4.3
Lu	0.1	0.4	0.7	0.3	1	0.3
Hf	16.5	16.5	22.7	76.8	51.2	26.1
Ta	250	320	240	460	420	560
Pb	73	79	100	45	233	156
Th	1940	2600	1100	1900	5200	5500
U	79	86	199	58	98	109

Figure 4. 14 (a) Chondrite-normalized REE diagram, and (b) Primitive mantle-normalized hygromagmatophile element diagram for perovskite from Pilot Butte (PLB) and Middle Table Mountain (MTM) lamproites, Leucite Hills.

4.3.3 Clinopyroxene

Major element compositions of clinopyroxene from the Middle Table Mountain, Pilot Butte and Badgers Teeth lamproites are very similar (Table 4.4). The pyroxenes are all diopside (Wo_{45-50}) and show limited compositional trends towards hedenbergite and enstatite. Clinopyroxene is characterized by a high degree of LREE enrichment ($\text{La}_\text{N}/\text{Yb}_\text{N} = 80 - 123$) (Fig. 4.15.a). However, unlike perovskite, diopside from the Middle Table Mountain appears to be less enriched in LREE relative to those from the Pilot Butte and Badgers Teeth lamproites. The HREE show little variation among diopsides from the three localities ($\text{Yb}_\text{N} = 2.2 - 2.6$). On the primitive mantle-normalized spider diagram (Fig. 4.15.b), diopside shows negative troughs in Nb and Ta relative to the neighboring elements with similar incompatibility.

In general, the trace element distribution patterns of the clinopyroxenes from the Leucite Hills madupitic and transitional madupitic lamproites are similar to those of apatite, except that diopside does not show a strong negative anomaly in Hf. Apatite also contains higher abundances of incompatible trace elements than diopside.

4.3.4 Phlogopite, olivine, amphibole, leucite and sanidine

The trace element contents of phlogopite, olivine, amphibole, leucite and sanidine were not measured by LAM-ICP-MS due to the very low concentrations of many of the trace elements and/or the difficulty in obtaining appropriate grain separates from whole rock samples. The following is a brief review of the major element composition of phlogopite, amphibole, leucite and sanidine from the Leucite Hills lamproites analyzed by electron microprobe at Carleton University. Representative compositions are listed in Table 4.5.

Phlogopite shows the most extensive chemical variations among the lamproite mineral phases. The phenocrysts are titanian (TiO_2 1-3 wt%) Al-poor (Al_2O_3 11-13 wt%) phlogopite and the poikilitic crystals are titanian (TiO_2 4-11 wt%) Al-deficient (Al_2O_3 5-11 wt%) tetraferriphlogopite. The mantling phlogopite around phlogopite phenocrysts in the transitional madupitic lamproites is similar in composition to the poikilitic groundmass phlogopite from the madupitic lamproites. The mantling mica around the olivine xenocrysts, however, is similar in composition to the phenocrystal phlogopite.

Table 4.4 Major element (wt%) and trace element (ppm) composition of clinopyroxene from the Middle Table Mountain, Pilot Butte and Badgers Teeth lamproites, Leucite Hills.

Sample	112BGT	114BGT	116BGT	135PLB	138PLB	140PLB	147MTM	149MTM
SiO ₂	54.66	55.06	54.77	53.98	54.77	55.52	54.96	52.71
Al ₂ O ₃	0.30	0.46	0.27	0.43	0.32	0.00	0.00	0.61
FeO _T	1.88	2.01	2.19	3.00	2.03	2.59	2.99	4.08
TiO ₂	0.43	0.85	0.70	1.43	0.70	0.37	1.27	2.78
Cr ₂ O ₃	0.26	0.00	0.14	0.20	0.13	0.21	0.41	0.39
MgO	18.56	18.02	18.06	16.90	17.83	17.73	17.09	16.53
CaO	24.20	25.12	25.23	24.98	25.55	24.89	23.84	22.96
Na ₂ O	0.00	0.00	0.00	0.55	0.00	0.00	0.54	1.30
Total	100.29	101.52	101.36	101.47	101.33	101.44	101.10	101.36
Sr	1840	1930	2270	2240	1730	1590	1570	1580
Y	6.0	9.0	10.0	6.0	6.0	5.0	7.0	8.0
Zr	166.0	196.0	246.0	133.0	152.0	144.0	194.0	217.0
Nb	0.48	0.75	1.94	0.23	0.54	0.36	1.69	1.10
Cs	*	*	0.05	*	*	0.03	0.13	0.14
Ba	42.0	20.0	37.0	43.0	38.0	39.0	85.0	66.0
La	58.0	55.0	91.0	55.0	54.0	87.0	46.0	44.0
Ce	138.0	161.0	202.0	124.0	127.0	204.0	120.0	123.0
Pr	19.0	23.0	30.0	17.0	17.0	25.0	19.0	19.0
Nd	75.0	101.0	119.0	70.0	68.0	90.0	76.0	80.0
Sm	11.0	15.0	17.0	12.0	10.0	13.0	12.0	12.0
Eu	2.7	3.4	4.5	2.5	2.6	2.7	2.7	3.3
Gd	4.7	7.5	8.4	5.2	4.9	4.9	5.2	6.1
Tb	0.47	0.74	0.86	0.51	0.53	0.52	0.56	0.56
Dy	1.5	3.2	3.3	2.0	1.8	1.6	1.9	2.6
Ho	0.24	0.46	0.50	0.34	0.22	0.21	0.32	0.32
Er	0.41	0.74	0.83	0.29	0.47	0.32	0.57	0.78
Tm	0.05	0.11	0.10	0.04	0.04	0.04	0.06	0.08
Yb	0.34	0.44	0.57	0.33	0.46	0.36	0.31	0.50
Lu	0.05	0.05	0.09	0.10	0.04	0.04	0.07	0.05
Hf	8.1	11.5	14.9	7.6	7.8	6.5	8.5	11.8
Ta	0.02	0.04	0.16	0.00	0.08	0.05	0.09	0.03
Pb	1.2	0.8	1.0	1.7	0.2	1.0	0.9	1.0
Th	0.77	0.61	3.35	1.32	1.04	3.12	0.62	0.40
U	0.06	0.13	0.13	0.15	0.05	0.14	0.11	0.06

* below detectoin limit

Table 4.5 Representative major element compositions (wt%) of phlogopite, olivine, amphibole, leucite and sanidine from the Leucite Hills lamproites.

phlogopite phenocrysts						
Sample	BGT	MTM	BT	DEB	ZM	NTM
SiO ₂	40.87	40.74	40.86	40.92	41.02	40.69
Al ₂ O ₃	12.24	12.36	12.05	12.52	12.03	11.87
Cr ₂ O ₃	0.98	0.86	0.54	0.79	0.50	0.61
TiO ₂	2.21	1.97	2.02	1.89	2.12	2.20
FeO _T	2.99	2.87	2.76	2.92	2.79	3.01
MgO	25.03	25.03	25.22	25.32	24.96	25.82
Na ₂ O	0.18	0.12	0.12	0.09	0.13	0.12
K ₂ O	10.88	10.99	10.87	11.02	10.95	10.73
Total	95.38	94.94	94.44	95.47	94.50	95.05

groundmass phlogopite			
Sample	BGT	MTM	PLB
SiO ₂	41.56	41.77	40.96
Al ₂ O ₃	4.69	5.9	7.26
Cr ₂ O ₃	0.12	0.01	0.15
TiO ₂	5.82	7.38	5.14
FeO _T	10.92	8.87	9.48
MgO	20.86	20.48	22.83
Na ₂ O	1.56	0.99	0.52
K ₂ O	9.77	10.38	10.13
Total	95.3	95.78	96.47

	olivine		amphibole		leucite	sanidine
SiO ₂	39.21	SiO ₂	52.81	SiO ₂	57.18	64.21
FeO _T	8.86	Al ₂ O ₃	0.83	Al ₂ O ₃	19.6	16.17
MnO	0.18	TiO ₂	8.66	Fe ₂ O ₃	2.01	3.29
MgO	51.76	FeO _T	4.01	Na ₂ O	0.01	0.15
CaO	0.01	MgO	17.69	K ₂ O	21.18	15.89
NiO	0.08	CaO	6.15	Total	99.98	99.71
Total	100.10	Na ₂ O	4.55			
		K ₂ O	4.79			
		Total	99.49			

BGT: Badgers Teeth, MTM: Middle Table Mountain, BT: Boars Tusk, DEB: Deer Butte, ZM: Zirkel Messa, NTM: North Table Mountain, PLB: Pilot Butte

Olivine is forsteritic in composition. Richterite is a K- and Ti- rich variety of amphibole (i.e., titanian potassium richterite, Mitchell and Bergman, 1991). Leucite is Fe-rich ($\text{Fe}_2\text{O}_3 = 1\text{-}2 \text{ wt}\%$) and Al-poor ($\text{Al}_2\text{O}_3 = 19\text{-}21 \text{ wt}\%$). Similarly, sanidine is Fe-rich ($\text{Fe}_2\text{O}_3 = 1\text{-}5 \text{ wt}\%$).

4.4 Geochemistry

Ten whole rock lamproites and two sedimentary xenoliths were analyzed for their major and trace element concentrations by ICP and ICP-MS at Activation Laboratories, Ancaster, Ontario. Sr, Nd and Pb isotopic ratios were measured on a Finnigan-MAT 261 multicollector mass spectrometry at Carleton University. Details of the sample preparation and analytical techniques are given in Appendices A and B.

4.4.1 Major element geochemistry

The major element compositions of the Leucite Hills lamproites have been reported by Carmichael (1967), Ogden (1979), Kuehner (1980) and Fraser (1987). Lamproites are Mg-rich igneous rocks ($\text{MgO} > 5 \text{ wt}\%$) that are considered peralkaline ($\text{K}_2\text{O} + \text{Na}_2\text{O}/\text{Al}_2\text{O}_3 > 1$) (molar), perpotassic $\text{K}_2\text{O}/\text{Al}_2\text{O}_3 > 1$ (molar) and ultrapotassic $\text{K}_2\text{O}/\text{Na}_2\text{O} > 3$ (molar). The general compositional ranges of lamproites are: 45 – 55 wt% SiO_2 , 4 – 10 wt% Al_2O_3 , 1- 5 wt% TiO_2 , 2 – 10 wt% CaO , 5 – 10 wt% K_2O , 0.2 – 1.5 wt% Na_2O and 0.5 – 2.0 wt% P_2O_5 (Bergman, 1987).

The major element abundances and CIPW normative compositions of the Leucite Hills lamproites are presented in Table 4.6. For normative calculation, FeO and Fe_2O_3 were recalculated from $\text{Fe}_2\text{O}_3\text{T}$ following the recommendation of Middlemost (1989). The oxide contents of the Leucite Hills lamproites have been plotted in Figure 4.16.

The Pilot Butte and Badgers Teeth lamproites have the highest MgO and lowest SiO_2 content of any of the Leucite Hills lamproites. The data from transitional madupitic lamproites from Middle Table Mountain with abundant sedimentary xenoliths plot between the madupitic lamproites and the phlogopite lamproites. However, if the effect of sedimentary xenoliths in changing the chemical composition of the whole rock is taken into account (see section 4.4.3 for more details), the data from the Middle Table Mountain lamproites would group with those from the Pilot Butte and Badgers Teeth

Table 4.6 Major element (wt%) composition of the Leucite Hills lamproites. The normative compositions have only been included for the uncontaminated lamproites.

Sample	110 BGT	112 BGT	114 BGT	116 BGT	133 PLB	135 PLB
SiO ₂	43.2	41.94	42	43.94	42.2	41.4
TiO ₂	2.29	2.19	2.16	2.24	2.28	2.23
Al ₂ O ₃	7.85	7.72	8.01	8.36	7.46	7.42
MgO	10.47	9.96	9.95	9.27	11.98	12.22
CaO	11.06	12.29	10.83	10.02	12.44	12.72
MnO	0.1	0.1	0.11	0.09	0.13	0.12
Fe ₂ O ₃ T	5.94	5.58	5.76	5.69	6.38	6.24
Na ₂ O	1.23	1.06	1.7	0.9	0.55	0.48
K ₂ O	6.85	4.74	6.2	8.14	5.39	5.13
P ₂ O ₅	2.92	3.01	2.84	2.99	1.63	1.87
LOI	5.42	9.05	6.8	5.86	6.51	6.43
Total	97.34	97.64	96.36	97.5	96.95	96.24
Mg#	0.76	0.76	0.76	0.75	0.77	0.78

$$\text{Mg\#} = \text{Mg}/(\text{Mg} + 0.899 \text{ Fe}^{+2})$$

Sample	110 BGT	112 BGT	114 BGT	116 BGT	133 PLB	135 PLB
Qtz						
An		2.31			1.97	2.94
Di	28.09	30.59	27.6	23.72	38.34	37.46
Ol	9.15	7.45	8.4	8.47	8.45	9.16
Hmt	1.68	3.92	1.69	1.66	4.48	4.38
Chr						
Ne	1.21	4.16	3.62		2.52	2.2
Cm						
Or	9.1	27.88	10.23	17.04	7.58	6.5
Ab		1.29				
Hy						
Mt						
Ilm	3.98	3.75	3.9	3.81	4.32	4.21
Ap	6.29	6.48	6.11	6.44	3.51	4.03
Lc	24.54		20.65	22.34	18.98	18.63

Table 4.6 (continued)

Sample	136 PLB	137 PLB	138 PLB	140 PLB	147 MTM	149 MTM
SiO ₂	42.4	45.27	42.89	42.96	47.39	48.26
TiO ₂	2.25	2.05	2.26	2.23	2.37	2.55
Al ₂ O ₃	7.2	7.64	7.57	7.51	8.66	8.63
MgO	11.19	10.63	11.31	10.92	8.87	9.09
CaO	12.38	11.04	12.06	12.13	9.34	8.78
MnO	0.12	0.12	0.12	0.12	0.09	0.1
Fe ₂ O ₃ T	6.16	5.89	6.3	6.41	5.45	5.65
Na ₂ O	0.89	1.67	0.54	0.94	1.18	0.96
K ₂ O	6.59	8.03	7.4	8.69	7.29	9.57
P ₂ O ₅	1.83	2.05	1.78	2.39	1.3	1.4
LOI	6.98	4.48	6.03	4.59	6.79	4.48
Total	97.99	98.86	98.27	98.89	98.74	99.46
Mg#	0.77	0.76	0.76	0.75	0.75	0.74

Sample	136 PLB		138 PLB	140 PLB		
Qtz						
An						
Di	30.68		32.58	27.09		
Ol	9.59		9.16	10.26		
Hmt	2.13		3.02	1.99		
Chr						
Ne	0.19					
Crn						
Or						
Ab						
Hy						
Mt				0.12		
Ilm	4.17		4.26	4.24		
Ap	3.94		3.83	5.15		
Lc	30.5		32.37	32.11		

Table 4.6 (continued)

Sample	150 STM	151 STM	104 BT	119 ZM	131 DEB	141 NTM	157 HAM
SiO ₂	53.86	53.07	51.64	55.06	52.58	55.82	49.71
TiO ₂	2.39	2.39	2.364	2.732	2.627	2.582	2.28
Al ₂ O ₃	9.12	9.07	9.79	9.87	9.70	9.99	9.87
MgO	9.79	9.72	7.49	6.28	7.48	6.44	8.03
CaO	4.02	3.86	5.99	4.11	5.09	3.30	7.32
MnO	0.06	0.06	0.07	0.06	0.07	0.05	0.09
FeO _T	4.72	4.61	5.11	4.28	5.08	4.13	5.52
Na ₂ O	1.08	1.17	1.60	0.98	1.28	1.69	1.85
K ₂ O	11.36	10.51	10.60	11.54	10.79	10.46	10.55
P ₂ O ₅	1.57	1.35	1.61	1.29	1.87	1.46	1.42
LOI	2.31	2.31	2.62	3.23	2.37	3.46	2.01
Total	100.28	98.12	98.88	99.42	98.93	99.38	98.64
Mg#	0.79	0.79	0.73	0.73	0.73	0.74	0.72

Sample	150 STM	151 STM		119 ZM	131 DEB	141 NTM
Qtz				3.62		4.93
An						
Di	7.71	8.19		9.43	10.37	5.45
Ol	1.86	2.12			2.53	
Hmt						
Chr						
Ne						
Cm						
Or	49.55	49.28		53.63	52.7	54.28
Ab						
Hy	19.51	18.63		11.58	11.59	13.9
Mt						
Ilm	4.55	4.55		5.19	5	4.91
Ap	3.38	2.91		2.78	4.03	3.14
Lc

Figure 4. 16.a-f Diagrams showing the major element chemical variations of the Leucite Hills lamproites. Dashed arrow shows the direction that the olivine-bearing phlogopite lamproite from the South Table Mountain (STM) would take if the effect of olivine xenocrysts is taken into account. If there were no sedimentary xenoliths in the Middle Table Mountain lamproite (MTM), the data from this lamproite would group with those from Pilot Butte (PLB) and Badgers Teeth (BGT).

lamproites. The olivine-bearing phlogopite lamproite from South Table Mountain has similar MgO to, and lower SiO₂ than, the madupitic and transitional madupitic lamproites. As listed in Table 4.6, the Mg# of the olivine-bearing phlogopite lamproites (0.79) is slightly higher than that of the madupitic and transitional madupitic lamproites (0.74 – 0.76) and the phlogopite lamproites (0.72 – 0.73). The SiO₂ content of the olivine-bearing phlogopite lamproites is similar to that of the olivine-barren phlogopite lamproites. In the other major element variation diagrams, the data from the olivine-bearing lamproites commonly plot within the phlogopite lamproite fields. The high MgO content and Mg# of the olivine-bearing phlogopite lamproites probably reflects the presence of olivine. Since the olivine is most likely xenocrystic (Sections 4.2), the high MgO and Mg# content of the lamproites from South Table Mountain cannot be considered characteristic of the melt. If the effect of increasing the MgO content of the whole rock is incorporated by taking into account the presence of olivine xenocrysts, data from the olivine-bearing phlogopite lamproites would plot within the field of phlogopite lamproites. The effect of this is indicated by the dashed arrows shown in the MgO - SiO₂ diagram (Fig. 4.16)

The Fe₂O₃T contents of lamproites follow more or less similar trends to the MgO contents (see Fig. 4.16). The Fe₂O₃T concentrations are the highest in the madupitic and transitional madupitic lamproites and the lowest in the phlogopite lamproites. The Fe₂O₃T contents of the olivine-bearing phlogopite lamproites are similar to the phlogopite lamproites and form one group in the Fe₂O₃T versus SiO₂ diagram.

Samples from madupitic lamproites and transitional madupitic lamproites have CaO contents higher than the phlogopite lamproites. The Na₂O concentration of the madupitic lamproites from Pilot Butte is lower than most of the other Leucite Hills lamproites. The Badgers Teeth transitional madupitic lamproites, like the phlogopite lamproites, show a wide range in Na₂O concentrations from 0.9 to 1.7 wt%. The phlogopite lamproites have noticeably higher K₂O contents than the madupitic and transitional madupitic lamproites. The Al₂O₃ abundance of the madupitic and transitional madupitic lamproites is also less than that of the phlogopite lamproites.

The different lamproite types from the Leucite Hills do not differ greatly in their TiO₂ abundances even though the phlogopite lamproites show slightly higher levels of

TiO_2 than the madupitic and transitional madupitic lamproites. The transitional madupitic lamproites from the Badgers Teeth locality show the highest concentrations of P_2O_5 among any of the lamproites in Leucite Hills and this is reflected in the high modal abundances of apatite in these rocks. The LOI contents of the madupitic and transitional madupitic lamproites ($> 6 \text{ wt\%}$) are higher than those of the phlogopite lamproites ($< 3 \text{ wt\%}$).

Regardless of the effects of sedimentary xenoliths and olivine xenocrysts on changing the chemical composition of some lamproites, the major element variations of the Leucite Hills lamproites can be explained by mineralogical controls. The higher MgO and FeO abundances of the madupitic and transitional madupitic lamproites relative to the phlogopite lamproites is due to the differences in the modal abundances of ferromagnesian minerals such as clinopyroxene or mica. The madupitic and transitional madupitic lamproites are characterized by higher CaO and LOI reflecting higher modal diopside and phlogopite in these rocks relative to the phlogopite lamproites. Modal phlogopite does not seem to affect the K_2O content of the lamproites as much as sanidine and leucite because the latter minerals have K_2O , Na_2O and Al_2O_3 abundances higher than phlogopite. As a result, due to the absence or low modal abundance of sanidine and leucite, the madupitic and transitional madupitic lamproites have lower K_2O and Al_2O_3 concentrations than those of the phlogopite lamproites. The madupitic and transitional lamproites contain higher modal perovskite than the phlogopite lamproites, yet their TiO_2 contents are comparable or slightly lower than those of the phlogopite lamproites, suggesting that minerals other than perovskite have contributed to the TiO_2 contents of the lamproites (e.g., phlogopite).

In addition to modal mineralogy, the normative composition of the Leucite Hills lamproites can be used to evaluate the petrogenesis of the lamproites. Table 4.6 includes the CIPW normative composition of the Leucite Hills lamproites calculated using the IGNEOUS Program. The madupitic and transitional madupitic lamproites have higher normative olivine than most of the phlogopite lamproites. The high normative olivine and the lack of normative hypersthene from the madupitic and transitional madupitic lamproites point to the silica undersaturated nature of their magmas. The higher normative diopside in the madupitic and transitional madupitic lamproites than that of the

phlogopite lamproites reflects the higher CaO of the former. CaO can also combine with P_2O_5 to form apatite and thus the normative diopside is less abundant in the Badgers Teeth than the rest of the madupitic and transitional madupitic lamproites. The only exception is sample 112BGT that has normative diopside comparable to that of other madupitic and transitional madupitic lamproites. However, the CaO content of this sample is higher than other samples from Badgers Teeth. This is also the only sample among the madupitic and transitional madupitic lamproites that does not contain any normative leucite. Careful examination of sample 112BGT shows the occurrence of secondary carbonate, thus explaining the high CaO content of this sample.

The olivine-bearing phlogopite lamproites from the South Table Mountain contain the highest normative hypersthene contents of any of the Leucite Hills lamproites. The phlogopite lamproites have high abundance of normative orthoclase with no normative leucite (except sample 104BT which has only 0.3% normative leucite). The North Table Mountain and Zirkel Mesa phlogopite lamproites are the only analyzed samples containing normative quartz. These samples are the most silica-saturated and have the highest SiO_2 and the lowest MgO and CaO of any of the Leucite Hills lamproites analyzed in this study. The normative composition of sample 157HAM, as well as samples from Middle Table Mountain, Boars Tusk and Deer Butte, were not calculated because of the presence of numerous crustal xenoliths and/or alteration (see section 4.4.3).

4.4.2 Trace element geochemistry

Like other ultrapotassic rocks, lamproites are characterized by high to extreme enrichment in incompatible elements (e.g., Ba > 2000 ppm, Sr > 1000 ppm, La > 200 ppm). Ni and Cr abundances of lamproites are higher than those of many other alkaline igneous rocks (Peccerillo, 1992) and on trace element spider diagrams, lamproites usually show negative Nb and Ta anomalies (Mitchell and Bergman, 1991). The trace element contents of the Leucite Hills lamproites have been analyzed and described by Kuehner (1980), Fraser (1987) and Mitchell and Bergman (1991). The compilation of REE data from different lamproites worldwide (Mitchell and Bergman, 1991) shows that the

Leucite Hills lamproites do not differ greatly from other lamproite types in terms of REE abundance and LREE enrichment.

A number of whole rock samples, including different types of lamproites from Leucite Hills, were analyzed for their trace element concentrations using ICP-MS. The results are given in Table 4.7.

The Ni contents of lamproites show some variations among the different lamproite types. For example, the madupitic and transitional madupitic lamproites have lower Ni concentrations than the phlogopite lamproites (Fig. 4.17). The olivine-bearing phlogopite lamproites from the South Table Mountain have the highest Ni concentrations (300 ppm) of any of the phlogopite lamproites. This is best attributed to the presence of xenocrystic olivine. The concentration of Ni in olivine xenocrysts ranges between 8000 ppm (this work) and 3000 ppm (Mitchell and Bergman, 1991). Considering the 10% modal olivine xenocrysts in the olivine-bearing phlogopite lamproites, the removal of this amount of olivine could lower the Ni content of the whole rock to values comparable to those of the phlogopite lamproites. Thus, if the effect of olivine xenocrysts is taken into account, the Ni concentration of the lamproite melt from the South Table Mountain may not be too different from the rest of the phlogopite lamproites. The dashed arrow on the Cr–Ni diagram shows the direction that the data would follow by removing the effect of xenocrystic olivine in South Table Mountain samples.

The REE concentrations of the Leucite Hills lamproites have been normalized to primitive mantle values and plotted in Figure 4.18.a. Only data from those lamproite samples not considered to be contaminated by the sedimentary xenoliths (section 4.4.3) have been included in this figure. The Leucite Hills lamproites are characterized by high concentrations of REE and enriched LREE over HREE ($(La/Yb)_N = 150 - 600$). When the normalized REE contents of the madupitic lamproites are compared to those of perovskite, apatite and clinopyroxene (Fig. 18.b), it can be seen that the mineral separates and the whole rocks follow similar patterns. One conclusion to be made from this observation is that the REE contents of the Leucite Hills lamproites are dominantly hosted by perovskite and apatite. LREE concentrations and the degree of LREE enrichment over HREE are higher in the madupitic and transitional madupitic lamproites

Table 4.7 Trace element (ppm) composition of the Leucite Hills lamproites.

Sample	110 BGT	112 BGT	114 BGT	116 BGT	133 PLB	135 PLB	136 PIB
V	129	65	116	114	31	21	54
Cr	469	489	471	471	546	560	523
Co	25	24	27	27	29	30	27
Ni	142	148	109	128	120	121	161
Cu	38	34	30	28	37	42	54
Zn	97	64	72	66	81	74	116
Ga	24	23	25	24	23	23	21
Ge	1.6	1.6	2.1	2	2.1	2.3	1.9
As	9	*	*	*	*	*	*
Rb	166	226	215	268	177	166	206
Sr	6223	5323	5745	5250	7233	6628	4881
Y	23	22	20	20	22	22	25
Zr	234	140	634	788	705	793	263
Nb	109	98	127	127	139	136	133
Mo	*	*	0.4	0.6	0.3	0.3	*
Ag	*	1.8	*	*	*	*	*
In	*	*	*	0.1	0.1	0.1	*
Sn	2	1	3.9	5.2	5.5	3.7	4
Sb	0.3	*	0.35	0.35	0.57	0.55	0.6
Cs	1.7	2.3	5.1	4.6	3.7	3.7	3.1
Ba	8171	10300	12500	9920	11020	11690	10590
La	397	402	370	395	338	321	371
Ce	735	754	739	778	685	653	695
Pr	81.5	83.3	81.4	85.6	77.4	73.5	79.5
Nd	292	300	286	299	276	265	294
Sm	33.0	33.8	34.6	35.6	35.7	33.9	35.7
Eu	7.0	7.0	7.1	7.6	7.7	7.3	7.5
Gd	20.6	21.5	13.4	13.9	15.9	14.1	22.2
Tb	1.6	1.6	1.3	1.3	1.5	1.4	1.8
Dy	5.4	5.1	5.1	5.5	5.8	5.5	5.8
Ho	0.7	0.7	0.6	0.7	0.7	0.7	0.7
Er	1.4	1.4	0.7	0.8	0.9	0.8	1.5
Tm	0.1	0.1	0.1	0.1	0.1	0.1	0.1
Yb	0.57	0.49	0.43	0.46	0.52	0.46	0.61
Lu	*	*	0.09	0.10	0.09	0.10	*
Hf	4.6	2.4	19	22	10	8.8	1.3
Ta	5.4	5.3	4.78	4.45	6.23	5.51	6.9
W	0.7	1.0	0.2	1.4	0.2	0.3	1.6
Tl	1.20	0.25	1.58	1.46	0.46	0.94	0.15
Pb	42	31	43	52	53	43	52
Bi	0.35	0.22	0.77	1.2	1.27	1.05	0.27
Th	44.2	43.4	45.9	48	47.4	47.1	42.8
U	11.4	10.1	10.2	11.4	11	10.7	10.7

* below detection limit

Table 4.7 (continued)

Sample	137 PLB	138 PIB	140 PIB	147 MTM	149 MTM	150 STM	151 STM
V	56	109	69	119	114	86	85
Cr	444	541	517	408	415	477	473
Co	27	26	26	25	24	28	26
Ni	104	155	157	115	156	333	275
Cu	49	44	45	28	39	32	22
Zn	99	70	68	61	115	96	74
Ga	24	20	23	23	25	22	22
Ge	1.9	2.1	1.6	2	1.9	1.5	1.8
As	*	*	6	*	*	*	*
Rb	212	258	220	231	302	266	281
Sr	5719	4126	3814	2965	3051	2132	2149
Y	27	20	27	16	18	16	15
Zr	529	850	253	808	737	417	588
Nb	136	123	123	97	104	47	46
Mo	*	0.8	*	0.3	*	*	0.8
Ag	*	*	*	*	*	*	*
In	*	*	*	*	*	*	*
Sn	5	3.3	3	3.9	5	5	4.1
Sb	0.4	0.49	0.4	0.41	0.3	0.2	0.27
Cs	3.8	4.1	2.9	2.9	3.1	2.0	2.5
Ba	7212	7556	9087	8453	6451	4471	4777
La	376	373	351	271	262	140	133
Ce	712	740	664	546	504	271	269
Pr	80.1	74.9	75.0	55.2	56.3	31.1	31.1
Nd	300	273	280	200	208	118	112
Sm	36.0	34.9	33.8	24.9	24.1	15.0	15.4
Eu	7.7	6.6	7.1	4.6	5.1	3.3	3.5
Gd	22.7	16.5	20.7	12.1	14.5	9.86	7.5
Tb	1.8	2.1	1.7	1.5	1.2	0.9	0.8
Dy	6.1	5.9	5.9	4.2	4.1	3.5	3.5
Ho	0.8	0.7	0.8	0.6	0.6	0.5	0.5
Er	1.7	0.8	1.7	0.7	1.2	1.1	0.8
Tm	0.2	0.1	0.2	0.1	0.1	0.1	0.1
Yb	0.77	0.4	0.82	0.37	0.60	0.59	0.52
Lu	*	0.11	*	0.09	*	0.06	0.09
Hf	4.7	20	1.3	12	8.1	3.3	3
Ta	6.7	8.81	6.5	6.76	4.9	2.0	2.22
W	0.4	1.7	0.3	1.6	0.5	1.1	0.9
Tl	0.47	1.24	0.39	0.46	0.74	0.85	1.01
Pb	120	52	26	80	47	23	29
Bi	0.31	1.97	0.44	0.83	0.38	0.18	0.68
Th	44.8	43.2	42.0	34.4	33.5	15.5	16.4
U	10.7	9.74	8.13	7.38	7.07	5.05	4.85

Table 4. 7 (continued)

Sample	104 BT	119 ZM	131 DEB	141 NTM	157 HAM
V	125	80	98	90	111
Cr	453	310	395	343	396
Co	21	18	22	21	25
Ni	186	180	266	226	136
Cu	86	29	44	35	37
Zn	114	62	71	67	75
Ga	22	23	23	24	24
Ge	1.9	1.8	1.7	1.9	2.1
As	9	*	*	*	-5
Rb	246	296	273	288	229
Sr	2270	2020	2720	1830	3293
Y	15.4	15.7	17.6	12.9	20
Zr	126	265	1180	159	529
Nb	57.5	39.6	53.1	41.9	71
Mo	*	*	*	*	0.4
Ag	*	*	*	*	*
In	*	*	*	*	*
Sn	4	6	7	5	3.3
Sb	1.1	0.4	0.6	0.6	0.3
Cs	4.5	1.8	3.1	2.2	2.5
Ba	8510	6670	6900	6240	6232
La	174	150	178	119	324
Ce	326	297	346	236	650
Pr	37.5	36.1	42.1	27.5	66.3
Nd	121	124	140	96.8	238
Sm	15.6	15.6	18.4	12.8	30.1
Eu	3.4	3.5	4.1	2.7	5.7
Gd	9.15	9.92	10.4	8.02	15.3
Tb	0.9	0.9	1.0	0.7	1.8
Dy	3.4	3.5	3.9	3.0	5.4
Ho	0.5	0.6	0.6	0.5	0.7
Er	1.1	1.1	1.1	0.9	0.8
Tm	0.1	0.1	0.1	0.1	0.1
Yb	0.5	0.37	0.48	0.37	0.43
Lu	0.06	0.03	0.05	0.04	0.12
Hf	1.2	2.5	1.2	1.4	11
Ta	3.6	2.4	3	2.6	4.95
W	*	*	*	0.7	6.7
Tl	0.35	0.29	0.46	0.54	0.41
Pb	35	25	24	32	119
Bi	0.21	0.1	0.26	0.19	1.14
Th	22.4	16.7	19.4	13.7	33.9
U	6.28	3.4	5.16	3.78	6.84

Figure 4. 17 Diagram showing variations in Ni and Cr contents of the Leucite Hills lamproites. Symbols as in Fig.4.16.

Figure 4.18 (a) Chondrite-normalized concentration of the REE from the Leucite Hills lamproites shows that madupitic and transitional madupitic lamproites are characterized by higher REE concentrations and higher degree of LREE enrichments than the phlogopite lamproites. (b) Separated mineral phases from madupitic and transitional madupitic lamproites follow similar REE trends to that of the Leucite Hills lamproites. (Prv = perovskite, Ap = apatite, Di = Diopside)

$((\text{La}/\text{Yb})_N = 300 - 650)$ than in the phlogopite lamproites $((\text{La}/\text{Yb})_N = 170 - 250)$. The olivine-bearing phlogopite lamproites from South Table Mountain and the phlogopite lamproites from other localities in the Leucite Hills have overlapping in their REE patterns. This again supports the proposal that olivine-bearing phlogopite lamproites should indeed be grouped together with other phlogopite lamproites from the Leucite Hills.

Trace element data from the Leucite Hills lamproites have been normalized to the primitive mantle values and plotted on a conventional spider diagram (Fig. 4.19.a) in order of decreasing chemical incompatibility from left to right. Data from those samples considered to have been contaminated by crustal materials have not been included. The normalized trace element patterns for different lamproite types from Leucite Hills have similar configurations with enrichment in many incompatible trace elements and a pronounced Nb-Ta trough. The patterns shown by the whole rocks are broadly similar to those of the mineral separates (Fig. 4.19.b). In general, the madupitic and transitional madupitic lamproites are more enriched in the LILE and HFSE than the phlogopite lamproites. The trace element concentrations of the olivine-bearing phlogopite lamproite from South Table Mountain are indistinguishable from those of the phlogopite lamproites.

The LILE (Cs, Rb, Ba, K) show less variation among the different types of lamproites compared to other incompatible elements. The concentration of Cs in the Leucite Hills lamproites varies from 1 to 4 ppm and no clear distinction is visible in terms of Cs contents of the different lamproite types. The Ba content of the Leucite Hills lamproites (4000 – 11,000 ppm) is substantially greater than those of the neighboring elements with similar incompatibility such as Cs (1 – 5 ppm) and Rb (150 – 300 ppm).

In the whole rock, the minerals that most likely host Ba are priderite $\{(\text{K}, \text{Ba})_{1-2}(\text{Fe}, \text{Cr}, \text{Al}, \text{V}, \text{Ce})_{1-2}(\text{Ti}, \text{Nb})_{7-6} \text{O}_{16}\}$, phlogopite, sanidine and Ba-rich apatite. The lack of any correlation between Ba and P_2O_5 abundances (Fig. 4.20) rules out the possibility of apatite as the major host for Ba. In addition, the trace element variation diagrams (Fig. 4.19b) shows that apatite and perovskite are not Ba-rich and thus could not have

contributed greatly to the Ba enrichment of the whole rock. The positive correlation between Ba and LOI shown in Fig. 4.20 points to phlogopite as the major Ba holder phase in the Leucite Hills lamproites. The concentrations of Rb is lower than Ba in the Leucite Hills lamproites. The strong positive correlation between Rb and K_2O indicates that Rb is likely to concentrate in phlogopite, sanidine, and leucite (Fig. 4.21). Rb is also more enriched in samples with higher abundances of Al_2O_3 and lower contents of LOI indicating that sanidine and leucite, in addition to phlogopite, control the Rb content of the phlogopite lamproites. On the other hand, the madupitic lamproites with almost no leucite or sanidine are characterized with lower Rb concentration.

Minerals that could contribute to the U and Th budget of the Leucite Hills lamproites are perovskite, priderite and apatite. In the phlogopite lamproites, Th and U show a strong positive correlation with CaO, Nb, Ce and P_2O_5 (Figs. 4.22 and 4.23), suggesting that perovskite and apatite control the Th and U contents of these lamproites. Such positive correlations are not observed in the madupitic lamproites, but madupitic lamproites have higher Th (45 ppm) and U (10 ppm) concentrations than those of the phlogopite lamproites (Th = 4 ppm and U = 15 ppm). Considering the higher Nb and Ce contents of the former, it is likely that U and Th substitute for Nb and Ce in the Nb-, Ce-rich perovskite and priderite.

The concentrations of Ta and Nb in the madupitic lamproites are 6 and 125 ppm, respectively, while those of the phlogopite lamproites are 3 and 47 ppm, respectively. Perovskite, priderite, ilmenite and magnetite are usually considered the major host of Nb and Ta in lamproites (Mitchell and Bergman, 1991). Microprobe analysis of perovskite (Table 4.3) shows that Nb can be as high as ~ 1 wt% which confirms that perovskite is the main host of Nb in lamproites. In addition, madupitic lamproites usually have groundmass priderite that could also contribute to the higher concentrations of Nb. Ta normally shows a geochemical affinity to Nb and it is expected that perovskite and priderite also host Ta.

The Sr concentration of the madupitic and transitional madupitic lamproites (3000-5000 ppm) is higher than that of the phlogopite lamproites (~2000 ppm). This clearly correlates with the CaO concentrations in the lamproites and reflects the control of Ca-rich minerals such as diopside, apatite and perovskite on the Sr abundances of the

Figure 4.19 (a) The extended trace element concentrations of the Leucite Hills lamproites normalized to primitive mantle. (b) The extended trace element concentrations (average) patterns of separated minerals from madupitic and transitional madupitic lamproites are included for comparison with those of the Leucite Hills lamproites.

Figure 4.20 a-d The variation of Ba versus P_2O_5 , K_2O , TiO_2 and LOI shows that phlogopite is the main carrier of Ba in the Leucite Hills lamproites. Note that the phlogopite lamproites and madupitic lamproites form two distinct groups. Symbols as in Fig. 4.16.

Figure 4.21 a-c The Rb versus K_2O , Al_2O_3 and LOI variation diagrams show that sanidine and leucite rather than phlogopite are the main minerals controlling the behavior of Rb in the Leucite Hills lamproites. Symbols as in Fig. 4.16.

Figure 4.22 a-d The positive correlation between Th and CaO, Nb and Ce shows that the Nb-bearing perovskite is the phase most likely controlling the Th content of the Leucite Hills lamproites. Symbols as in Fig. 4.16.

Figure 4.23 a-d The positive correlation between U and CaO , Nb and Ce shows that the Nb-bearing perovskite is the phase most likely controlling the U content of the Leucite Hills lamproites. The positive correlation between P_2O_5 and U shows that apatite is also an important phase in affecting the U content of the Leucite Hills lamproites. Symbols as in Fig. 4.16.

whole rock. The microprobe and LAM-ICP-MS analyses indicate that perovskite is much more enriched in Sr than the apatite and diopside (Fig. 4.19.b) and thus plays a key role in contributing to the Sr budget of the whole rock.

The average concentrations of Zr and Hf from the Leucite Hills lamproites (500 ppm and 7 ppm, respectively) reported in Table 4.7 are much lower than the values compiled by Mitchell and Bergman (1991) (> 1000 ppm and > 40 ppm). These authors have cited Powell and Bell (1970), Kuehner (1980) and Fraser (1987) as references for their compilation of Hf and Zr from the Leucite Hills lamproites. A search among these references reveals that Fraser (1987) reports only Hf concentrations (> 40 ppm). Neither Zr nor Hf values are found in any of the other references. The LAM-ICP-MS data for the mineral separates from the Leucite Hills lamproites show a pronounced negative Zr and Hf anomaly in perovskite and apatite (Figs. 4.13-14) that is consistent with what is observed in most of the whole rocks. Because the trace element data are not available for phlogopite, sanidine and leucite from the Leucite Hills, any conclusions drawn from the LAM-ICP-MS should be taken with care. In addition, the whole rock Zr and Hf values show more scatter than any of the other trace element data plotted on the spider diagram (Fig. 4.19a). Because of the uncertainties inherited in measuring Zr and Hf concentrations by the ICP method, the Zr and Hf values of the Leucite Hills lamproites are not discussed here.

Y is a moderately incompatible element and shows little deviation in concentration from the HREE in both whole rock and minerals separates.

4.4.3 Identifying altered and contaminated lamproites

Some of the analyzed lamproite samples contain carbonate veins and crustal xenoliths that are usually impossible to avoid during sample preparation. Before discussing the isotope geochemistry of the Leucite Hills lamproites, evaluation is made of the effect of alteration and possible crustal contamination on the geochemistry of the analyzed samples.

The effects of crustal contamination and secondary carbonate alteration are evaluated by: (1) analyzing the leachate, residual and unleached lamproite samples for their isotopic ratios, (2) comparing the major and trace element composition and isotopic

ratios of lamproites and their crustal xenoliths, and (3) measuring the isotopic ratios of separated mineral phases.

4.4.3.1 Leaching tests

Leaching experiments have already been shown to be important in detecting the presence of crustal components in some North American kimberlites (Alibert and Albarede, 1988). In contrast to kimberlites that are CO₂-rich, lamproites are normally H₂O-rich, and do not contain primary carbonate phases. Hence, any high CO₂ content in the lamproites is probably related to secondary alteration.

A selected number of whole rock lamproites from the Leucite Hills with the possible presence of secondary alteration and/or interaction with crustal xenoliths, were leached with 1.5 molar HCl to remove any carbonate minerals. Leachates, residual materials and whole rocks were analyzed for Sr, Nd and Pb isotopic ratios. Table 4.8 compares the isotopic compositions of the whole rock, leachate and residual portions.

It can be seen that the Nd isotopic ratios of the whole rock lamproites are similar to the leachates for all runs. This means that any secondary carbonate alteration has not affected the Nd isotopic composition of the lamproites. On the other hand, it appears that the Sr isotopic ratios from half of the samples (i.e., those from Boars Tusk, Deer Butte, Pilot Butte and Middle Table Mountain) show differences among the whole rock, leachate and residual portions (Fig. 4.24). For example, the ⁸⁷Sr/⁸⁶Sr ratio of the leachate from sample 147MTM is 0.70694 compared with values of 0.70636 and 0.70606 in the unleached and residual samples, respectively. Furthermore, a subset of these five samples also appears to have Pb isotopic systematics which show alteration effect. Because leachate portions contain HCl-soluble fractions such as carbonate, the isotopic ratios of the residual parts should normally be closer to the true isotopic composition of the fresh lamproites.

The leaching tests show that the isotopic compositions of certain samples (e.g., MTM, DEB, BT and 137PLB) have been affected by secondary carbonate alteration. The phlogopite lamproites from Boars Tusk and Deer Butte show variations in the Sr isotopic ratios in the leachate and residual portions but not in their Nd and Pb isotopic ratios.

Table 4.8 The Sr, Nd and Pb isotopic compositions of the leachates (lcht), residual (res) and the unleached (unlch) portions of the analyzed lamproites from the Leucite Hills.

Rock Type	Sample	$^{87}\text{Sr}/^{86}\text{Sr}$	$^{143}\text{Nd}/^{144}\text{Nd}$	$^{206}\text{Pb}/^{204}\text{Pb}$	$^{207}\text{Pb}/^{204}\text{Pb}$	$^{208}\text{Pb}/^{204}\text{Pb}$
Trans Madp Lamp	110BGT unlch	0.70545	0.51203	17.457	15.484	37.272
Trans Madp Lamp	110BGT-licht	0.70545	0.51201	17.461	15.479	37.264
Trans Madp Lamp	110BGT-res	0.70539	0.51200	17.450	15.470	37.150
Madp Lamp	133 PLB unlch	0.70549	0.51210	17.556	15.508	37.489
Madp Lamp	133 PLB-licht	0.70550	0.51204	17.576	15.518	37.503
Madp Lamp	133 PLB-res	0.70546	0.51211	17.563	15.502	37.493
Madp Lamp	137 PLB unlch	0.70550	0.51205	17.576	15.506	37.559
Madp Lamp	137 PLB-licht	0.70562	0.51205	17.571	15.506	37.582
Madp Lamp	137 PLB-res	0.70548	0.51202	17.542	15.482	37.502
Phg Lamp	141NTM unlch	0.70591	0.51188	17.273	15.482	37.280
Phg Lamp	141NTM-licht	0.70591	0.51187	17.318	15.479	37.289
Phg Lamp	141NTM-res	0.70596	0.51186	17.292	15.475	37.291
Phg Lamp	143NTM unlch	0.70603	0.51186	17.281	15.482	37.278
Phg Lamp	143NTM-licht	0.70609	0.51186	17.279	15.484	37.273
Phg Lamp	143NTM-res	0.70613	0.51187	17.270	15.485	37.289
Ol Phg Lamp	151STM unlch	0.70592	0.51181	17.250	15.460	37.170
Ol Phg Lamp	151STM-licht	0.70595	0.51181	17.247	15.466	37.158
Ol Phg Lamp	151STM-res	0.70598	0.51180	17.271	15.487	37.199
Phg Lamp	119ZM unlch	0.70574	0.51187	17.227	15.464	37.318
Phg Lamp	119ZM-licht	0.70572	0.51187	17.239	15.475	37.325
Phg Lamp	119ZM-res	0.70575	0.51191	17.253	15.492	37.424

Red numbers represent data from samples that show isotopic ratio differences between leachate, residual and unleached portions. Trans Madp Lamp = transitional madupitic lamproites, Madp Lamp = madupitic lamproites, Phg Lamp = phlogopite lamproites, Ol Phg Lamp = olivine-bearing phlogopite lamproites.

Table 4.8 (continued)

Rock Type	Sample	$^{87}\text{Sr}/^{86}\text{Sr}$	$^{143}\text{Nd}/^{144}\text{Nd}$	$^{206}\text{Pb}/^{204}\text{Pb}$	$^{207}\text{Pb}/^{204}\text{Pb}$	$^{208}\text{Pb}/^{204}\text{Pb}$
Trans Madp Lamp	147MTM unlich	0.70637	0.51209	17.534	15.491	37.501
Trans Madp Lamp	147MTM-licht	0.70694	0.51208	17.592	15.513	37.521
Trans Madp Lamp	147MTM-res	0.70606	0.51207	17.492	15.490	37.385
Phg Lamp	104BT unlich	0.70597	0.51193	17.469	15.499	37.581
Phg Lamp	104BT-licht	0.70609	0.51195	17.445	15.478	37.523
Phg Lamp	104BT-res	0.70589	0.51193	17.450	15.482	37.518
Phg Lamp	126DEB unlich	0.70597	0.51184	17.585	15.514	37.498
Phg Lamp	126DEB-licht	0.70585	0.51184	17.583	15.523	37.519
Phg Lamp	126DEB-res	0.70573	0.51184	17.580	15.524	37.508
Phg Lamp	131DEB unlich	0.70581	0.51184	17.565	15.535	37.589
Phg Lamp	131DEB-licht	0.70609	0.51185	17.586	15.522	37.507
Phg Lamp	131DEB-res	0.70587	0.51184	17.566	15.518	37.491

Figure 4.24 Diagram showing the Sr and Nd isotopic variations among the leachate, residual and unleached samples from the Leucite Hills lamproites.

4.4.3.2 Chemical composition of the sedimentary xenoliths

The major and trace element abundances of two of the sedimentary xenoliths are listed in Table 4.9. The major element compositions of the xenoliths are different from those of the lamproites (Fig. 4.25). The sedimentary xenoliths have higher SiO_2 and Al_2O_3 and lower MgO and FeO contents than their lamproite hosts. The chemical composition of lamproite sample 137PLB suggests that some sedimentary xenoliths may have been digested by the lamproites. Sample 137PLB has higher SiO_2 , Al_2O_3 and lower MgO , FeO , CaO than the rest of the madupitic lamproites from Pilot Butte. On the major element diagrams, the data from sample 137PLB plot towards the data from the sedimentary xenoliths (Fig. 4.25). Similarly, the Middle Table Mountain lamproites also trend towards the sedimentary xenoliths on the major element diagrams. Since the sedimentary xenoliths enclosed in the Middle Table Mountain lamproites cannot be avoided during sample crushing, the deviation of their major element compositions from the rest of the madupitic and transitional madupitic lamproites is not surprising. If the effect of sedimentary contamination is taken into account, the Middle Table Mountain lamproites should be included with the other madupitic and transitional madupitic lamproites from Pilot Butte and Badgers Teeth.

As was noted earlier, the presence of abundant sedimentary xenoliths (but not necessarily the assimilation of these by the lamproite host) has shifted some of the major element abundances of the madupitic lamproites to levels closer to those of the phlogopite lamproites. The only exceptions are the TiO_2 and P_2O_5 abundances that seem to be unaffected by contamination. The Middle Table Mountain lamproites contain abundant microphenocrysts of perovskite and needle-like apatite in the matrix. Because these lamproites are rich in Ti and P, they may have buffered the whole rocks from the effect of the sedimentary xenoliths contamination.

Five sedimentary xenoliths were analyzed for their Sr, Nd and Pb isotopic compositions and abundances (Tables 4.10). It is not intended here to examine in detail the role of assimilation in the petrogenesis of lamproites, because it will be discussed later. However, a brief discussion of these data may help identify those samples whose isotopic compositions have been changed by crustal assimilation and/or the physical presence of the sedimentary xenoliths. On a $^{87}\text{Sr}/^{86}\text{Sr}$ versus Sr concentration

Table 4.9 Major element (wt%) and trace element (ppm) composition of the sedimentary xenoliths from the Leucite Hills lamproites.

Sample	106 BT _(xth)	133 PLB _(xth)	Sample	106 BT _(xth)	133 PLB _(xth)
SiO ₂	59.05	54.8	V	51	101
TiO ₂	0.462	0.43	Cr	30	68
Al ₂ O ₃	14.62	11.19	Co	3	5.9
MgO	1.05	4.17	Ni	17	16
CaO	5.09	10.64	Cu	32	21
MnO	0.03	0.04	Zn	*	95
FeO _T	1.71	3.23	Ga	23	18
Na ₂ O	1.08	0.35	Ge	0.6	1.1
K ₂ O	12.66	10.7	As	10	*
P ₂ O ₅	2.41	2.4	Rb	259	144
LOI	2.77	2.6	Sr	2170	2756
Total	100.95	100.55	Y	34	22
			Zr	20	171
			Nb	27.9	14
			Mo	*	1.1
			Ag	*	*
			In	*	*
			Sn	*	*
			Sb	0.7	0.95
			Cs	1	1
			Ba	4820	2722
			La	63.4	29.1
			Ce	114	55.4
			Pr	12.9	6.8
			Nd	45.8	27
			Sm	7.69	5.41
			Eu	1.7	1.0
			Gd	6.18	4.53
			Tb	0.9	0.7
			Dy	5.1	4.2
			Ho	1.1	0.8
			Er	3.1	2.2
			Tm	0.4	0.4
			Yb	2.88	2.37
			Lu	0.41	0.38
			Hf	0.7	3.9
			Ta	1.4	2.44
			W	*	3.6
			Tl	1.51	0.4
			Pb	89	272
			Bi	0.16	0.83
			Th	9.13	8.76
			U	5.89	3.92

* below detection limit

Figure 4. 25.a-f Diagrams showing the major element chemical variations of the Leucite Hills lamproites and their sedimentary xenoliths (Sed Xlth). The data from the transitional madupitic lamproites from Middle Table Mountain (MTM) with abundant sedimentary xenoliths points towards the composition of sedimentary xenoliths. If there were no sedimentary xenoliths in the MTM lamproite, the data from these lamproites would group with those from the the madupitic lamproites (Madp lamp) and away from the phlogopite lamproites (Phg lamp). Sample 137PLB contains abundant sedimentary xenoliths and thus plots to the extreme right of the madupitic lamproite field.

diagram (Fig. 4.26.a), the data from sample 147MTM, a transitional madupitic lamproite which contains sedimentary xenoliths, plot closest to the sedimentary xenoliths suggesting assimilation. It is difficult to estimate to what extent sample 137PLB has been affected by the sedimentary xenoliths on either the $^{87}\text{Sr}/^{86}\text{Sr}$ versus Sr or the $^{143}\text{Nd}/^{144}\text{Nd}$ versus Nd diagrams (Fig. 4.26.a, and b). A plot, however, of $^{206}\text{Pb}/^{204}\text{Pb}$ versus Pb concentration (Fig. 4.26.c) shows clearly that the data from this sample plots closer to the Pb composition of the sedimentary xenoliths.

Comparison between the major and trace element content and isotopic composition of the Leucite Hills lamproites with those of the sedimentary xenoliths reveals that the chemical composition of a number of lamproites such as the transitional madupitic lamproites from Middle Table Mountain (147MTM and 149MTM) and one madupitic lamproite from Pilot Butte (137PLB) have been changed due to the presence of and/or reaction with sedimentary xenoliths.

4.4.3.3 Isotopic composition of the mineral separates

The isotopic composition of the mineral separates from the Leucite Hills lamproites is expected to more accurately represent the initial isotopic ratios of the magma. Because certain minerals are less susceptible to alteration than the whole rocks, the isotopic composition of the mineral phases can help place some constraints on the initial isotopic compositions of the lamproite magma. For example, the studies of isotopic composition of the perovskite grains from the eastern North America kimberlites (Heaman, 1989) shows that the Sr and Nd isotopic ratios of the perovskites are less radiogenic than the whole rock values.

In an attempt to evaluate the effect of alteration and crustal contamination on the isotopic compositions of the lamproites, separated mineral phases (i.e., apatite, diopside and perovskite) from Badgers Teeth, Pilot Butte and Middle Table Mountain were analyzed for their Sr, Nd and/or Pb isotopic ratios (Table 4.11). The approach was restricted only to the madupitic and transitional madupitic lamproites and was not extended to the phlogopite lamproites due to the rarity of favorable minerals (i.e., apatite, diopside and perovskite) and the very small size of the mineral grains. Because apatite, diopside and perovskite are the major carriers of Sr and Nd in lamproites, these

Figure 4. 26 (a) $^{87}\text{Sr}/^{86}\text{Sr}$ versus Sr, (b) $^{143}\text{Nd}/^{144}\text{Nd}$ versus Nd, and (c) $^{206}\text{Pb}/^{204}\text{Pb}$ versus Pb diagrams for the sedimentary xenoliths and lamproites from the Leucite Hills. STM samples are included with phlogopite lamproites. Symbols as in Fig. 4. 16.

minerals provide a unique opportunity to look at the combined Sr and Nd isotopic compositions of mineral phases from the same rock.

The $^{87}\text{Sr}/^{86}\text{Sr}$ ratios show that the Pilot Butte (excluding sample 137PLB) and Badgers Teeth transitional madupitic/madupitic lamproites are in isotopic equilibrium with their mineral phases, but in the case of the Middle Table Mountain lamproite the isotopic ratios for the diopside and perovskite are different from those of the whole rock, residual and leachate (Table 4.11). The leaching test has previously shown that the real Sr and Pb isotopic composition of the Middle Table Mountain lamproites should be lower than that of the whole rock. It is interesting to note that even though the whole rock $^{87}\text{Sr}/^{86}\text{Sr}$ ratio (0.706062) for sample 147MTM is different from that of sample 149MTM (0.705685), the Sr isotopic ratio is identical for the mineral separates from these two samples (0.70551). Assuming the mineral grains were in isotopic equilibrium with the whole rocks, the Sr isotopic ratio of the mineral separates are thus considered to be the true isotopic composition of the transitional madupitic lamproites from Middle Table Mountain. The Nd isotopic ratios show no substantial differences between the whole rock and the mineral separates from the Leucite Hills transitional madupitic/madupitic lamproites thus suggesting that the sedimentary xenoliths and secondary alteration processes have had no effect on the Nd isotopic compositions of the madupitic lamproites.

Because of the very low Pb concentrations in the mineral separates, the Pb isotopic ratios could only be obtained for diopside separates from Pilot Butte and diopside and perovskite from Middle Table Mountain (Table 4.11). The Pb isotopic variations between the whole rock samples and minerals separates are within analytical uncertainty and therefore they are in isotopic equilibrium.

Consistent with the leaching test and the chemical analysis of the sedimentary xenoliths, the isotopic analysis of the individual grains from the Leucite Hills lamproites shows that the Sr isotopic composition of the whole rock samples from Middle Table Mountain is not representative of the parental melt. It appears that the isotopic composition of the mineral grains more accurately represent the isotopic composition of the lamproitic melt from which the lamproites from the Middle Table Mountain

Table 4.10 The Sr and Pb isotopic composition and concentration of sedimentary xenoliths (xlth) from the Leucite Hills lamproites.

Sample	$^{87}\text{Sr}/^{86}\text{Sr}$	$^{143}\text{Nd}/^{144}\text{Nd}$	$^{204}\text{Pb}/^{204}\text{Pb}$	$^{207}\text{Pb}/^{204}\text{Pb}$	$^{208}\text{Pb}/^{204}\text{Pb}$	Sr	Nd	Pb
106BT-xlt	0.707861	0.511809	17.68	15.508	37.722	1871	40	89
114BGT-xlt	0.706855	0.512161	19.893	15.726	39.462	3253	120	
133PLB-xlt	0.706396	0.512003	17.659	15.494	37.55	2390	25	272
137PLB-xlt	0.705755	0.512033	17.55	15.495	37.47	2982	112	
140PLB-xlt	0.705694	0.512073	17.595	15.501	37.517	3182	168	

Table 4.11 The isotopic composition of separated minerals from the madupitic lamproites, Leucite Hills.

Sample	$^{87}\text{Sr}/^{86}\text{Sr}$	$^{143}\text{Nd}/^{144}\text{Nd}$	$^{204}\text{Pb}/^{204}\text{Pb}$	$^{207}\text{Pb}/^{204}\text{Pb}$	$^{208}\text{Pb}/^{204}\text{Pb}$
112 BGT w.r.	0.705380	0.511972			
112 ap	0.705309	0.511974			
112 cpx	0.705323	0.511968			
113 BGT w.r.	0.705343	0.512027			
113 ap	0.705296	0.511982			
133 PLB w.r.	0.705455	0.512100	17.556	15.508	37.489
133 cpx	0.705463	0.512071			
135 PLB w.r.	0.705511	0.512085	17.542	15.489	37.48
135 cpx	0.705452	0.512044	17.531	15.482	37.463
136 PLB w.r.	0.705450	0.512088	17.547	15.496	37.517
136 cpx	0.705463	0.512062			
147 MTM w.r.	0.706062	0.512072	17.534	15.491	37.501
147 cpx	0.705529	0.512067			
149 MTM w.r.	0.705685	0.512056	17.535	15.496	37.512
149 cpx	0.705514	0.512055	17.527	15.458	37.375
149 pv	0.705501	0.512050	17.519	15.468	37.455

w.r. = whole rock, ap = apatite, cpx = clinopyroxene, pv = perovskite.

crystallized. Thus, a preference is given to the isotopic composition of the mineral grains if isotopic data from both the whole rock and mineral separates are available.

4.4.4 Isotope geochemistry of whole rocks

The Sr, Nd and Pb isotopic ratios of the whole rock lamproites from this study are listed in Table 4.12 and plotted on Figs. 4.27 and 4.28. Only the data from those samples that are not considered contaminated or altered have been used. In addition, in cases where the isotopic composition of the whole rock and mineral separates were different, those of the minerals were used for the isotopic interpretations.

The Nd-Sr isotopic plot (Fig. 4.27) shows that the madupitic and transitional madupitic lamproites have higher $^{143}\text{Nd}/^{144}\text{Nd}$ ratios (0.51203 - 0.51211) than the phlogopite lamproites (0.51186 - 0.51194). The $^{143}\text{Nd}/^{144}\text{Nd}$ in the olivine-bearing phlogopite lamproites vary from 0.51178 to 0.51183. The $^{87}\text{Sr}/^{86}\text{Sr}$ ratios also indicate differences between the two groups. Ratios vary from 0.70534 - 0.70563 for the madupitic and transitional madupitic lamproites to 0.70566 - 0.70606 for the phlogopite lamproites. The Sr isotopic ratios of the olivine-bearing phlogopite lamproites are comparable to those of the phlogopite lamproites and range from 0.70584 to 0.70608.

On the Pb – Pb isotopic ratio diagrams, the data from the phlogopite lamproites and olivine-bearing phlogopite lamproites usually plot at lower Pb isotopic ratios than those of the madupitic and transitional madupitic lamproites (Fig. 4.28a and 4.28b). The $^{206}\text{Pb}/^{204}\text{Pb}$ (17.23–17.25) and $^{207}\text{Pb}/^{204}\text{Pb}$ (15.46–15.48) ratios of olivine-bearing phlogopite lamproite and the phlogopite lamproite (17.13–17.23 and 15.46–15.48) overlap. However, the $^{208}\text{Pb}/^{204}\text{Pb}$ ratios of the former (37.13–37.19) are slightly lower than the latter (37.20 – 37.32). The madupitic lamproites have Pb isotopic ratios ($^{206}\text{Pb}/^{204}\text{Pb}$ 17.4–17.58, $^{207}\text{Pb}/^{204}\text{Pb}$ 15.47–15.50, $^{208}\text{Pb}/^{204}\text{Pb}$ 37.42–37.52) higher than those of the phlogopite lamproites.

4.5 Summary

Based on petrographic studies, the Leucite Hills lamproites can be divided into two main groups: phlogopite lamproites and madupitic lamproites, and two sub-groups:

Table 4.12 The Sr, Nd and Pb isotopic composition of the whole rock unaltered and uncontaminated Leucite Hills lamproites.

Rock type	Sample	$^{87}\text{Sr}/^{86}\text{Sr}$	$^{143}\text{Nd}/^{144}\text{Nd}$	$^{206}\text{Pb}/^{204}\text{Pb}$	$^{207}\text{Pb}/^{204}\text{Pb}$	$^{208}\text{Pb}/^{204}\text{Pb}$
madp lamp	110BGT	0.705451	0.512029	17.457	15.484	37.272
madp lamp	112BGT	0.705380	0.511972	17.446	15.484	37.463
madp lamp	113BGT	0.705343	0.512027	17.436	15.473	37.424
madp lamp	114BGT	0.705512	0.512043	17.442	15.481	37.459
madp lamp	116BGT	0.705370	0.512007	17.444	15.484	37.455
madp lamp	147MTM	0.705514	0.512086	17.534	15.491	37.501
madp lamp	149MTM	0.705512	0.512055	17.535	15.496	37.512
madp lamp	133PLB	0.705488	0.512100	17.556	15.508	37.489
madp lamp	135PLB	0.705511	0.512085	17.542	15.489	37.480
madp lamp	136PLB	0.70545	0.512088	17.547	15.496	37.517
madp lamp	138PLB	0.705561	0.512112	17.563	15.490	37.485
madp lamp	140PLB	0.705631	0.512075	17.583	15.504	37.523
phg lamp	119ZM	0.705741	0.511872	17.227	15.464	37.318
phg lamp	120ZM	0.705659	0.511908	17.182	15.462	37.258
phg lamp	122ZM	0.705683	0.511940	17.220	15.467	37.320
phg lamp	141NTM	0.705909	0.511879	17.273	15.482	37.280
phg lamp	143NTM	0.706025	0.511862	17.281	15.482	37.278
phg lamp	144NTM	0.70606	0.511888	17.282	15.478	37.203
phg lamp	146NTM	0.70603	0.511871	17.239	15.470	37.239
phg lamp	150STM	0.705852	0.511782	17.227	15.470	37.159
phg lamp	151STM	0.705843	0.511830	17.254	15.477	37.188
phg lamp	152STM	0.706079	0.511781	17.250	15.460	37.130
phg lamp	153STM	0.70592	0.511809	17.253	15.459	37.170
phg lamp	154STM	0.705951	0.511806	17.247	15.466	37.158

Sr and Nd isotopic ratios normalized to $^{86}\text{Sr}/^{86}\text{Sr} = 8.31465$ and $^{146}\text{Nd}/^{144}\text{Nd} = 0.7219$, respectively.

NBS 987 Sr standard = 0.710251 ± 0.00002 , La Jolla $^{143}\text{Nd}/^{144}\text{Nd} = 0.511870 \pm 0.00003$.

Uncertainties are given at 2σ level.

An average fractionation correction of 0.1% per mass unit was applied to all measured Pb isotopic ratios based on analysis of NBS981. The reproducibility of Pb isotopic ratios is 0.1% at the 2σ level.

Figure 4.27 $^{143}\text{Nd}/^{144}\text{Nd}$ versus $^{87}\text{Sr}/^{86}\text{Sr}$ diagram. The transitional madupitic (BGT: Badgers Teeth, MTM: Middle Table Mountain) and madupitic lamproites (PLB: Pilot Butte) are characterized by higher $^{143}\text{Nd}/^{144}\text{Nd}$ and lower $^{87}\text{Sr}/^{86}\text{Sr}$ ratios than the phlogopite lamproites (ZM: Zirkel Messa, NTM: North Table Mountain). The olivine-bearing phlogopite lamproites from South Table Mountain (STM) have similar isotopic compositions to those of the phlogopite lamproites.

Figure 4.28 (a) $^{207}\text{Pb}/^{204}\text{Pb}$ versus $^{206}\text{Pb}/^{204}\text{Pb}$ and (b) $^{208}\text{Pb}/^{204}\text{Pb}$ versus $^{206}\text{Pb}/^{204}\text{Pb}$ isotopic diagrams. The madupitic and transitional madupitic lamproites have higher Pb isotopic ratios than the phlogopite lamproites. The olivine-bearing phlogopite lamproite shows similar Pb isotopic composition to those of the phlogopite lamproites. Symbols as in Fig. 4.27.

transitional madupitic lamproites and olivine-bearing phlogopite lamproites. The leaching test and analysis of the sedimentary xenoliths and the separated minerals from lamproites show that the transitional madupitic lamproites resemble the madupitic lamproites in many aspects of their geochemical characteristics. The Pb isotopic ratios of the Boars Tusk and Deer Butte phlogopite lamproites are similar to those of the transitional madupitic/madupitic lamproites but different from those of the other Leucite Hills phlogopite lamproites. This is because the phlogopite lamproites from Boars Tusk and Deer Butte have probably been affected by secondary alteration or crustal assimilation. The olivine-bearing phlogopite lamproites share many geochemical features with the phlogopite lamproites and thus these two lamproite types can be grouped together.

The madupitic lamproites and phlogopite lamproites form two distinct groupings on many major and trace element diagrams. The chemical differences are clearly reflected in the mineralogical variability of the lamproites. This, along with the isotopic variation between the two lamproite types, can also indicate variation in source mineralogy and composition.

CHAPTER 5

SOURCE EVALUATION OF THE LEUCITE HILLS LAMPROITES

5.1. Introduction

In this chapter, the nature of the mantle source of the Leucite Hills lamproites is discussed. What will be considered are the geochemical and mineralogical constraints needed to generate the different lamproite types in the Leucite Hills region. However, it is also important to assess the primary character of the Leucite Hills lamproite melts prior to any discussion regarding the source evaluation because only primitive magmas can preserve the geochemical characteristics of the source from which they are derived.

5.2 Evaluating the primary character of the Leucite Hills lamproite melts

The two most important processes that can influence changes in the chemical composition of the primary magmas are fractional crystallization and crustal contamination. Therefore, the effect of these processes in the petrogenesis of the Leucite Hills lamproites is discussed here.

5.2.1 Evaluation of fractional crystallization

Because the phlogopite lamproites are more Si-rich than the madupitic lamproites (Section 4.4.1), the latter might be considered as parental to the former. However, the observation that the phlogopite lamproites and madupitic lamproites form two distinct groups in the major and trace element diagrams suggest that they cannot be related by fractional crystallization. Although this observation may also indicate that madupitic and phlogopite lamproites originate from the same source but under varying degrees of partial melting, the distinct isotopic compositions among various lamproite types from the Leucite Hills clearly support their derivation from distinct mantle sources.

5.2.2 Evaluation of crustal contamination

Ogden (1978) suggested that the madupitic lamproites from the Pilot Butte locality may have attained their peculiar mineral assemblage and chemical composition

by reaction between a phlogopite lamproite liquid and sedimentary xenoliths. This was the first time that the proposal had been made that the Leucite Hills lamproites were related to one another by contamination. However, this hypothesis was solely based on petrographic observations without any geochemical evidence.

Contamination can occur in two forms. One involves magma mixing between mantle-derived and crustal melts while the second involves assimilation of crustal rocks by mantle-derived melts. Two types of assimilation can occur: 1. bulk assimilation, and 2. assimilation accompanied by fractional crystallization (AFC). Bulk assimilation can be evaluated using a simple binary mixing model. A more realistic model involves the AFC model as formulated by DePaolo (1981). The parameters needed to quantify the AFC model include the composition of the initial magma, the composition of the host rock, the composition of the fractionating phases, the partition coefficients between the fractionating phases and the residual melt, and the rate of assimilation and crystallization. Due to the lack of constraints on many of the parameters, the AFC calculation would be subject to oversimplification and thus it is not performed here.

For bulk assimilation, the formula $X_M = X_A f + X_B(1-f)$ is used to quantitatively model the simple mixing of two components. X_M is the concentration of the element in the contaminated magma, X_A the concentration of the element in the contaminant is and X_B is the concentration of the element in the original magma. f is the weight fraction and is calculated as $f = A/(A+B)$, where A and B are the weight fractions of the contaminant and the original magma, respectively. The isotopic ratios of the contaminated magma (R^X_M) is given as $(R^X_A X_A f + R^X_B X_B(1-f))/(X_A f + X_B(1-f))$ (Faure, 1986). The contaminant materials that are likely to participate in the assimilation process of the Leucite Hills lamproites are the lower crust, upper crust and the sedimentary rocks. The chemical composition of the lower crustal xenoliths (Section 3.4) and the sedimentary xenoliths (section 4.4.3.2) from the Leucite Hills, already reported in Tables 3.4 and 4.9 and 4.10, have been averaged and used for the assimilation calculations. Abundant upper crustal igneous rocks can be found at the Deer Butte but they are extremely altered and not suitable for geochemical analysis. Instead, the upper crust composition compiled by Wedepohl (1995) has been used for the purpose of the binary mixing calculation.

The main purpose of the elemental modeling is to find out whether the crustal

assimilation of one type of lamproite can lead to the formation of another type. The major and trace element contents and the isotopic ratios of the contaminated lamproite melt under varying values of f are then compared with those of the Leucite Hills lamproites to see whether they can reach comparable levels in terms of element abundances and isotopic ratios.

5.2.2.1 Major element calculations

The effects of lower and upper crustal, and sedimentary xenolith contamination on the major element contents of the contaminated madupitic lamproites can be evaluated from Figs. 5.1 – 5.3. In these figures, the major element abundances of the contaminated magmas and the lamproites are normalized to the chemical composition of either madupitic or phlogopite lamproite. Fig. 5.1 shows that bulk assimilation of lower crust by the madupitic lamproites would increase the Al_2O_3 , FeO and Na_2O and decrease the TiO_2 , MgO, K_2O and P_2O_5 contents of the contaminated melts. Assimilation by the phlogopite lamproites of lower crust increases Al_2O_3 , CaO, MnO, FeO_T , Na_2O and decreases the TiO_2 , K_2O and P_2O_5 contents of the contaminated magma. Assimilation of upper crustal rocks by Leucite Hills lamproitic melts would have similar effects to the assimilation of these lamproites by the lower crust but the melt would contain lower MgO, CaO, MnO and FeO_T and higher Na_2O contents (Fig. 5.2). The lamproite melts contaminated by the sedimentary rocks usually have higher CaO, and lower FeO_T , and Na_2O than those contaminated by upper crust. The K_2O content of the lamproite does not change with contamination by the sedimentary rocks (Fig. 5.3).

In all cases, the major element composition of the contaminated lamproite magmas does not approach the chemical composition of any of the existing lamproite types in the Leucite Hills under varying degrees of f . This means that the major element composition of the madupitic lamproites cannot result from the crustal contamination of the phlogopite lamproites and vice versa.

5.2.2.2 Trace element calculations

A similar approach to the one taken above was used to calculate the trace element composition of the contaminated lamproite melt. The trace element concentrations of the

Figure 5.1 (a) Crustal contamination of the madupitic lamproites by lower crust increases the Al_2O_3 , MnO , FeO and Na_2O and decreases the TiO_2 , MgO , K_2O and P_2O_5 contents of the contaminated melts but never reaches those of the phlogopite lamproites. (b) Crustal contamination by the phlogopite lamproites of lower crust has similar effects to those of the madupitic lamproites on the major element composition of the contaminate melt. The fraction of contaminated melt is indicated by f in the legend.

Figure 5.2 (a) Upper crustal contamination of the madupitic lamproites increases the Al_2O_3 , MnO , FeO and Na_2O and decreases the TiO_2 , MgO , K_2O and P_2O_5 contents of the contaminated melts. However, these normalized values are different from those of the phlogopite lamproites. (b) Upper crustal contamination of the phlogopite lamproites has similar effects on the major element composition of the contaminated melt as above (except the MnO and FeO contents). The chemical compositions of the contaminated melts do not resemble those of the madupitic lamproite.

Sedimentary xenoliths

Figure 5.3 (a) Contamination of the madupitic lamproites by sedimentary rocks increases the Al_2O_3 , K_2O and P_2O_5 and decreases the TiO_2 , MgO , and Na_2O contents of the contaminated melts. These oxides never reach the levels shown by the phlogopite lamproites. (b) The contamination of the phlogopite lamproites by the sedimentary rocks has similar effects on the major element composition of the contaminated melt as above. The chemical composition of the contaminated melt does not resemble that of the madupitic lamproite.

contaminated magmas have been normalized to the average trace element concentration of the madupitic or the phlogopite lamproites and plotted on Figs. 5.4 – 6 (the Hf and Zr are not included in the calculations because of the reasons explained in section 4.4.2).

It is seen that the contamination by lower crust decreases the concentration of most of the incompatible elements and increases the concentration of the HREE (Fig. 5.4a). Although unreasonable and high degrees of contamination ($f > 0.5$) of the madupitic lamproite can lower the concentration of some of the incompatible elements to levels similar to those of the phlogopite lamproites, HREE contents of the contaminated melt are still higher than those of the phlogopite lamproites (Fig. 5.4a). Contamination by lower crust of the phlogopite lamproite melt increases the HREE contents and decreases the incompatible elements concentration (Fig. 5.4b). The incompatible element concentration of the resulting contaminated phlogopite lamproite melt is completely different from the madupitic lamproites.

The upper crust, even though characterized by higher concentrations of incompatible elements than lower crustal rocks, in general, has trace element concentrations (except the HREE) that are much lower than the lamproites. Therefore, the effect of the upper crustal contamination is similar to that of contamination by lower crust because the incompatible element contents are lowered and the HREE concentrations of the contaminated melt are similar (Fig. 5.5). Assimilation of the upper crust could also influence the compatible elements of the contaminated lamproite magma dramatically and the contaminated magma is depleted in these elements compared to those of the Leucite Hills lamproites. For example, 10% contamination by the upper crust can decrease the Cr content from 450 ppm in the madupitic lamproite to 80 ppm in the contaminated magma which is much lower than the Cr concentration in the phlogopite lamproites (400 ppm). The contamination by sedimentary xenoliths lowers the incompatible and compatible elements and increases the HREE contents of the contaminated lamproites. None of the chemical compositions of the contaminated magmas is similar to those of the known lamproite types in the Leucite Hills (Fig. 5.6).

5.2.2.3 Isotopic calculations

The isotopic data can also put additional constraints on models proposing crustal

Figure 5.4 (a) With increasing lower crustal contamination of the madupitic lamproites, the LILE - LREE and HREE contents of the contaminated melt changes to higher and lower levels, respectively. The trace element pattern of the phlogopite lamproites is different to that of the contaminated madupitic lamproite. (b) The contamination of phlogopite lamproites by the lower crust decreases the LILE - LREE abundances and increases the HREE of the contaminated melt.

Figure 5.4 (a) With increasing lower crustal contamination of the madupitic lamproites, the LILE - LREE and HREE contents of the contaminated melt changes to higher and lower levels, respectively. The trace element pattern of the phlogopite lamproites is different to that of the contaminated madupitic lamproite. (b) The contamination of phlogopite lamproites by the lower crust decreases the LILE - LREE abundances and increases the HREE of the contaminated melt.

Sedimentary xenoliths

Figure 5.6 (a) Contamination of the madupitic lamproite by sedimentary rocks increases the LILE and LREE contents and decreases the HREE contents of the contaminated melt. K and P contents of such contaminated melt do not vary to great extent. The general trace element patterns of the phlogopite lamproites differ from the contaminated madupitic melts. (b) Contamination of phlogopite lamproites by sedimentary rocks decreases the incompatible elements (except Ta, Sr and Nd) and increases the HREE contents of the contaminated melt. The resulting melt is totally different from the madupitic lamproite in terms of trace element concentrations.

assimilation. Fig. 5.7a shows that the Sr concentration and $^{87}\text{Sr}/^{86}\text{Sr}$ ratio of the lamproite melt contaminated by lower crust plots away from the isotopic compositions of both types of lamproites from the Leucite Hills. Even though the Sr concentration of the madupitic lamproites contaminated by the lower crust decreases to values comparable to those of the phlogopite lamproites, the $^{87}\text{Sr}/^{86}\text{Sr}$ ratios of the contaminated madupitic lamproites decrease from 0.7055 ($f = 0.1$) to 0.7053 ($f = 0.5$) which are different from the $^{87}\text{Sr}/^{86}\text{Sr}$ ratio of the phlogopite lamproite (0.7058). Similar degrees of contamination ($f = 0.1 - 0.5$) of a phlogopite lamproite using lower crust contaminant, lowers the Sr concentration as well as the $^{87}\text{Sr}/^{86}\text{Sr}$ ratios of the contaminated melt which are different to those from the madupitic lamproites.

Sr isotopic modeling shows that the contamination by upper crust (Fig. 5.7b) is able to lower the $^{87}\text{Sr}/^{86}\text{Sr}$ ratios of the madupitic lamproites to values comparable to those of the phlogopite lamproites but only for f values > 0.5 . In other words, if the madupitic lamproites were to assimilate the upper crustal materials in equal amounts or greater than 50%, then the contaminated magma would have a Sr isotopic composition similar to that of the phlogopite lamproites. The contamination of the Leucite Hills lamproites by the sedimentary xenoliths increases the Sr isotopic ratios of the contaminated melt (Fig. 5.7c).

The effect of lower crustal contamination on the Nd content and Nd isotopic ratios (Fig. 5.8a) is to lower the $^{143}\text{Nd}/^{144}\text{Nd}$ ratio and Nd concentration from 0.51204 and 260 ppm in the madupitic lamproites to 0.51197 and 145 ppm in the contaminated magma ($f = 0.5$). The Nd content and $^{143}\text{Nd}/^{144}\text{Nd}$ ratio of such a contaminated magma at f value of 0.5 is comparable to that of the phlogopite lamproite (Nd = 144 ppm, $^{143}\text{Nd}/^{144}\text{Nd} = 0.51197$). Lower crustal contamination of the phlogopite lamproite would decrease the Nd isotopic ratio and Nd concentrations in the contaminated melt to levels that are very different from the madupitic lamproites (Fig. 5.8a). In the case of contamination by the upper crust, the madupitic melt needs to assimilate moderate amounts of upper crustal rocks ($f > 0.7$) in order for its $^{143}\text{Nd}/^{144}\text{Nd}$ ratio to reach values close to of those of the phlogopite lamproites (0.5118) (Fig. 5.8b). However, even though contamination of the madupitic lamproite may generate a melt with Nd isotopic composition similar to those of the phlogopite lamproites, the major, trace element and Sr

isotopes of such melt are still very different. Assimilation of upper crust by a melt with the composition of a phlogopite lamproite cannot generate a melt with Nd isotopic ratios similar to the madupitic lamproite. The assimilation of the sedimentary xenoliths (Fig. 5.8c), on the other hand, would lower the Nd isotopic ratios of those of the phlogopite lamproites to values comparable to those of the madupitic lamproites. However, the phlogopite lamproites must digest extensive quantities of sedimentary material (> 70%) before its Nd isotope ratios reach values comparable to those of the madupitic lamproites.

The Pb isotopic compositions of the upper crustal rocks from the Wyoming Province are variable (Kolker et al., 1991) with the result that any modeling using estimates for upper crust can generate a large and complicated database. However, the Pb isotopic ratios and Pb concentrations of the lower crust and sedimentary xenoliths probably cover the extreme Pb isotopic ratios of the upper crustal rocks and thus a discussion on the lower crust and sedimentary assimilation should suffice for evaluating upper crust involvement.

Very small degrees of contamination by the lower crust (<10%) can lower the $^{206}\text{Pb}/^{204}\text{Pb}$ ratios of the madupitic lamproites down to ratios comparable to those of the phlogopite lamproites (Fig. 5.9a), but the Pb concentration of the contaminated melt is still higher than that of the phlogopite lamproites. Contamination of the phlogopite lamproite melt by lower crustal materials produces Pb isotopic ratios that are similar to those of the madupitic lamproites but the Pb concentration of the resulting melt is lower than that found in the latter.

The sedimentary xenoliths can increase the $^{206}\text{Pb}/^{204}\text{Pb}$ isotopic ratios of phlogopite lamproites to levels comparable to those of the madupitic lamproites (Fig. 5.9b). For example, assimilation of sedimentary xenoliths by a phlogopite lamproite melt at $f = 0.4$ can change the $^{206}\text{Pb}/^{204}\text{Pb}$ ratios of the phlogopite lamproites from 17.39 to 17.53 which is very similar to that of the madupitic lamproites. However, the Sr and Nd isotopic ratios of such a contaminated melt are completely different from those of the madupitic lamproites. The assimilation of the sedimentary xenoliths by melts similar in composition to the madupitic lamproites would increase the Pb isotopic ratios of the melt to much higher values than those of the phlogopite lamproites.

Figure 5.7 The progressive contamination of madupitic and phlogopite lamproites by lower crust (a), upper crust (b) and sedimentary rocks (c) generates contaminated melts with the Sr isotopic ratios and concentrations that differ significantly from the known lamproite types in the Leucite Hills. Numbers indicate the fraction of the original and uncontaminated melt.

Figure 5.8 The progressive contamination of madupitic and phlogopite lamproites by lower crust (a), upper crust (b), and sedimentary rocks (c), generates melts that have Nd isotopic ratios and Nd concentrations significantly different from the known lamproite types in the Leucite Hills. The only exception is contamination of the madupitic lamproites by upper crust in which the Nd isotopic ratios and concentration of the contaminated melt reach those of the phlogopite lamproites at high degrees of assimilation. However, such high degrees of assimilation ($f < 0.6$) are unlikely to occur.

Figure 5.9 The progressive contamination of madupitic and phlogopite lamproites by lower crust (a), and sedimentary rocks (b), generates contaminated melts with the Pb isotopic ratios and concentrations that differ significantly from the known lamproite types in the Leucite Hills.

A combination of major, trace element and isotope ratios modeling shows that the two major lamproite types in the Leucite Hills (i.e., phlogopite lamproite and madupitic lamproite) cannot be related to each other, either by fractional crystallization or by crustal contamination. Therefore, it is more likely that these lamproites represent primitive magmas derived from a heterogeneous source.

5.3 The chemical composition and mineralogy of the source of the Leucite Hills lamproites

Phase equilibrium studies suggest a variety of sources for lamproites, ranging from phlogopite lherzolite to apatite-amphibole-phlogopite veins. It has still not been shown whether these sources could closely or even remotely generate melts with trace element contents and isotopic ratios similar to those of the lamproites. It is clear, however, that the variations in the source mineralogy combined with variables such as the degree of partial melting, fO_2 and volatile contents can generate a variety of lamproite magma types. Obtaining information on such variables requires some knowledge about the nature of the source and an approximation on the modal abundances and chemical compositions of the mineral phases likely to be present in the mantle source. The most common and widespread approaches employed to infer the source compositions of lamproites are: (1) mantle xenoliths, (2) high P – T experiments and (3) evaluation of the geochemical characteristics of the whole rock data. Each of these will be discussed in turn.

5.3.1 Mantle xenoliths

One way to identify the mantle sources of volcanic rocks is to evaluate the mantle xenoliths brought up to the surface by volcanic eruptions. Unfortunately, mantle xenoliths are exceptionally rare in lamproites (Mitchell and Bergman, 1991). The rarity is presumably related to the physical and mechanical properties of lamproite magmas and their inability to carry the xenoliths to the surface. Mantle xenoliths so far recovered from lamproites are dominantly dunite, garnet harzburgite and garnet lherzolite (Atkinson et al., 1984). The chemical composition of these xenoliths, along with the nature of the

minerals recovered from the heavy mineral concentrates from lamproites, point toward the dominant presence of a refractory mantle source.

An alternative way to evaluate the mantle source of lamproites is to investigate xenoliths from other igneous rocks with geochemical affinities to lamproites, such as Group II kimberlites. Mantle xenoliths from Group II kimberlites can range from peridotite-pyroxenite, metasomatised peridotites rich in amphibole and/or mica and glimmerites (Dawson, 1980; Mitchell, 1995b). The peridotite-pyroxenite group consists of dunite, harzburgite, lherzolite, websterite and pyroxenite. The nomenclature is based on the IUGS recommendation with the exception that modal clinopyroxene of at least 5% suffices to name the rock as lherzolite. The metasomatised peridotites are composed of metasomatic minerals including mica, amphibole and apatite. Other, less common, metasomatic mineral phases such as ilmenite, rutile, sulphides and carbonate have also been reported. The metasomatic phases are either disseminated throughout the xenolith with modal abundances between trace amounts to 5% (modally metasomatised xenoliths) or they have developed into distinct layers (veined xenoliths). Textural, mineralogical and geochemical criteria indicate that the metasomatic phases formed at the same temperature and pressure as the xenoliths and are thus unrelated to contamination by the host magma (Erlank and Rickard, 1977; Gurney and Harte, 1980). Moreover, the mineral chemistry shows that the metasomatic minerals are more enriched in the incompatible elements than in the host rock and partial melting of such phases can generate magmas enriched in incompatible trace elements.

5.3.2 Phase equilibrium studies

High P-T experiments on lamproites have been conducted by several workers to assess the composition of the source region, to examine the composition of extracted melts under varying P and T, and to evaluate the effect of volatiles on the source and partial melt compositions. Experimental studies on ultrapotassic rocks usually fall into the three following categories: (a) melting experiments on natural and synthetic lamproites, (b) melting experiments on mantle xenoliths or synthesized materials resembling the mantle sources, and (c) experimental studies on the stability fields of the characteristic mantle minerals.

5.3.2.a Melting experiments on the natural and synthetic lamproites

In high P-T experiments, a vital prerequisite in evaluating the source composition of lamproites is the selection of what is considered to be the primary lamproite magma. This is controversial among many petrologists and a number of 'primary' lamproite magmas have been proposed. These include madupitic lamproite (Barton and Hamilton, 1979), transitional madupitic lamproite (Arima and Edgar, 1983; Mitchell and Edgar, in press), leucite phlogopite lamproite (Foley, 1989; Mitchell and Edgar, 1989), sanidine phlogopite lamproite (Barton and Hamilton, 1982; Mitchell, 1995a; Mitchell and Edgar, 1989), olivine lamproite (Foley, 1993) and phlogopite armalcolite lamproite (Edgar and Mitchell, 1997).

The selection of the starting material can influence the results of the experiments and accordingly the inferred mineralogy of the source. The experimental studies, so far, have failed to agree on the precise nature of the mantle source of lamproites. Other factors such as the lack of control over oxidation state during the experimental runs (e.g., Barton and Hamilton, 1979) could also produce problems in source evaluation. For example, Fe loss could lead to higher Mg# in the crystallized phases. Also of concern is that some experiments were undertaken at pressures < 40 kb, corresponding to depths lower than those estimated for the generation of lamproites (e.g., Barton and Hamilton, 1979). Finally, many of the experiments refer to the near-liquidus phase assemblages of lamproitic melt to delineate the mineralogy of the source. This is based on the assumption that the near-liquidus phases of a primary melt at high pressure denote mineral assemblages that are in equilibrium with the mantle and melt at pressures corresponding to the pressures that existed near the source. As a result, the near-liquidus mineral assemblages have been used to deduce the mineralogy of the mantle source at certain pressures, temperatures and fO_2 . This was considered as a rational argument among petrologists studying lamproites until Foley (1992) introduced his vein-plus-wall-rock melting model in which the near-solidus phases of the primary melt at high pressures supposedly reflect the mineralogy of the small volume partial melt from the veined mantle. The near-liquidus represents the residual source in equilibrium with the melt at the time of separation. Mitchell (1995a), Edgar and Mitchell (1997), and Mitchell and Edgar (in press) adopted the same approach to deduce the mineralogy of the mantle

source of lamproites from the Leucite Hills and Smoky Butte, respectively. They also concluded that the partial melts from the veins should be mixed with those from the peridotite wall rock to give rise to a magma with major element composition similar to the starting materials. In the melting experiments of Mitchell (1995a), Edgar and Mitchell (1997), and Mitchell and Edgar (in press) some new and uncommon minerals, including two varieties of K-Ti silicate and a K-Ba phosphate, were observed as near-solidus phases at high pressure. With the exception of the K-Ti silicates that have been reported from lamproites, none of the mentioned minerals has been found in mantle xenoliths from lamproites. As Edgar and Mitchell (1997) state “the near-solidus source for the Smoky Butte and Leucite Hills lamproites suggested by the high P-T experiments need not contain all the ‘exotic’ minerals present in the experiment”. It was also impossible for these authors to recalculate the K-Ti and Ba-rich silicate phases into conventional lower pressure mineral formulas.

Table 5.1 summarizes the available data on high P-T experiments on lamproites. Except for Mitchell (1995a), Edgar and Mitchell (1997), and Mitchell and Edgar (in press) the other papers quoted only the near-liquidus mineral assemblages as the source mineralogy of lamproites. Based on the careful examination of the published phase diagrams, the near-solidus minerals are added to Table 5.1 in order to infer sources that are more consistent with the vein-plus-wall rock melting model of Foley (1992). The near-liquidus phase assemblage introduces four different residual sources remaining after lamproite extraction: phlogopite clinopyroxenite, garnet clinopyroxenite, garnet lherzolite and phlogopite harzburgite. The results of the experiments suggest a clinopyroxenite/pyroxenite/lherzolite source for phlogopite lamproites, and a harzburgite source for the madupitic lamproites.

Based on the near-solidus phase assemblage (Table 5.1), the vein assemblages can include: phlogopite $\pm$ clinopyroxene $\pm$ apatite $\pm$ richterite $\pm$ garnet $\pm$ rutile $\pm$ K-Ti silicate $\pm$ K-Ba phosphate. The wide variations of the inferred veined-mantle sources are essentially due to the differences in the selection of the starting material and the experimental conditions (e.g., volatile type) under which the experiments were run.

In addition to high P-T experiments on the natural samples, studies of the

Table 5.1 Summary of the high P-T experimental studies on lamproites.

Starting Material	Sanidine lamproite		
Locality	Leucite Hills		
Reference	Barton and Hamilton (1982)		
Volatiles	Not Added		
Pressure (kb)	(5 - 11)	(11 - 27)	(27 - >30)
Near liquidus phases	Ol	Opx+Ol	Cpx+Gnt
Inferred Source	Garnet lherzolite		
Pressure (kb)	(5 - 14)	(14 - >30)	
Near Solidus phases	Phg+cpx+San	Phg+cpx+Gnt+San	
Inferred Source	Phlogopite clinopyroxene vein		

Starting Material	Sanidine lamproite		
Locality	Leucite Hills		
Reference	Mitchell (1995)		
Volatiles	Not Added		
Pressure (kb)	(40 - 45)		(45 - > 70)
Near liquidus phases	Cpx		Cpx+Gnt
Inferred Source	Garnet clinopyroxenite		
Pressure (kb)	(40 - 55)		(55 - >70)

Table 5.1 (continued)

Starting Material	Sanidine richterite lamproite
Locality	Murcia-Almeria (Spain)
Reference	Mitchell and Edgar (1998)
Volatiles	Not Added
Pressure (kb)	(40 - 60) (40 - >55)
Near liquidus phases	Gnt+Opx+Cpx Gnt+Cpx
Inferred Source	Pyroxenite
Pressure (kb)	(40 - 45) (45 - 65)
Near Solidus phases	Gnt+Opx+Cpx+Coes+Phg Gnt+Cpx+Coes+K-Ti silicate
Inferred Source	Phlogopite clinopyroxene K-Ti silicate vein

Starting Material	Leucite lamproite
Locality	Gaussberg
Reference	Foley S. F. (1989)
Volatiles	H ₂ O>CH ₄
Pressure (kb)	(5 - 8) (8 - 21)
Near liquidus phases	Ol Phg
Inferred Source	Phlogopite harzburgite
Pressure (kb)	(5 - 20?) (20? - >30)
Near Solidus phases	Phg+cpx Phg+Rut
Inferred Source	Phlogopite clinopyroxene rutile vein

Table 5.1 (continued)

Starting Material	Leucite lamproite
Locality	W. Kimberly
Reference	Mitchell and Edgar (in press)
Volatiles	Not Added
Pressure (kb)	(45 - 60)
Near liquidus phases	Opx (60 - 65) Gnt+Rut+Coos+Cpx
Inferred Source	Pyroxenite
Pressure (kb)	(45 - 55)
Near Solidus phases	Cpx+Rut+Coos+Phg+Ap Gnt+Ru+Coes+Cpx+Phg+Ap
Inferred Source	Phlogopite clinopyroxene apatite rutile vein

Starting Material	amalcotite-phg lamproite
Locality	Smokey Butte
Reference	Edgar and Mitchell (1997)
Volatiles	
Pressure (GPa)	(1 - 2.5)
Near liquidus phases	Ol (2.5 - 4) Cpx (5 - 12) Gnt (>12)
Inferred Source	Garnet clinopyroxenite
Pressure (GPa)	(1 - 4)
Near Solidus phases	Cpx+Opx+Phg+Anp Gnt+Cpx+Phg+Rut+Coes+K-Ti Sil
Inferred Source	Phlogopite clinopyroxene K-Ti silicate vein

Table 5.1 (continued)

Starting Material	Madupitic lamproite
Locality	Leucite Hills
Reference	Barton and Hamilton (1979)
Volatiles	Not Added
Pressure (kb)	(3 - 7) (7-18) (18 - >30)
Near liquidus phases	Cpx+Ol Cpx+Sp Cpx
Inferred Source	Phlogopite clinopyroxenite
Pressure (kb)	(4 - 12) (12 - 16) (16 - >30)
Near Solidus phases	Cpx+Phg+Apt Cpx+Phg+Sp+Apt Cpx+Phg+Grt+Apt
Inferred Source	Phlogopite clinopyroxene apatite vein

Starting Material	Transitional madupitic lamproite
Locality	W. Kimberly
Reference	Arima and Edgar (1983)
Volatiles	Not Added
Pressure (kb)	(10 - 15) (15 - >30)
Near liquidus phases	Ol Opx
Inferred Source	Phlogopite harzburgite
Pressure (kb)	(13 - >30)
Near Solidus phases	Opx+Phg
Inferred Source	Phlogopite vein

Table 5.1 (continued)

Starting Material	Transitional madupitic lamproite
Locality	W. Kimberly
Reference	Arima and Edgar (1983)
Volatiles	H ₂ O (13%)
Pressure (kb)	(10 - 23) (23 - >40)
Near liquidus phases	Ol+Opx Opx
Inferred Source	Phlogopite harzburgite
Pressure (kb)	(15 - >30)
Near Solidus phases	Opx+Cpx+Phg+Rut
Inferred Source	Phlogopite clinopyroxene rutile vein

Starting Material	Transitional madupitic lamproite
Locality	Leucite Hills
Reference	Mitchell and Edgar (in press)
Volatiles	Not Added
Pressure (kb)	(30 - 65)
Near liquidus phases	Cpx
Inferred Source	Clinopyroxenite
Pressure (kb)	(30 - 55) (55 - 65)
Near Solidus phases	Cpx+Phg+Ap+Titnt Cpx+Phg+Ap+K-Ti silicate
Inferred Source	Phlogopite clinopyroxene apatite K-Ti silicate vein

Table 5.1 (continued)

Starting Material	Olivine lamproite
Locality	W. Kimberly
Reference	Foley S. F. (1992)
Volatiles	H ₂ O>CH ₄
Pressure (kb)	(20 - 40)
Near liquidus phases	Ol (40 - >55) Ol+Opx+Phg
Inferred Source	Phlogopite harzburgite
Pressure (kb)	(20 - 32?) (32? - 55)
Near Solidus phases	Phg+Ol+Cpx Phg+Opx+Rut
Inferred Source	Phlogopite clinopyroxene rutile vein

anhydrous and volatile-bearing (H_2O , CO_2 , F) varieties of the synthetic systems Ne-Fo-Qtz and Ks-Fo-Qtz at high pressures have been widely used to examine the melting relations in the mantle origin of alkaline igneous rocks and the effect of pressure and volatile components on the chemical composition of the melts (Kushiro, 1980; Wendlandt and Eggler, 1980a, b; Sekine and Wyllie, 1982; Gupta and Green, 1988; Foley, 1989). The system Ks-Fo-Qtz is particularly important in evaluating lamproite petrogenesis because it involves high concentrations of K_2O . The Ks-Fo-Qtz system also involves key minerals such as phlogopite, olivine, orthopyroxene, leucite and sanidine that are observed in both the mode and norm of the many lamproites. The main drawback of this system is the lack of normative compositions of some minerals that are common in lamproites such as diopside. Still, a brief review of this system is presented here since it has been shown to have important bearings on the origin of the Leucite Hills lamproites.

Kuehner (1980) was the first to apply the Ks-Fo-Qtz system, although anhydrous, to the petrogenesis of lamproites by plotting the chemical composition of madupitic and phlogopite lamproites from Leucite Hills on the Fo-Ks-Qtz triangle (Fig. 5.10). In this diagram, madupitic lamproites plot between the kalsilite, leucite and forsterite stability fields whereas phlogopite lamproites reside within the forsterite, sanidine and enstatite stability fields. At high pressure (34 kb) a peritectic point resides between Fo-En-San field and within the Fo stability field. Melting of a mantle source at high pressure starts at this peritectic point that marks a melt low in K_2O , SiO_2 and Al_2O_3 and high in MgO contents and with a chemical composition similar to that of the madupitic lamproites. At lower pressure (19 kb), the peritectic point shifts towards the En side of the diagram and the stability field of leucite decreases and that of sanidine increases. At lower pressure, the invariant point thus resides within the Fo-San-En subsolidus compositional triangle, where the normative composition of the majority of phlogopite lamproites plots. Partial melts produced at low pressure (19 kb) from the K_2O -bearing harzburgite mantle source is thus higher in K_2O , SiO_2 and Al_2O_3 and lower in MgO content than the melt generated at higher pressure (34 kb) and thus have a chemical composition similar to that of the phlogopite lamproites. From the preceding discussion, Kuehner (1980) and Kuehner et al. (1981) concluded that madupitic lamproites are generated from partial melting of a peridotite mantle source at great depths (~ 90 km) whereas phlogopite lamproites are

generated from the same source at shallower levels (~ 65 km). As Mitchell and Bergman (1991) correctly point out, the main concern regarding the application of the anhydrous Fo-Ks-Qtz system to lamproites genesis is the absence of phlogopite in the system. The volatile contents of the ultrapotassic rocks are normally higher than most other alkaline igneous rocks and thus the volatile-bearing systems are more appropriate to the study of lamproites and kimberlites.

Foley et al. (1986) have discussed, in detail, the control of F-O-H-C vapor system on the Fo-Ks-Qtz system and its application to the origin of lamproites. Their experiment was prompted by the fact that the variation in the chemical composition of lamproite melts reflects not only changes in the degree of partial melting, source mineralogy and composition but also variations in the fO_2 and volatile composition and abundances. CO_2 seems to be more abundant in the source region of kimberlite, whereas H_2O is considered the dominant volatile in the source of lamproites. The high abundance of F in the Leucite Hills lamproites (~ 0.8 wt%) and the positive correlation between F and K in lamproites show that F component (e.g., HF) could also be an important volatile in the origin of lamproites (Edgar, 1979). In many mantle conditions and particularly those prevailing in the source of lamproites, H_2O is present in excess of F and thus HF is formed to counterbalance the effect of F. It can be concluded that the source region of the silica-rich phlogopite lamproites from the Leucite Hills is probably richer in H_2O and HF (i.e., more reduced) than the source areas of the silica-poor lamproites such as madupitic lamproites. In other words, the addition of F and CO_2 to the source region of the Leucite Hills lamproites can change the composition of the resulting melts towards a madupitic composition (i.e., lower SiO_2 , K_2O and higher MgO), whereas H_2O , HF and CH_4 can produce melts with chemical compositions similar to the phlogopite lamproites (i.e., higher SiO_2 , K_2O and lower MgO). It should be noted that the low abundance of F in mineral constituents of some mantle xenoliths such as phlogopite, apatite and K-richterite (e.g., southwest Uganda; Edgar and Pizzolato, 1995) may imply that the F enrichment in the potassic melts such as lamproites is not due to the source enrichment in F. Thus the F content of the mantle source of lamproites is probably lower than that estimated by Foley et al. (1986). More importantly, the close examination of Fo-Ks-Qtz system shows that the position of the phlogopite-forsterite-enstatite-liquid peritectic point is more drastically

Figure 5.10 The position of madupitic lamproites and phlogopite lamproites on the 3 kb phase field of the system $\text{KAlSiO}_4\text{-Mg}_2\text{SiO}_4\text{-SiO}_2\text{-H}_2\text{O}$ (Luth, 1967). Field of lamproites from this work and after Kuehner (1980).

Figure 5.11 System Fo-Ks-Qtz showing the effects of pressure (a) and volatile (b) on the composition of lamproite melt. Melts formed at higher pressures are less silica-rich relative to those generated at lower pressures. (After Foley, 1992).

affected by changes in their pressure rather than variations in the volatile components (Fig. 5.11a, b). Foley (1993) has emphasized that the variations in the depths of formation could easily explain the diversity of chemical compositions among different lamproites, and that localized variations in the volatile types seem improbable. Foley et al. (1986) also propose that the olivine lamproites from the West Kimberley originated from a reduced source and at greater depths than the more SiO_2 -rich, MgO-poor lamproites. The high-pressure experiments using the West Kimberley olivine and leucite lamproites (Foley, 1989), indicates that both lamproite types originate from similar sources but at different depths. This also explains the presence of diamond in the olivine lamproites from West Kimberley.

Based on the above observations, Foley (1993) proposes that the phlogopite lamproites from Leucite Hills are the most extreme, low-pressure natural examples of lamproites because their high SiO_2 (Quartz normative) compositions correspond to a depth of origin around 10 – 15 kb. By extending the results of Foley's experiments (1989) and observation (1993) to the Leucite Hills lamproites, it can be concluded that the silica-undersaturated madupitic lamproites, if generated from the deeper parts of the mantle than the phlogopite lamproites, are more likely to contain diamond than the other lamproites types. The Sr, Nd and Pb isotopic data of this study (Table 4.12) also support such a hypothesis as the isotopic composition of madupitic lamproites rather than the phlogopite lamproites show proximity to the data from the Group II kimberlites (Mirnejad and Bell, 1999). The important conclusion drawn from studying the Ks-Fo-Qtz system is that depleted peridotite dominates the mantle source that generated the ultrapotassic melts. The results also indicate that processes other than fractional crystallization are involved in generating the different lamproite types. However, confining various lamproites to a common source that is able to generate a variety of melts under varying degrees of partial melting, volatile compositions and P – T conditions could be difficult to justify due to the variations in the isotopic compositions of lamproites (Section 4.4.4). The petrogenesis of lamproites should be examined by combining major and trace elements, isotopic ratios and high P – T experimental studies.

5.3.2.b Experiments on the natural mantle samples

Partial melting experiments on the natural mantle materials have focused mainly on lherzolite xenoliths (e.g., Gudfinsson and Presnall, 2000; and reference therein) that are not directly related to the petrogenesis of ultrapotassic rocks and lamproites. The only high P-T experiment on a mantle xenolith that is relevant to the study of ultrapotassic rocks is that of Lloyd et al. (1987) who used a mica-clinopyroxenite from southwest Uganda. Previously Lloyd and Bailey (1975), in a detailed study of numerous metasomatised alkali clinopyroxenite xenoliths from southwest Uganda, showed that mantle sources other than lherzolite or harzburgite could also generate the K-rich lavas at this region. Lloyd et al. (1987) used a mica clinopyroxenite, containing minor titanomagnetite, titanite and apatite in runs at 20 and 30 kb and 1175 and 1300 °C for their melting experiments. The results of their experiments show that at high degrees of partial melting (20 – 30%), the composition of the melt is comparable to that of the high-potassic mafic lavas. Even though the composition of the melt is not lamproitic, the experiments of Lloyd et al. (op.cit.) have two important findings that bear on the origin of lamproites. First, they show that a mantle source containing mica and clinopyroxene is able to produce high potassic mafic melts. Second, the experiments provide important information on the melting relationship among phlogopite, apatite, ilmenite and clinopyroxene that is useful in the geochemical modeling of the mantle source. It shows that at 20 to 30 kb pressure, apatite is consumed at low degrees of partial melting, followed by ilmenite, phlogopite and clinopyroxene.

High P-T experiments on synthetic mantle assemblages are also lacking but the peritectic point olivine + orthopyroxene + phlogopite + liquid in the synthesized Fo-Ks-Qtz system can represent the partial melts of phlogopite harzburgite if excess H₂O is used in the system (Foley, 1989; 1993). The petrogenetic significance of Fo-Ks-Qtz system has already been discussed. It should be remembered that the addition of H₂O to the system expands the stability field of phlogopite (Foley, 1990) which ensures that phlogopite will stay on the liquidus during most of the melting.

5.3.2.c Stability of mantle minerals

The main purpose of the experimental studies on mantle minerals is to evaluate the stabilities of mineral phases at varying P–T and volatile components in the mantle and to find out to what extent the chemistry of the mantle minerals and thus the resulting melts are affected by changing conditions in the mantle. Of the various metasomatic minerals assumed to be present in the mantle, phlogopite, apatite and amphibole have the tendency to incorporate trace elements in their crystal structure and have been proven to be important mineral constituents in the source of lamproites. These minerals can be completely or partly melted out of the source and contribute to the trace element budget of the partial melts. Thus, both the existence (stability) and persistence (solubility) of the metasomatic mantle minerals should be evaluated prior to any petrogenetic modeling.

Phlogopite

The high modal abundance of phlogopite in lamproites, the high K₂O content (>10 wt%) of phlogopite, and the presence of phlogopite in the near-liquidus and near-solidus of high P – T experiments suggest that phlogopite may be an important mineral phase in the source of lamproite magmas. The stability of phlogopite under upper mantle conditions has been experimentally investigated by a number of workers (e.g., Kushiro et al., 1967; Sato et al., 1997) and the results show that phlogopite can be stable in the subcratonic lithosphere down to 200 km. This is mainly due to the low geotherm of subcratonic lithosphere. In the oceanic upper mantle, which has a higher geothermal gradient than that of the subcontinental mantle, the stability of phlogopite is limited to lower depths.

Phlogopite in the mantle can form by reaction between the K-rich metasomatic agents and garnet or other mantle minerals. Metasomatic agents can be generated either by the release of hydrous melts from subducted slabs or the partial melting of K-bearing peridotites in an uprising plume. The released metasomatic agents can thus react with the overlying harzburgite or lherzolite to form metasomatised peridotites such as phlogopite orthopyroxenite, garnet phlogopite websterite or garnet phlogopite harzburgite (Sato et al., 1997). Such metasomatised phlogopite-bearing peridotites can easily melt out of the sub-continental lithosphere and therefore are potential sources for potassic magmas such

as lamproites. F and CO₂ increase and decrease, respectively, the stability of phlogopite whereas pressure expands the stability field of phlogopite. Moderisk and Boetcher (1972, 1973) evaluated the stability field of phlogopite up to 3 GPa (30 kb) pressure and under various volatile conditions. At pressures greater than 0.2 GPa (20 kb), phlogopite breaks down to forsterite and liquid rich in K₂O. Foley (1990) examined the stability and chemistry of a phlogopite from the West Kimberley and Gaussberg lamproites under varying aH₂O, fO₂, P and T and showed that the chemistry of phlogopite was slightly affected by changing P and T at fixed volatile compositions. Conversely, the chemistry of phlogopite changes significantly at constant P – T and varying volatile compositions. The results of the experimental studies on phlogopite thus suggest that phlogopite melting in the mantle source can contribute significantly to K₂O enrichment during partial melting.

Apatite

Fluorine-rich apatite has been reported frequently from mantle xenoliths and the experimental studies (Watson, 1980) and thermodynamic calculations (Beswick and Carmichael, 1978) show that it can be stable at P-T conditions pertinent to mantle conditions. Apatite, along with phlogopite and amphibole, is considered a late stage metasomatic mineral phase crystallizing from volatile-rich fluids or melts in the form of either disseminated grains or layered veins.

Amphibole

Experimental studies have constrained the stability of amphibole to shallow depths (<70 km), whereas phlogopite is expected to be stable up to 200 km (Arai, 1986). Recent experiment by Konzett et al. (1997), however, suggests that K-richterite, in association with clinopyroxene and phlogopite, can be stable up to ~ 250 km at T < 1400 °C. In addition, K-richterite along with clinopyroxene, olivine, rutile, phlogopite, apatite and garnet have been reported as inclusions in diamonds (Leung et al.; in Konzette et al., 1997). Amphibole is normally considered to be the main phase in generating low K/Na to medium K/Na magmas (Fraser, 1987) Accordingly, amphibole should not be considered as an important mineral constituent of the mantle source of lamproites.

5.3.3 Constraints on mantle source compositions imposed by the geochemical characteristics of whole rocks

The chemical compositions of the primary magmas carry critical information about the geochemical compositions of their source regions. The general petrographical and geochemical features of lamproites such as the presence of diamond and olivine xenocrysts, peridotite xenoliths and high Mg# and Ni and Cr contents are consistent with a mantle origin. Nevertheless, lamproites also contain the highest concentrations of incompatible elements among the alkaline igneous rocks. Generation of such extreme trace element enrichments by partial melting of a primitive or depleted mantle would require unrealistically small degrees of partial melting. This clearly rules out an unmodified or primitive mantle as the potential source and indicates that a major metasomatic activity must have occurred in the source of the lamproites.

It is known that each potential mantle phase has at least one element which is compatible with that phase alone, and thus the trace element variations of primary melts can be used to help specify the mineralogy of mantle sources. Accordingly, the K, Rb, Ba, LREE, Sr and P enrichment of the lamproites reflects the abundant presence of high K- and Ca-bearing metasomatic minerals such as clinopyroxene, apatite, amphibole and phlogopite in the mantle source. All of these minerals have been shown to be stable in upper mantle peridotites and have been reported from the metasomatised mantle xenoliths. The question, however, remains as to what extent these phases have affected the isotopic and trace element variations among different lamproite types in the Leucite Hills. The higher distribution coefficient of Sr and LREE in apatite relative to clinopyroxene indicates that the former mineral is more influential in controlling the Sr and LREE contents of the lamproites. In addition, the relatively high P and Ca contents of the lamproites can indicate the presence of apatite in the source. Similarly, the distribution coefficients of Ba and Rb are higher in phlogopite than amphibole and the high concentrations of these elements in the Leucite Hills lamproites reflect the importance of phlogopite rather than amphibole in the source.

Lloyd and Bailey (1975) attribute the scarcity of the ultrapotassic rocks in oceanic settings to the high geothermal gradient because it will limit the stability of the phlogopite without amphibole being present. On the other hand, due to the low

geothermal gradient in the continental regions, phlogopite can be stable to greater depths than the stability field of amphibole and thus high K/Na magmas are more common. It should be emphasized that phlogopite and apatite are considered here as the major metasomatic minerals that store high concentrations of K, Rb, P and REE. Until further data are available to confirm the presence of other Rb and REE-bearing phases in the source, it seems justified to assume apatite, phlogopite and some K-Ti silicate phases are the major metasomatic mineral phases residing in the source of lamproites. As such, these minerals can largely influence the isotopic composition and trace element content of the source mantle and the derivative lamproite magmas.

5.4 Relationships among the olivine-bearing phlogopite lamproites, madupitic lamproites and phlogopite lamproites

In the studies of lamproites, olivine lamproites and madupitic lamproites are usually grouped together and are considered to be distinct from the phlogopite lamproites (e.g., Mitchell and Bergman, 1991). The main reason for this grouping appears to be related to the similarity in the major and trace element compositions of olivine lamproites and madupitic lamproites. There have been, however, arguments among petrologists about the origin of different lamproite types and particularly the primary character of the olivine and madupitic lamproites. The debate over the origin of olivine lamproites and madupitic lamproites was initiated by Jaques et al. (1986) who proposed that phlogopite lamproites are not primary melts but are derived from olivine lamproite magmas by fractional crystallization. Foley et al. (1986) and Foley (1993), however, suggested that under varying pressures in a metasomatised phlogopite harzburgite mantle source, several primary magmas ranging from phlogopite lamproite to olivine lamproite could be generated. Contrary to both hypotheses, Mitchell and Bergman (1991) and Mitchell (1995a) proposed that: (1) a silica-rich lamproite melt (phlogopite lamproite) cannot be derived from a silica-poor lamproite magma (olivine lamproite and madupitic lamproite), (2) olivine lamproites do not have the geochemical characteristics of a primary magma, rather they are fractionated magmas that have been hybridized with mantle materials, and (3) madupitic lamproites are not primary melts but they are the result of fractionation and contamination of a primitive phlogopite lamproite within the mantle. They also add that

the ultrabasic character of the madupitic lamproites results from high abundances of phlogopite.

The three points by Mitchell and Bergman (1991) and Mitchell (1995a), however, are open to discussion, and each is discussed in turn below:

(1) Mitchell and Bergman's (1991) argument that silica-rich lamproite melt (phlogopite lamproite) cannot be derived from silica-poor lamproite magma (olivine lamproites and madupitic lamproites) is based on the assumption that the silica-poor madupitic lamproites are more evolved than the phlogopite leucite/sanidine lamproites and thus the former cannot be a parental melt to the latter. Support for this includes higher levels of incompatible trace elements in the madupitic lamproites and more evolved phlogopite (i.e., depleted in Cr and Al and enriched in Fe, Ti, Ba and Na) in the madupitic lamproites relative to the phlogopite lamproites (Mitchell, 1989). As far as the REE are concerned, there are other factors more influential in altering the REE abundances of igneous rocks than fractional crystallization and degree of melt evolution. For example, the mineralogy, chemical composition and degree of partial melting of the source can highly influence the trace element content of the lamproite melts and the chemistry of crystallizing mineral phases. As a result, the higher REE contents of madupitic lamproites relative to phlogopite lamproites do not necessarily indicate the former is more evolved than the latter.

Even if it is assumed that the madupitic lamproites cannot be parental to phlogopite lamproites, as suggested by Mitchell and Bergman (1991) and Mitchell (1995a), there is still the question as to why the phlogopite lamproites are the only primary lamproitic magmas in the Leucite Hills. Madupitic lamproites indeed contain many geochemical characteristics of a primary melt and thus should be considered as such.

(2) The second point of Mitchell and Bergman (1991) is based on least-squares mixing calculations that suggest the olivine lamproites from West Kimberley may be derived from phlogopite lamproites if phlogopite and orthopyroxene are retained and olivine and minor clinopyroxene are added to the melt (Mitchell and Bergman, 1991). Mitchell and Bergman (op. cit.) contend that the trace element data do not fit the results derived from the least-squares mixing calculations. However, it remains unclear as to

whether the data from the olivine lamproite samples used for the mixing calculations reflect the presence of phenocrystic or xenocrystic olivine. If the olivine is phenocrystic, then the primary character of such lamproite should be clearly evident based on petrography and geochemical data (e.g., Mg#, Ni and Cr content). On the other hand, if the olivine is xenocrystic, then such a calculation would be meaningless. Mitchell and Bergman (1991) and Mitchell (1995a) do, however, consider the olivine lamproites as hybrid rocks formed as a result of mixing between differentiates of primitive lamproite with olivine derived from the lithospheric mantle. Thus, the least-square calculations likely involved the xenocrystic olivine-bearing lamproites.

The detailed examination of the petrography and chemical composition of olivine-bearing phlogopite lamproites from the South Table Mountain reveals that olivine is xenocrystic in origin and thus it should not have any bearing on the chemical composition of the host lava. This is supported by the similarity in the major element abundances of the phlogopite lamproites to those of the olivine-bearing phlogopite lamproites. The trace element composition of the olivine-bearing phlogopite lamproites and phlogopite lamproites (excluding Ni and Cr) also show little variation and in some cases there is similarity in abundance levels between the two. Likewise, the Sr, Nd and Pb isotopic compositions of the phlogopite lamproites and olivine-bearing phlogopite lamproites are comparable to each other, but completely dissimilar from those of the madupitic lamproites. In the case of the olivine-bearing phlogopite lamproites from the South Table Mountain, contrary to Foley's (1993) hypothesis, the olivine lamproites from the Leucite Hills are not distinct melts but they are only phlogopite lamproites containing xenocrystic olivine. The olivine-bearing phlogopite lamproites thus appears to be no more evolved than the phlogopite lamproites. Even if there has been any interaction between the phlogopite lamproite melt and the olivine, it has occurred to such a minimal extent that no substantial geochemical changes, particularly in the trace element and isotopic compositions, have occurred.

It is still not clear why the olivine xenocrysts dominate at South Table Mountain and are absent or rare in other localities in the Leucite Hills. This, however, may be the result of small variations in the viscosity of the magma, its volatile content, and the mineralogy of the source.

(3) The third point rejects the primary character of madupitic lamproites by assuming that their ultrabasic character simply results from the high modal abundances of phlogopite (Mitchell and Bergman, 1991). This argument is, however, normally applied to cumulate rocks. Since the madupitic lamproites from the Leucite Hills do not show any cumulate textures, the ultrabasic nature of the whole rocks most likely reflects the chemical character of the melt itself. Mitchell and Bergman (1991) also point out that phlogopite in madupitic lamproites is chemically more evolved than that in the phlogopite lamproites. If that is the case and the presence of phlogopite can influence the geochemical characteristic of madupitic melts, as proposed by these authors, then madupitic lamproites should be more evolved in all aspects of their geochemistry compared to phlogopite lamproites. However, it is clear from the major element compositions that the madupitic lamproites are more mafic and less silica-saturated than the phlogopite lamproites (i.e., less evolved).

It has previously been shown that madupitic lamproites and phlogopite lamproites cannot be related by fractional crystallization or by crustal assimilation. The madupitic lamproites and phlogopite lamproites form two distinct groups on most of the major, trace element and isotope diagrams. This is an intriguing observation, that places some constraints on the origin of both types of lamproites. It is suggested here that partial melting in a veined mantle can produce the observed variations in the chemical composition of the Leucite Hills lamproites. As will be discussed in the next chapter, metasomatic veins are considered important in the petrogenesis of the lamproites (Foley, 1992). In order to explain the distinct isotopic signatures of the two groups of lamproites, it is important to assume that various minerals in the mantle source have retained their isotopic compositions since their crystallization in the mantle. In other words, it is assumed that the isotopic composition of the mineral grains have not been homogenized prior to the partial melting event.

If an heterogeneous mantle source composed of apatite-rich, phlogopite-poor and apatite-poor, phlogopite-rich veins undergoes partial melting, then the phlogopite-poor veins would generate melts with higher LREE, higher $^{143}\text{Nd}/^{144}\text{Nd}$ and lower $^{87}\text{Sr}/^{86}\text{Sr}$ ratios than the phlogopite-rich veins. As suggested by Foley (1992), wall rock olivine and orthopyroxene do not play a significant role in changing the incompatible trace element

and isotopic composition of the melt, although they can affect the compatible trace element and major element composition of the melt. Earlier melts are normally dominated by vein components, whereas later melts have a greater contribution from the wall rock (Foley, 1992). Once the partial melting process is completed, the melt leaves the site of melting and erupts at the surface to crystallize as either madupitic lamproite or phlogopite lamproite. As an alternative model, it can be assumed that the modal mineralogy of the veins is relatively uniform. In addition, the isotopic composition of the mineral phases do not homogenize during partial melting. Apatite is normally consumed prior to phlogopite because of its lower melting point. As a result, such a melt would be characterized by high LREE, high $^{143}\text{Nd}/^{144}\text{Nd}$ and low $^{87}\text{Sr}/^{86}\text{Sr}$ ratios. Melting at higher degrees of partial melting would generate melts with a greater contribution from phlogopite resulting in higher $^{87}\text{Sr}/^{86}\text{Sr}$ ratios than in the earlier formed magmas.

It is important to note that the models described above are probably two of many that can explain the geochemical variability of the Leucite Hills lamproites. In addition, phlogopite and apatite are considered only as simple examples of metasomatic minerals that can influence the compositions of the partial melts. In general, the described models show that variations in source mineralogy or degree of partial melting are able to create melts with distinct geochemical compositions. However, other factors such as variations in pressure, temperature, $f\text{O}_2$ and volatile contents on the source can also influence the chemical composition of the partial melts.

5.5. Discussion of source evolution

The geochemical characteristics of the Leucite Hills lamproites reveal that they originate from the partial melting of a metasomatised mantle source. However, the nature of such a source and its history should be understood in detail before any petrogenetic model can be proposed for the origin of the Leucite Hills lamproites. In the following sections, several aspects of the Leucite Hills lamproites will be discussed. These are: 1. the relations between the Leucite Hills lamproites and known mantle reservoirs, 2. whether the source region of the Leucite Hills lamproites resides in the sub-continental lithospheric or sub-lithospheric mantle, and 3. the origin of the distinctive trace element and isotope geochemistry of the Leucite Hills lamproites.

5.5.1 Relations to known mantle components: trace element and isotopic systematics

The mantle has been divided into a number of components based on the isotopic composition of basalts from oceanic environments. Such components show that the mantle is not homogeneous and that the isotopic variations have existed for millions of years. The following possible end-member mantle components or reservoirs have been identified: DMM (Depleted MORB Mantle), HIMU (High U/Pb), BSE (Bulk Silicate Earth), PREMA (PREvalent Mantle), EMI and EMII (Enriched Mantle I, II) (Zindler and Hart, 1986; Hart, 1988). It has been shown that mixing among the mantle components can explain many of the geochemical differences observed in MORB and OIB. DMM, characterized by low $^{87}\text{Sr}/^{86}\text{Sr}$ (0.7015 – 0.7025), $^{206}\text{Pb}/^{204}\text{Pb}$ (16 – 17.5) and high $^{143}\text{Nd}/^{144}\text{Nd}$ (0.51325 – 0.5136), is the dominant component in the source of many MORBs. HIMU has very high $^{206}\text{Pb}/^{204}\text{Pb}$ (21.0 – 22.0) that reflects the high μ value ($^{238}\text{U}/^{204}\text{Pb}$) in the mantle source. HIMU is also characterized by low $^{87}\text{Sr}/^{86}\text{Sr}$ (0.7028) and intermediate $^{143}\text{Nd}/^{144}\text{Nd}$ (0.51285) that indicates the source was enriched in U and Th relative to Pb without concomitant increase in Rb and Nd relative to Sr and Sm, respectively. The loss of Pb from mantle into the core and the removal of Pb from the mantle by metasomatic fluids are among a number of mechanisms that have been proposed for the origin of the HIMU mantle component (Dickin, 1997). BSE represents a mantle component which has the chemistry of the bulk silicate Earth. The composition of such reservoir is equivalent to the composition of a homogeneous primitive mantle which formed during core formation but prior to the formation of the continents. BSE has the following isotopic compositions: $^{87}\text{Sr}/^{86}\text{Sr}$ (0.7045 – 0.7052), $^{143}\text{Nd}/^{144}\text{Nd}$ (0.51264), $^{206}\text{Pb}/^{204}\text{Pb}$ (18.4), $^{207}\text{Pb}/^{204}\text{Pb}$ (15.58) and $^{208}\text{Pb}/^{204}\text{Pb}$ (38.9). Although some oceanic basalts have isotopic compositions similar to the BSE, there is no geochemical evidence that such a reservoir still exists. PREMA is envisioned as the most common and frequently reported mantle component in the oceanic environment that has the following isotopic compositions: $^{87}\text{Sr}/^{86}\text{Sr}$ (0.7033), $^{143}\text{Nd}/^{144}\text{Nd}$ (0.5130), $^{206}\text{Pb}/^{204}\text{Pb}$ (18.2 – 18.5). The enriched mantle components, EMI and EMII are both characterized by low $^{143}\text{Nd}/^{144}\text{Nd}$ (0.51235 – 0.5124) but EMI has lower $^{87}\text{Sr}/^{86}\text{Sr}$ (0.7052), $^{206}\text{Pb}/^{204}\text{Pb}$ (17.5), and $^{207}\text{Pb}/^{204}\text{Pb}$ (15.47) than EMII $^{87}\text{Sr}/^{86}\text{Sr}$ (0.7080), $^{206}\text{Pb}/^{204}\text{Pb}$ (19.2), $^{207}\text{Pb}/^{204}\text{Pb}$ (15.7). The two major contributors to the geochemical characteristics of the enriched

mantle are oceanic sediments and continental crust. For example, Weaver (1991) favors subduction of pelagic sediments and terrigenous sediments, respectively, to generate EMI and EMII components while Tatsumoto and Nakamura (1991) propose that subduction of the lower and upper continental crusts, respectively can generate sources with the EMI and EMII isotopic compositions, respectively.

Young oceanic basalts are ideal for understanding mantle geochemistry because they are barely affected by the crustal contamination, which means their isotopic compositions are similar to that of the mantle source from which they were derived. Continental volcanic rocks, on the other hand, could have contributions from both the mantle and the crust. However, by studying fresh and unaltered continental flood basalts, alkaline volcanic rocks, carbonatites, kimberlites, lamproites and mantle xenoliths some information about the chemical and isotopic composition of the subcontinental mantle (i.e., lithospheric and sub-lithospheric mantle) can be provided. In particular, the lamproites and kimberlites provide evidence for enriched domains within the subcontinental mantle distinctive from the sub-oceanic mantle, whereas carbonatites suggest that sub-continental upper mantle is similar to the sub-oceanic mantle (Bell and Blenkinsop, 1989).

Samples from the Leucite Hills lamproites that have not been affected by crustal contamination and alteration (Section 4.4.3) can be used to probe the underlying mantle source region. Another advantage of studying the lamproites from the Leucite Hills is that they are young and hence direct measurement of the isotopic ratios corresponds to initial isotope composition and no age corrections need be made to the measured isotope ratios. A comparison is made between the chemical characteristics of the sub-oceanic mantle reservoirs with the trace element ratios, and Sr–Nd–Pb isotopic compositions of the Leucite Hills lamproites in order to find out whether the mantle source of the Leucite Hills lamproites can be linked with any of the known oceanic mantle components.

5.5.1.a Trace Element Ratios

It is generally believed that in the case of the primary igneous rocks, the ratios of trace elements with similar incompatibility can reflect the composition of the mantle source because such ratios are unaffected by the degrees of partial melting and crystal

fractionation. This is feasible if the degree of partial melting in the mantle is >10% because under small degrees of partial melting (<10%) the REE and other trace element ratios can fractionate. Alkaline magmas, including lamproites, are produced by relatively small degrees of partial melting in the mantle (e.g., <1% as proposed by Fraser, 1987), and thus the ratios of those incompatible trace element with very similar degrees of incompatibility (e.g., Ce/Pb, Ba/Th) should be used to assess the chemical evolution of the mantle source. Table 5.2 shows a set of such trace element ratios compiled by Weaver (1991) for the various mantle components. Considering the fact that the interpretation of the incompatible element abundance ratios of direct mantle melts are more ambiguous than the isotopic ratios, the fractionation of the trace element ratios during the partial melting episode is evaluated before such ratios are used for petrogenetic interpretation of the Leucite Hills lamproites. Quantitative modeling of the incompatible trace element ratios during partial melting was conducted using the NEWPET program (1994). In this program, the concentration of trace element in the liquid (C_L) is calculated as: $C_L = C_o (1/D_o + F(1-P))$ where:

C_o = the concentration of the trace element in the unmelted source rock

D_o = the initial bulk crystal/liquid partition coefficient

$P = \sum p_\alpha D_\alpha$ where p_α represent the proportions of phases entering the melt

F = the fraction of melting

There are a number of references citing the average chemical composition of mantle sources (e.g., Maaløe and Aoki, 1977). However, these data usually deal with a fertile lherzolite and/or a disseminated metasomatised source. As was discussed in section 5.3, the mineralogy and chemical composition of the mantle source of the lamproites are different from the normal source of basaltic rocks. An extensive literature search shows that data dealing with the modal mineralogy of the metasomatised mantle are scarce and thus any attempt to determine the chemical composition of such unusual sources is difficult. The approach used here involves the trace element ratios rather than absolute abundances of the trace element in the partial melts. Accordingly, no data regarding the trace element concentration in the source are needed and an arbitrary unit value may be used instead. The phase relations in the anhydrous system forsterite –

Table 5.2 The trace element ratios of mantle components (Weaver, 1991). Ce/Pb and Nb/U data from Hofmann et al. (1986) and Hofmann (1997).

	P Mantle	N-MORB	Cont. Crust	HIMU	HIMU	EM I	EM I	EM I	EM II	EM II
				min	max	min	max	min	max	max
Zr/Nb	14.8	30	16.2	2.7	5.5	3.5	13.1	4.4	7.8	
La/Nb	0.94	1.07	2.2	0.64	0.82	0.78	1.32	0.79	1.19	
Ba/Nb	9	4.3	54	4.7	6.9	9.1	23.4	6.4	13.3	
Ba/Th	77	60	124	39	85	80	204	57	105	
Rb/Nb	0.91	0.36	4.7	0.3	0.43	0.69	1.41	0.58	0.87	
K/Nb	323	296	1341	66	187	194	523	203	378	
Th/Nb	0.117	0.071	0.44	0.08	0.12	0.09	0.13	0.11	0.17	
Th/La	0.125	0.067	0.204	0.99	0.164	0.089	0.147	0.108	0.183	
Ba/La	9.6	4	25	6.2	9.3	11.3	19.1	7.3	13.5	
Ce/Pb	25	24	4	29	38	8.3	16	?	?	
Nb/U	34	49	10	45	49	40	49	25	45	

P Mantle = Primitive mantle

diopside – enstatite (Yoder, 1976) were used to calculate the proportions of phases entering the melt at the eutectic point for simple sources such as lherzolite. However, in the case of the more complicated systems a multi-dimensional phase diagram should be used to correctly determine the proportions of phases entering the melt (p_a). Therefore, p_a was calculated on the basis of both available melting point data and phase relations in the lherzolite system. A realistic batch melting model was preferred over Rayleigh melting because in the former the newly-formed melt is able to equilibrate with the residual phases until it passes the permeability threshold of the source and thus segregates to higher levels (Maaløe, 1985). It is also important to account for the variable melting points of mineral phases stable in the source because phases such as clinopyroxene, garnet, apatite, phlogopite and amphibole complete their melting earlier than olivine and orthopyroxene. For example, in a source that is composed of olivine 40%, clinopyroxene 40%, and orthopyroxene 20%, the proportion of the minerals that enter the melt in the eutectic composition is olivine 10%, clinopyroxene 70%, and orthopyroxene 20% which clearly differs from the initial proportions in the source (Yoder, 1976). Data for the distribution coefficient between the minerals and the melt (D_o) were mainly obtained from the Geochemical Earth Reference Model (GERM) partition coefficient compilation (<http://earthref.org/GERM>).

A large number of parameters, such as degrees of partial melting and variable source mineralogy, were involved in the trace-element modeling calculation. This has resulted in the creation of a large database of the calculated trace element ratios, the details of which are outside the scope of the present discussion. To deduce the most likely mantle components involved in the petrogenesis of the Leucite Hills lamproites, only those trace element ratios least likely to fractionate during the course of partial melting have been compared with the data from Table 5.2 and plotted on Fig. 5.12. The results of the trace element modeling show that under varying degrees of partial melting (0.1% - 20%) and various source mineralogies, Ba/La, Ba/Rb, La/Rb and U/Nb ratios are the least fractionated in the liquid. On the other hand, the ratios of Ba/Th, Ba/Nb, Rb/Nb and Ce/Pb are less than unity in the resulting melt. The results of the trace element modeling indicate that those ratios plotted on Fig. 5.12 can be used to compare the mantle source of the Leucite Hills lamproites with the known mantle components. Halliday et al.

Figure 5.12 The ratios of incompatible elements normally used to discriminate the mantle components have been used to see if the mantle sources of the Leucite Hills lamproites are close in composition to any of the known mantle components. (a) Ba/La and Ba/Th ratios of the Leucite Hills lamproites form a linear trend and extend from the fields of EMI and average lower crust composition to compositionally higher values, (b) Madupitic and phlogopite lamproites extend from the low Ba/La and Ba/Nb ratios with close to EMI and average lower crustal granulites to high ratios, (c) On Rb/Nb and Ba/Th ratio diagram, data from madupitic lamproites plot at a lower Rb/Nb ratio than those of the phlogopite lamproites and are close to data from EMI. Madp lamp: madupitic lamproites, Phg lamp: phlogopite lamp, U.C. = upper crust, L.C. = lower crust, N-MORB: normal MORB, LC xth, LH: lower crust xenoliths, Leucite Hills.

Figure 5.12 (continued) (d) The Leucite Hills lamproites have lower Nb/U and Ce/Pb ratios relative to N-MORB indicating the extensive involvement of subducted crustal materials in the mantle source of lamproites, (e) La/Nb versus Ce/Pb diagram shows the effect of crustal materials in altering the source composition of the Leucite Hills lamproites.

(1995) also performed trace element modeling on a set of normal and metasomatised (i.e., up to 2% modal phlogopite and amphibole) peridotites and showed that the trace element ratios such as Nb/U, Ce/Pb, Ba/Nb and Ba/Rb approach unity at $> 0.1\%$ degree of partial melting.

Trace element ratios of the Leucite Hills lamproites vary over short distances indicating that their mantle source is very heterogeneous even on the scale of hundreds of meters. The data from the Leucite Hills lamproites form a positive correlation on the Ba/Th versus Ba/La diagram (Fig. 5.12a). The trend extends from the EMI component to high Ba/Th and Ba/La ratios. The data from the madupitic lamproites plot at lower Ba/La and Ba/Th ratios than do the data from the phlogopite lamproites. Data from both the madupitic and phlogopite lamproites also show a correlation in the Ba/Nb versus Ba/La diagram (Fig. 5.12b). On the Ba/Th versus Rb/Nb diagram, madupitic and phlogopite lamproites form two isolated groups. The madupitic lamproites have lower Ba/Th and Rb/Nb ratios than the phlogopite lamproites. The Ce/Pb ratios of the Leucite Hills lamproites are similar to those of EMI but the Nb/U and La/Nb ratios of the former are lower and higher, respectively than those of the latter (Fig. 5.12d,e). This indicates that the source of the Leucite Hills lamproites contains higher U and LREE and lower Nb contents relative to the EMI source.

The model proposed for metasomatizing the mantle source ranges from recycled lower continental crust (Mahoney et al., 1991) to oceanic pelagic sediments (Chauvel et al., 1992). Because the possible involvement of various components in the petrogenesis of the Leucite Hills lamproites will be discussed later in this chapter, the compositions of the average upper, lower crust and pelagic sediments (Taylor and McLennan, 1985) and the lower crustal xenoliths from the Leucite Hills have been plotted in Fig. 5.12.

5.5.1.b Isotopic Ratios

Figure 5.13.a shows the Nd – Sr isotopic data for the Leucite Hills lamproites along with the three mantle components, EMI, EMII and HIMU. Even though the Leucite Hills lamproites have lower Nd isotopic ratios than EMI, they still plot closer to EMI than any other mantle components. On the same diagram, the madupitic lamproites lie closer to EMI than the phlogopite lamproites. Considering the fact that the uniformity of EMI is

difficult to assess (Hart, 1988), Menzies (1989) proposes that the EMI in the continental environments may possibly have lower $^{143}\text{Nd}/^{144}\text{Nd}$ ratios than its oceanic counterparts. On $^{207}\text{Pb}/^{204}\text{Pb}$ versus $^{206}\text{Pb}/^{204}\text{Pb}$ and $^{208}\text{Pb}/^{204}\text{Pb}$ versus $^{206}\text{Pb}/^{204}\text{Pb}$ diagrams (Figs. 5.13.b and 5.13.c), the EMI mantle signature completely or partially overlaps with the field of the Leucite Hills lamproites. This might seem as an additional support for the possible involvement of EMI mantle domain in the petrogenesis of the Leucite Hills lamproites. The proximity of the Leucite Hills lamproites to EMI component is also evident on the $^{143}\text{Nd}/^{144}\text{Nd}$ versus $^{206}\text{Pb}/^{204}\text{Pb}$ and $^{87}\text{Sr}/^{86}\text{Sr}$ versus $^{206}\text{Pb}/^{204}\text{Pb}$ diagrams (Fig. 5.14a and 5.14b).

5.5.2 The lamproite source: sub-continental lithospheric or sub-lithospheric (plume) mantle?

As with many other alkaline rocks, the major controversy concerning the petrogenesis of lamproitic magmas centers around whether the parental melts are derived from lithospheric or sub-lithospheric mantle sources. In the first of these two models, it is assumed that parts of the subcontinental lithosphere have been metasomatised by fluids/melts, and then melted (e.g., Menzies, 1989). In the second model, the assumption is made that magmas originate from partial melting of material entrained within a mantle plume (e.g., Bell, 2001).

Lamproites are commonly found in old continental regions such as Archean terrains and no counterparts exist in oceanic settings. The lack of oceanic environments containing isotopically evolved components with ultrapotassic geochemical characteristics can be considered as evidence for the derivation of kimberlites and lamproites from subcontinental lithospheric sources. More importantly, the isotopic compositions of mantle xenoliths from the subcontinental lithosphere exhibit a variable range that overlaps with those from lamproites (Wilson, 1989). It is thus possible to generate such volcanic rocks by partial melting of a lithospheric source. Based on this and other geochemical variations, Skinner (1989) assumed that kimberlites (i.e., both Group I and II) are derived from different parts of a subcontinental lithosphere. The involvement of metasomatised sub-continental lithosphere in generating continental magmatism has also been proposed by Bell and Simonetti (1996).

Figure 5.13 (a) $^{143}\text{Nd}/^{144}\text{Nd}$ versus $^{87}\text{Sr}/^{86}\text{Sr}$, (b) $^{207}\text{Pb}/^{204}\text{Pb}$ versus $^{206}\text{Pb}/^{204}\text{Pb}$, and (c) $^{208}\text{Pb}/^{204}\text{Pb}$ versus $^{206}\text{Pb}/^{204}\text{Pb}$ diagrams show that data from the EMI component either overlap or plot close to the data from the Leucite Hills lamproites.

Figure 5.14 (a) $^{143}\text{Nd}/^{144}\text{Nd}$ versus $^{206}\text{Pb}/^{204}\text{Pb}$, and (b) $^{87}\text{Sr}/^{86}\text{Sr}$ versus $^{206}\text{Pb}/^{204}\text{Pb}$ diagrams show that data from the EMI component either overlap or plot close to the data from the Leucite Hills lamproites. Symbols as in Fig. 5.13.

There are, however, some continental rocks with affinities to lamproites that have been derived from sub-lithospheric sources. For example, the presence of asthenospheric signatures in the Sr, Nd and Pb isotopic compositions of Group I kimberlites points to a sub-lithospheric rather than lithospheric mantle source for these rocks (Smith, 1983). le Roex (1986), on the basis of geochemical data and the plate tectonic reconstruction of southern Africa, proposed that the source region of Group II kimberlite magmatism is related to hot spot activity and thus resides within the asthenosphere. Based on these observations, it is therefore possible that the ultrapotassic melts such as those erupted in the Leucite Hills can be produced directly within a plume, even though they are largely restricted to continents and their major and trace element contents differ from those of present-day OIBs. If a plume is able to generate a variety of magma types in both oceanic and continental environments, then it may be possible to involve melting within a plume rather than melting solely of metasomatised lithosphere.

5.5.2.a Evaluating plume involvement

Before choosing any of the models, a comparison is made between the geochemical characteristics of the Leucite Hills lamproites and those of known mantle plumes that currently exist in the area.

The Yellowstone hotspot, located about 200 km northwest of the Leucite Hills (Fig. 5.15) is believed to be a surface manifestation of a mantle plume (Craig, 1997). Edgar (1983) attributed the K enrichment of the Leucite Hills lamproites to involvement of the Yellowstone plume. Mitchell and Bergman (1991) suggest that “The Yellowstone mantle plume plays a fundamental role in the formation of the Leucite Hills lamproites”. However, such a statement does not specify the nature of the Yellowstone plume and several questions need to be answered. Does the Yellowstone plume currently reside beneath Leucite Hills? Has direct melting in the plume head generated the Leucite Hills lamproites or has the plume only triggered partial melting in the overlying metasomatised lithosphere? In order to evaluate the contribution of the Yellowstone plume to the genesis of the Leucite Hills lamproite, the nature of the Yellowstone activity and the geochemical composition of its products are compared with seemingly related other volcanic rocks erupted onto the Wyoming craton.

Figure 5.15 Map showing the paleotrack of the Yellowstone hotspot, the location of alkaline igneous rocks, and the boundary of the Wyoming Craton (after Smith and Brallie, 1994; O'Brien et al., 1995). YSRP = Yellowstone - Snake River, CM = Crazy Mountains, ID = Independence, LH = Leucite Hills, MB = Missouri Break, SB = Smoky Butte.

The Yellowstone hotspot is believed by some authors (e.g., Nash and Perkins, 1997) to be the surface manifestation of an old mantle plume that can be traced back, over a period of 17 Ma, from southwestern Idaho to the present day locus of the Yellowstone Plateau magmatism in northwest Wyoming (Fig. 5.15). The Yellowstone Plateau and the Snake River Plain volcanic fields, two major products of the Yellowstone hotspot, consist of a bimodal basalt-rhyolite association (Smith and Braille, 1994). In the Yellowstone Plateau and Snake River volcanic fields, volcanism began with voluminous eruptions of high silica rhyolites, which was then followed by basaltic eruptions. The rhyolites and some of the more evolved basalts are interpreted as the products of crustal melting, and contamination of mantle-derived melts by continental crust (Hildreth et al. 1991). The more primitive olivine tholeiites, with their uniform composition and $\delta^{18}\text{O}$ values similar to mantle values, have not interacted significantly with the crust and therefore can indicate geochemical characteristics of melts from a mantle source (Menzies et al., 1983).

On a Nd versus Sr isotopic diagram (Fig. 5.16), data from the primitive basalts from Yellowstone plot to the right of the MORB – OIB array, and lie in the enriched quadrant. Noteworthy is that on the same diagram, data from the Leucite Hills lamproites and the primitive Yellowstone basalts plot reasonably close to one another and their Nd and Sr isotopic compositions more closely resemble EMI than any of the other mantle components. On the Pb-Pb isotopic diagrams shown in Fig. 5.17, some of the Yellowstone basalts and the Leucite Hills lamproites show isotopic compositions comparable to EMI. The similarities between the isotopic compositions of the Leucite Hills lamproites and some of the primitive Yellowstone basalts are also seen on the Sr-Pb and Nd-Pb isotopic diagrams (Figs. 5.18).

The Yellowstone basalts (2.2 – 0.6 Ma) and the Leucite Hills lamproites (3.1 – 0.8) have similar eruption ages, and the similarity in their isotopic signatures may indicate that melting of a common source is responsible for their generation. The main question, however, is whether the similarity in both age and isotopic composition is conclusive evidence for producing the Leucite Hills magmas by melting within the Yellowstone plume, or whether this feature is simply fortuitous.

Figure 5.16 $^{143}\text{Nd}/^{144}\text{Nd}$ and $^{87}\text{Sr}/^{86}\text{Sr}$ isotopic variation diagram for the igneous rocks emplaced in the Archean Wyoming craton. madp lamp = madupitic lamproite (Leucite Hills), phg lamp = phlogopite lamproite (Leucite Hills), BSE = bulk silicate Earth, HIMU = high μ , EMI = enriched mantle I, EMII = enriched mantle II, ylws = Yellowstone, crz mt = Crazy Mountains, smk bt = Smoky Butte, indp = Independence, misr brk = Missouri Breaks. Dashed lines mark the position of the mantle array.

Figure 5.17 (a) $^{207}\text{Pb}/^{204}\text{Pb}$ versus $^{206}\text{Pb}/^{204}\text{Pb}$, and (b) $^{208}\text{Pb}/^{204}\text{Pb}$ versus $^{206}\text{Pb}/^{204}\text{Pb}$ isotopic diagrams showing Pb isotopic ratios for Cretaceous to Quaternary igneous rocks emplaced into the Archean Wyoming craton. Symbols as in Fig. 5.16.

Figure 5.18 (a) $^{87}\text{Sr}/^{86}\text{Sr}$ versus $^{206}\text{Pb}/^{204}\text{Pb}$, and (b) $^{206}\text{Pb}/^{204}\text{Pb}$ versus $^{143}\text{Nd}/^{144}\text{Nd}$ isotopic diagrams for the Cretaceous to Quaternary igneous rocks emplaced into the Archean Wyoming craton. Symbols as in Fig. 5.16.

To assess the nature of the mantle underlying the Wyoming craton and to evaluate the possible involvement of the Yellowstone plume, the Sr, Nd and Pb isotopic compositions of some Cretaceous to Tertiary igneous rocks emplaced into the Wyoming Archean Province are compared with those of the Leucite Hills lamproites and the Yellowstone basalts (Figs. 5.16-5.18). These igneous rocks are from the Crazy Mountains, Smoky Butte, Independence, and Missouri Breaks.

The Crazy Mountains complex, consisting of alkalic and subalkalic igneous rocks located 150 km north of Yellowstone (see Fig. 5.15) lies close to the northern margin of the Wyoming craton (Dudás et al., 1987; Dudás, 1991). On a plot of $^{143}\text{Nd}/^{144}\text{Nd}$ versus $^{87}\text{Sr}/^{86}\text{Sr}$ (Fig. 5.16), the data from the Crazy Mountains igneous rocks overlap with the field of the Leucite Hills lamproites. Similar features are also evident on the $^{207}\text{Pb}/^{204}\text{Pb}$ versus $^{206}\text{Pb}/^{204}\text{Pb}$, $^{87}\text{Sr}/^{86}\text{Sr}$ versus $^{206}\text{Pb}/^{204}\text{Pb}$ and $^{143}\text{Nd}/^{144}\text{Nd}$ versus $^{206}\text{Pb}/^{204}\text{Pb}$ diagrams (Figs. 5.17 and 5.18). The isotopic similarity between the Crazy Mountains alkaline and sub-alkaline rocks with the Yellowstone basalts may indicate that the Crazy Mountains magmatism was related to the Yellowstone plume activity. However, the spatial and temporal relationships between the Crazy Mountains and Yellowstone volcanic rocks may contradict this idea. The Crazy Mountains igneous rocks are 50 Ma old, and at the time of eruption, were located further to the north of their present-day position. Considering the 3 cm/yr rate of North American plate motion to the west (Pierce et al., 1993), the Crazy Mountains must have been located, at 50 Ma ago, 1500 km to the northeast of the present-day Yellowstone hotspot. It seems unlikely that the Yellowstone plume 50 Ma ago could have extended 1500 km from its current location and generated the volcanic rocks now observed in the Crazy Mountains. In addition, it is questionable whether the plume could have retained the same isotopic composition over 50 Ma. There is also evidence that contradicts the temporal and spatial feature of the Yellowstone plume inferred from the Crazy Mountains volcanism. Pierce et al.'s (1993) study of the uplift and tectonic events in southern Idaho indicated that the plume head encountered the WSW-moving North American plate around 10 Ma ago.

The Smoky Butte lamproite, a 27 Ma old complex (Marvin et al., 1980; in Fraser, 1987), is located 300 km northeast of Yellowstone National Park (see Fig. 5.15). The Sr, Nd and Pb isotopic composition of the Smoky Butte lamproites (Fraser, 1987) have the

lowest $^{143}\text{Nd}/^{144}\text{Nd}$ (0.51127 – 0.51142), $^{206}\text{Pb}/^{204}\text{Pb}$ (16.02 – 16.47) and $^{207}\text{Pb}/^{204}\text{Pb}$ (15.19 – 15.28) among any of data used in Figs. 5.16 to 5.18. In contrast, the $^{87}\text{Sr}/^{86}\text{Sr}$ (0.7058 – 0.7063) and $^{208}\text{Pb}/^{204}\text{Pb}$ (36.20 – 36.58) of the Smoky Butte lamproites are comparable to those of the Leucite Hills and Independence volcanic rocks. The 92 to 84 Ma old volcanic rocks from Independence are divided into two series: intrusive low-K high alumina tholeiitic basaltic-andesite through to shoshonite to high-K dacite (HATS series) and the extrusive high-magnesium andesite to banakite (HAB series). The most primitive and uncontaminated samples from these series show large variations in their initial Sr, Nd and Pb isotopic ratios ($^{87}\text{Sr}/^{86}\text{Sr} = 0.7047\text{--}0.7057$, $^{143}\text{Nd}/^{144}\text{Nd} = 0.51140\text{--}0.51190$, $^{206}\text{Pb}/^{204}\text{Pb} = 15.59\text{--}17.35$, $^{207}\text{Pb}/^{204}\text{Pb} = 15.32\text{--}15.41$, and $^{208}\text{Pb}/^{204}\text{Pb} = 36.78\text{--}37.21$). On the basis of their Sr, Nd and Pb isotopic data, the Independence volcanic rocks define two distinct groups (Figs. 5.16-5.18).

The last group of rocks to consider is the Missouri Breaks kimberlite – alnöite suite (45 - 50 Ma), located on the northern margin of the Archean Wyoming craton. They show the most radiogenic initial Nd and Pb isotopic signatures among any of the Cretaceous and younger igneous rocks emplaced into the Wyoming craton (Scambos, 1987; Scambos, 1991). The Sr, Nd and Pb isotopic signatures of the Missouri Breaks kimberlites and alnöites are similar to BSE (Fig. 5.16 and 5.18) and may indicate derivation from a sub-lithospheric mantle source.

In general, the isotopic compositions of the Cretaceous and younger volcanic rocks emplaced into the Wyoming craton show that their mantle source is extremely heterogeneous. Moreover, the isotopic data from the Smoky Butte, Independence and Missouri Breaks igneous rocks are different from those of the Yellowstone basalts. If it is assumed that a very heterogeneous plume generated all of these melts, then the Yellowstone plume must have extended for thousands of kilometers beneath the Wyoming craton over several millions of years.

Another argument that can be used against any model linking the Cretaceous to younger volcanism in Wyoming Province to the partial melting in the Yellowstone plume head is the random scattering of these volcanic rocks throughout the Wyoming craton. Hotspot activity would be expected to create a linear array of eruption centers during plate motion.

As a final note, the ultimate origin of the Yellowstone hotspot has been questioned by some authors. For example, Christiansen (1993) examined the relations between rift zones in the surrounding areas and the propagation of the Yellowstone hotspot from 12 Ma to the present time and proposed that these relations are inconsistent with the geometry of a deep mantle plume. Rather, the regional lithospheric extension and basal-lithospheric shear melting might have caused magmatism at the Yellowstone hotspot (Christiansen, 1993). Similar arguments have been invoked for other well-known volcanic eruptions previously associated with the plume activity (e.g., Anderson et al., 1992).

5.5.2.b Evaluating lithospheric involvement

According to Menzies (1989), the isotopic and trace element data from mantle xenoliths and volcanic rocks (Smoky Butte, Crazy Mountains and Independence) from the Wyoming craton clearly exhibit an EMI signature. Such an enriched reservoir is considered by Menzies (1989) to have resided within the lithospheric mantle since the Archean.

An important geochemical feature to emerge from the Leucite Hills lamproites is a negative Nb anomaly shown on the normalized trace element diagrams (Fig. 4.19). A negative Nb anomaly, observed in the Etendeka HTZ (High Ti-Zr) basalts and some basalts from the Walvis Ridge (Milner and le Roex, 1996) was considered to be produced by melting of continental lithospheric mantle. In contrast, Tristan, Inaccessible and Gough samples, considered to be plume-derived do not show a Nb anomaly, a feature that is consistent with a derivation from a source located below the subcontinental lithosphere (i.e., sub-lithospheric mantle source).

The depletion in Nb and other HFSE is normally considered to be a characteristic of subducted slab materials (Pearce, 1982; Cox, 1988; Hawkesworth et al., 1990). Even though lamproites (except those from Spain) are rarely found in subduction-related tectonic settings, Bergman (1987) observed that lamproites are commonly located above fossil Benioff Zones. Similarly, Mitchell and Bergman (1991) suggest that the tectonic settings and compositional traits of lamproites indicate they might represent partial melts of ancient subduction zones. If subduction indeed took place, then subcontinental

lithosphere might be the most appropriate place for the storage of ancient, isotopically evolved components.

5.5.2.c EMI component in the Wyoming mantle: fact or fiction?

The assumption that an EMI signature exists within volcanic rocks of the Wyoming Province (Menzies, 1989) might be considered as support for the possible involvement of a heterogeneous EMI plume or enriched sub-continental lithosphere in generating the Cretaceous to Quaternary volcanic rocks found in the Wyoming craton. However, before accepting any model that involves EMI in the petrogenesis of the Leucite Hills lamproites, two basic questions have to be answered: (1) does the EMI component reside in the sub-continental mantle or plume?, and (2) is the proximity of the isotopic data from the Leucite Hills lamproite to EMI merely coincidental?

EMI, along with other mantle components considered to characterize oceanic plumes, were originally proposed on the basis of the isotopic variations found among OIBs (Zindler and Hart, 1986). As such, if EMI is observed in continental environments, it is commonly considered to be related to plume activity. In the most recent compilation of the isotopic compositions of carbonatites, Bell and Tilton (2001) pointed out that the EMI signature tends to occur in regions where seismic tomography indicates prominent low-velocity zones in the lower mantle. This and other observations, such as the dominant restriction of EMI to the oceanic environment, led Bell and Tilton (op. cit) to suggest that EMI is contained within an isotopically heterogeneous mantle plume. These authors also restrict the EMI component, along with HIMU, to the lower 1000 km of the mantle. In addition, Reid and Leeman (1993) analyzed a number of primitive basalts from the Snake River Plain emplaced in the Wyoming craton and noticed that the paired Th-Pb isotopic signatures of the basalts are similar to those of basalts derived from enriched mantle (EMI) plumes. They therefore suggested that the Snake River Plain basalts may have originated from an EMI mantle plume.

The strong EMI isotopic character from continental environments has been found in the Etendeka HTZ basalts (Erlank et al., 1984), the alkaline basalts from Eastern China (Tatsumoto et al., 1992), the Parana basalts (Hawkesworth et al., 1986), the Deccan (Allègre et al., 1982) and the Columbia River flood basalts (Church, 1985). However, all

of these volcanic rocks plot away from the field of their proximal present-day plume on the Nd-Sr isotope plot (e.g., Etendeka from the Tristan plume, and Madagascar from the Mario plume). Based on these observations, Milner and le Roex (1996) proposed that those continental basalts with EMI isotopic and trace element signatures are largely derived from small degrees of partial melting of metasomatised parts of the subcontinental lithosphere. Tatsumoto et al. (1992) contend that if EMI was derived from the core/mantle boundary layer, then alkali basalts with EMI signature erupted in areas such as Sino-Korean Platform would necessarily lie over hot spots, which is not the case. In addition, the eastern Eurasian continental margin which exhibits EMI signatures in lavas is an area of cold subduction, not a region of hotspot activity (Richards and Engebreston, 1992) and thus is unrelated to plume activity. The presence of EMI in the sub-continental lithosphere is further supported by the existence of this component in mantle xenoliths from alkaline basalts from eastern China (Song and Fery, 1990; in Liu et al. 1994). Sweeney and Watkeys (1990) in a study of the relationship between the lithospheric architecture and the composition of the Gondwana flood basalts found that those basalts with an EMI isotopic and trace element signature were derived from enriched lithosphere beneath old Archean terrains, while the normal basalts with no enriched signature were associated with late Proterozoic mobile belts.

Menzies (1989) has also suggested that the EMI signature resides in the Wyoming subcontinental lithosphere and is restricted to Archean terrains in the western United States. To explain the lower $^{143}\text{Nd}/^{144}\text{Nd}$ ratios of the Leucite Hills lamproites relative to the EMI, Menzies (1989) proposed that the $^{143}\text{Nd}/^{144}\text{Nd}$ ratio of EMI from the subcontinental lithosphere is lower than that of the sub-oceanic mantle domain. There are many reasons to disagree with Menzies's suggestion. First, Menzies (1989) employed only the Nd-Sr isotopic diagram to show the proximity of data from four volcanic rocks from the Wyoming craton to that of EMI. However, as seen from Figs. 5.17-18, other isotopic diagrams such as Sr-Pb, Sr-Nd and Pb-Pb isotopic plots do not show affinities between the Wyoming rocks and the EMI component. Second, Menzies (op. cit) plotted the data only from the Leucite Hills, Smoky Butte, Independence, and Crazy Mountain volcanic rocks whereas other Cretaceous to Quaternary volcanic rocks such as those from Yellowstone and Missouri Breaks were not included in the Nd-Sr isotopic diagram. A

plot of all the data from the Cretaceous to Quaternary volcanic rocks emplaced in the Wyoming craton on different isotopic diagrams show a significantly large scatter relative to EMI. Third, the proposed EMI component for the Wyoming subcontinental lithosphere has a wide range in isotopic compositions (Menzies, 1989), which is in contrast to the relative uniform EMI component originally defined using data from oceanic islands (Zindler and Hart, 1986). The question arises as whether such heterogeneous continental EMI-like component can still be considered EMI?! Fourth, the volcanic rocks from the Wyoming craton form a crude linear array on many of the isotopic diagrams (Figs. 5.17-18). On these diagrams EMI plots midway along the array and the proximity of some of the data from the Wyoming volcanic rocks to EMI seems to be coincidental. The Hf isotopic data obtained by Slaters and Hart (1991) also show that the Leucite Hills lamproites do not plot close to EMI on the $^{176}\text{Hf}/^{177}\text{Hf}$ versus ϵNd diagram.

It, therefore, seems unlikely that EMI is involved in the source of the Leucite Hills lamproites and the idea that EMI is present in the Archean Wyoming mantle should be treated with caution. Accordingly, the mere proximity of some isotopic data from the Wyoming volcanic rocks to those from EMI cannot be used as evidence to ascribe the origin of Wyoming magmatism to either partial melting within a plume head, or within the subcontinental lithosphere.

5.5.3 The origin of the distinctive trace element and isotope geochemistry of the Leucite Hills lamproites

Overall, the sequence of major events needed to produce the geochemical, isotopic and tectonic setting for the Leucite Hills lamproites can be summarized in terms of depletion, enrichment and melting. These processes are discussed in detail here in the light of recent studies of metasomatism, and mantle evolution.

5.5.3.a Depletion processes

The subcontinental mantle below Archean cratons normally exhibits evidence of depletion in compatible major and trace elements (McDonough, 1990). The depletion occurs by melt removal from the mantle source leading to the formation of the overlying

continental crust. Fast seismic velocities and strong seismic anisotropy extending to hundreds of kilometers below cratons are considered to be geophysical evidence for mantle depletion (Silver and Chan, 1988), as well as high MgO, Ni and Cr abundances and low Al₂O₃ and CaO concentrations in the peridotite xenoliths and the primitive mantle-derived melts. The depletion of Al, Na and Ca by a factor of two in lamproites, relative to alkaline basalts, indicates the involvement of a residual mantle, dominantly harzburgitic in composition in the petrogenesis of the Leucite Hills lamproites (Foley, 1992).

Lamproitic and kimberlitic melts clearly originate from sources that have undergone both depletion and enrichment events. Thus it is of interest to estimate the timing of source depletion and the volume of extracted melt from the mantle source. The major melt removal episode in the Wyoming Archean Province can be attributed to formation of continental crust between 3.5 and 2.5 b.y. ago. This age is inferred from dates from the Granite Mountains gneisses (Wooden and Mueller, 1988) and the Beartooth blocks (Muller et al., 1985), the oldest crustal rocks from the Wyoming craton. The depletion event is also supported by Re/Os work on spinel peridotite, pyroxenite and glimmerite xenoliths in Eocene minette dikes from southern Montana (Carlson and Irving, 1994). Very low ¹⁸⁷Os/¹⁸⁸Os ratios (<0.110) correspond to Re depletion model ages of 2.7 to 2.9 Ga and therefore to a melt removal episode in the Wyoming mantle during the Archean era. This is also consistent with the “dates” obtained from the lower crustal xenoliths from the Leucite Hills lamproites (Section 3.5). The most reasonable time of the depletion event that affected the mantle below the Wyoming Craton is sometime between 3.5 to 2.5 Ga.

5.5.3.b Enrichment processes

Hawkesworth et al. (1985) consider that lamproites and kimberlites are the most extreme products of various mantle enrichment processes. Given the high concentrations of incompatible elements in ultrapotassic igneous rocks, it is difficult for the ultrapotassic melts to be derived by partial melting of a depleted or even a normal garnet lherzolite.

Of the many hypotheses that have been suggested for the high incompatible element enrichment in the mantle source, metasomatism is the most widely accepted

(e.g., Menzies and Hawkesworth, 1987). The components that introduce such peculiar geochemical characteristics could be either melts or fluids. The volatile- and K_2O – rich character of the lamproites requires the addition of hydrous melts or fluids enriched in K_2O and incompatible elements. Volatile-rich, low degree, partial melts or H_2O – CO_2 fluids are among the most accepted metasomatic agents which can affect the mantle source that generated ultrapotassic magmas (Menzies and Wass, 1983; Hawkesworth et al., 1985; Foley et al., 1986; Menzies, 1987). Mantle enrichment processes are thus diverse and can be caused by either water-rich volatiles that emanated from plume heads rising through the asthenospheric mantle or by hydrous metasomatic fluids from the dehydration of subducted slabs within a destructive margin (Hawkesworth et al., 1990). However, variations in the composition of the metasomatic melts/fluids can significantly influence the chemical composition of the melts derived from the metasomatised sources. For example, Hawkesworth et al. (1984) have argued that the low Sm/Nd and Rb/Sr style of enrichment reflects the migration of small-volume silicate melts in the upper mantle, while Menzies and Wass (1983) suggest that such enrichment may also result from the migration of LREE-enriched, low-Rb, CO_2 -rich fluids.

The two main mechanisms responsible for metasomatism of the mantle are subduction and plume activity. So far, it seems that the geochemical characteristics of the Leucite Hills lamproites (e.g., depletion in HFSE) favor subduction of crustal materials as the main process for metasomatism. Fig. 5.19 compares the normalized trace element patterns of the Leucite Hills lamproites with those of the N-MORB, calc-alkaline basalts, pelagic sediments and the lower crustal xenoliths from the Leucite Hills.

Some pronounced chemical features such as Nb and Ta depletion and LILE and LREE enrichment can be observed in the Leucite Hills lamproites as well as most of the Quaternary – Cretaceous volcanic rocks emplaced into the Wyoming Craton (Fig. 5.20). Such geochemical characteristics, similar to island arc and calc-alkaline volcanic rocks, are normally observed in subduction-related volcanic rocks produced at convergent margins. The proponents of plume models, however, would argue that the subduction of crustal materials into the mantle and their subsequent trapping by upwelling plumes could explain some of the geochemical peculiarities in plume-related magmatism in the continental or oceanic domains (McKenzie and O’Nions, 1983; Ringwood, 1989).

5.5.3.c Models to describe the evolution of the Wyoming mantle source

A number of models have been proposed to explain the geochemical characteristics of the Leucite Hills lamproites and the Cretaceous to Tertiary volcanic rocks that were emplaced onto the Wyoming Craton. Vollmer et al. (1984) suggested two models for the evolution of the mantle source of the Leucite Hills lamproites. In model one, it was assumed that the source of the Leucite Hills magmas was metasomatised during the stabilization of the Archean lithosphere in the Wyoming Province about 2.7 Ga. An enrichment of 20 – 30% relative to bulk Earth is considered sufficient to generate the observed Nd and Sr isotopic variations in the lamproites. In the second model, mixing of an old metasomatised mantle source with a LILE-enriched metasomatizing fluids, some time prior to partial melting, was suggested. Vollmer et al. (1984) emphasized that the strongly negative ϵ_{Nd} value of the Leucite Hills lamproites had to be a pre-metasomatic feature in order for the mixing model to stand. In the second model, the two end-members consist of an old metasomatised mantle source (negative ϵ_{Nd}) and a recent metasomatised fluid (high ϵ_{Nd}). Vollmer et al. (1984), however, emphasized that their data were insufficient to estimate the timing of the more recent metasomatic event and to distinguish which of the models better describes the observed isotopic characteristics of the Leucite Hills lamproites.

Meen and Eggler (1987) subsequently proposed a model that attributed the variations in the Sr, Nd and Pb isotopic compositions of the Independence volcanic rocks to an ancient trace element enrichment event. Correlations of $^{207}\text{Pb}/^{204}\text{Pb}$ and $^{206}\text{Pb}/^{204}\text{Pb}$ among these rocks were considered secondary isochrons that recorded ages of mantle enrichment at 3.8 and 2.0 Ga ago (Meen and Eggler, 1987). These authors question the existence of the Farallon plate beneath the Montana-Wyoming area and thus disregard the enrichment of the Wyoming subcontinental lithosphere during late Cretaceous subduction. Based on the strongly negative ϵ_{Nd} , the moderate Sr and unradiogenic Pb isotopic compositions of the Leucite Hills and Smoky Butte lamproites, Fraser (1987) also considered the metasomatism of the source to be the result of either subduction or intra-mantle processes. Based on the Pb isotopic compositions of the Smoky Butte lamproites, Fraser (1987) suggested that the U/Pb fractionation and, thus source

metasomatism took place at 2.5 Ga. The available data were, however, insufficient to select between the pelagic sediment subduction or intra-mantle processes and more detailed work in this area was suggested.

The geochemical features of the Cretaceous to Quaternary volcanic rocks within the Wyoming craton were attributed by Eggler et al. (1988) to multistage evolution of the Wyoming mantle, involving Archean depletion and mid-Proterozoic enrichment events. The mid-Proterozoic event is related to collisional and accretionary magmatotectonic events in mobile belts about the continental core. Even though Eggler et al. (1988) emphasize that the Farallon Plate subduction was not directly involved in the arc-related magmatism in the Wyoming Province, they proposed that the Cretaceous to Cenozoic magmatism occurred during back-arc rifting involving extension and thinning of lithosphere that induced flow of asthenosphere as a result of Farallon Plate subduction.

Trace element and Sr, Nd and Pb isotopic analyses from the Crazy Mountains subalkalic and alkalic igneous rocks indicate that the source region is heterogeneous and strongly enriched in most incompatible elements. Accordingly, Dudás et al. (1987) proposed that the mantle source below the Wyoming craton had a multistage differentiation history. The mantle initially developed its distinct isotopic signature in Late Archean and was significantly influenced by a mantle-enrichment event in the mid-Proterozoic (Dudás et al., 1987). Based on the Nd model age calculations, these authors point out that the mid-Proterozoic enrichment event is regional in scale and has been recorded from most of the Cretaceous to Quaternary igneous rocks emplaced in the Wyoming craton. No specific mechanism for the mid-Proterozoic enrichment was proposed by Dudás et al. (1987). However, Dudás (1991), based on the assessment of the geologic and tectonic constraints in the Wyoming craton, argues against Eocene subduction of the Farallon Plate under the Wyoming craton. Instead, Dudás (1991) favoured a mid-Proterozoic enrichment event in the Wyoming mantle source based on Nd model ages.

Melt removal during the Archean was also favoured by Carlson and Irving (1994). Evidence included Os, Sr, Nd and Pb isotopic compositions of mantle xenoliths from the northwestern part of the Wyoming craton. Very low $^{187}\text{Os}/^{188}\text{Os}$ ratios as well as Pb-Pb and Sm-Nd isochrons from xenoliths are considered to be related to a mid-

Proterozoic enrichment event. The metasomatizing materials introduced during the Proterozoic were derived from much older, presumably Archean, crustal materials.

The involvement of young and small volume asthenospheric melts in the Wyoming lithosphere was proposed by O'Brien et al. (1995). The idea was based on the isotopic dissimilarity between the Leucite Hills - Smoky Butte lamproites and mantle xenoliths from the Wyoming lithosphere. As a result, the isotopic and trace element systematics of the volcanic rocks emplaced in the Wyoming craton originated from mixing of one or more ancient metasomatised mantle components with a dominantly asthenospheric component with an ϵNd value near or greater than zero. O'Brien et al. (1991; 1995) consider the Archean subduction event at 2.7 to 2.8 Ga as the most likely source of an ancient metasomatised component in the lithospheric mantle.

Any model proposed for the origin of the Leucite Hills lamproites should take into account the following features:

- a. The geochemical characteristics of the Archean lower crustal xenoliths from the Leucite Hills lamproites (Nb-Ta depletion) indicating subduction-related processes.
- b. The much lower Ce/Pb (17) ratio of the Leucite Hills lamproites than that of MORB and OIB (25) indicating the presence of crustal materials in their source region (Fig. 5.12). Since the lamproites contain much higher LREE enrichment than MORBs, the lower Ce/Pb ratios in the Leucite Hills lamproite might reflect the addition of crustal Pb to their source.
- c. The trace element patterns showing that average pelagic sediments and lower crustal xenoliths from the Leucite Hills lamproites exhibit similar degrees of trace element enrichment and depletion relative to the lamproites (Fig. 5.19).
- d. The strongly negative ϵNd signature from the Leucite Hills lamproites that is considered to be a pre-metasomatic feature inherited from ancient continental crustal materials (Vollmer et al., 1984). This is also supported by the work of Carlson and Irving (1994) based on dating of zircons and monazites from glimmerite and mica harzburgite xenoliths from the Wyoming craton.
- e. The similarity of trace element patterns of the Leucite Hills lamproites and other Cretaceous to Quaternary volcanic rocks from the Wyoming craton to those of volcanic rocks produced in subduction-related regimes (Fig. 5.19-20).

f. The fact that the Pb isotopic data from the Leucite Hills lamproite and their lower crustal xenoliths plot to the left of the geochron. A correlation is also observed among the Sr, Nd and Pb isotopic data from the lower crustal xenoliths and those from the Wyoming Cretaceous to Quaternary volcanic rocks (Fig. 5.21-22).

g. The relatively low Sr isotopic compositions of the Wyoming volcanic rocks (< 0.707) that indicate low time-integrated Rb/Sr ratios in the mantle source.

h. The high K and LILE contents of the ultrapotassic rocks from the Wyoming craton.

Based on these observations, three different models can be proposed for the petrogenesis of the Leucite Hills lamproites:

1. Eocene subduction within the Wyoming mantle.

Because all of the Cretaceous – Quaternary volcanic rocks in the Wyoming craton show geochemical characteristics consistent with subduction-related magmatism, they may be related to recent subduction processes. It is generally accepted that the Tertiary magmatism in western North America is the result of Eocene continental subduction. However, in order to explain the occurrences of magmatism far inland from the continental margin (i.e., on the Wyoming craton), the angle of subduction had to have been relatively shallow (Lipman, 1980). There is, nevertheless, some evidence to argue against the relation of Tertiary volcanic rocks erupted within the Wyoming craton to Eocene subduction such as the timing, spatial distribution and geochemistry (Dudás, 1991). Although the Cretaceous to Quaternary volcanic rocks in the Wyoming craton show some geochemical affinities to Eocene magmatism, they erupted earlier (Independence volcanism, 90 Ma) and later (Leucite Hills lamproites) than Eocene. These eruption ages indicate that Cretaceous to Quaternary magmatism within the Wyoming craton was not dependent on subduction of the Farallon plate in Western North America. Moreover, the Nd isotopic data from the Cretaceous–Quaternary volcanic rocks emplaced into the Wyoming Archean craton are not consistent with Eocene enrichment but point to an older metasomatic event in the mantle source.

Figure 5.19 The normalized trace element patterns of the Leucite Hills lamproites (LH lamp), lower crustal xenoliths (LC xlth), pelagic sediments (Pelg sed), N-MORB and calc-alkaline basalts (Arc bslt). Data set from: Sun and McDonough, 1989 (N-MORB), Gertisser and Keller, 1999 (arc basalt), <http://earthref.org/GERM/> (pelagic sediment).

Figure 5.20 The presence of arc-like geochemical patterns (i.e., depletion in HFSE and enrichment in LILE) in Cretaceous to Quaternary igneous rocks from the Wyoming craton is evidence for the presence of subducted materials into the mantle source. Data from Fraser (1987), Dudas (1991), and Meen and Eggler (1987). LH: Leucite Hills, smk bt: Smokey Butte, ylws: Yellowstone, cr mt: Crazy Mountains, indp: Independence, alc-alkaline basalts (Arc bslt).

Figure 5.21 $^{143}\text{Nd}/^{144}\text{Nd}$ and $^{87}\text{Sr}/^{86}\text{Sr}$ isotopic variation diagram for the Cretaceous to Quaternary volcanic rock emplaced in the Archean Wyoming craton in relation to lower-crustal xenoliths.

Figure 5.22 $^{207}\text{Pb}/^{204}\text{Pb}$ versus $^{206}\text{Pb}/^{204}\text{Pb}$ isotopic diagram for the Cretaceous to Quaternary volcanic rocks emplaced into the Archean Wyoming craton. Also shown are data from the lower crustal xenoliths. The samples plot to the left of the geochron and below the Stacey-Kramers growth line showing an old U/Pb depletion in the source. LC xlth = lower crustal xenoliths from Leucite Hills. Symbols as in Fig. 5.21.

2. Reactivation of entrapped late-Archean metasomatic materials during the mid-Proterozoic.

This model is based on model Nd age (T_{DM}) calculations, that indicates that during the mid-Proterozoic, the Wyoming lithospheric mantle was affected by pervasive metasomatism (Dudás et al., 1987; Eggler et al., 1988; Carlson and Irving, 1994). Some problems exist with the mid-Proterozoic enrichment event. First, the exact nature of the mid-Proterozoic event is not well defined. The main mechanisms for metasomatizing the mantle source involve either plume or subduction-related activities. However, the Cretaceous to Quaternary volcanic rocks emplaced into the Wyoming craton show subduction-related geochemical features, and thus the involvement of plume in the old enrichment of the source is ruled out. There is little or no evidence, however, for mid-Proterozoic subduction (e.g., Dudás, 1991) such as a subduction geometry or arc volcanism in the Wyoming craton. Second, the Nd model age calculations are based on the Nd isotopic composition and the Sm/Nd ratio measured in the lamproites. If, however, the Sm/Nd ratio of the sample is changed during recent metasomatism or partial melting, as seems to be the case for the Leucite lamproites, then Nd model ages cannot be used to place restrictions on the timing of the metasomatic event and the nature of the metasomatic components in the source.

3. Late-Archean subduction combined with recent metasomatic activity.

This model seems to be most consistent with the geochemical features of the Leucite Hills lamproites and other recent volcanic rocks emplaced in the Wyoming craton. In this model, a late-Archean lithospheric component is mixed with a recent sub-lithospheric metasomatic component (Fig. 5.23). The Wyoming mantle was either depleted during the late Archean, or represented by an unfractionated CHUR-like source. Fluids/melts derived from the subducting slab infiltrated the overlying mantle across a large area and froze either at the base of or within the sub-continental lithosphere.

There is now sufficient evidence to suggest that the late-Archean subduction event caused metasomatism in the Wyoming mantle source. In addition, incompatible element patterns, particularly the Nb and Ta depletion, of the Archean lower crustal xenoliths imply that the Wyoming mantle source was metasomatised by Archean,

Figure 5.23 (a) Pelagic sediments were subducted under the Wyoming craton and metasomatized the overlying lithosphere at about 2.6 Ga. The underplated magma from partial melting of the mantle formed the lower crust. (b) More recently, an uprising plume provided heat for partial melting and added volatiles, K_2O and incompatible elements to the site of melting in the Wyoming lithospheric mantle.

subduction-related, metasomatic components. In general, the pronounced Nb and Ta depletion in lamproites, including that of the Leucite Hills, is considered evidence for a fossil Benioff Zone (Bergman, 1987). The Archean subduction process is also indicated by the arc-like geochemical features of a series of compositionally diverse, late-Archean rocks from the eastern Beartooth Mountains (Mueller et al. 1983; Wooden and Mueller, 1988). There is no evidence for any subduction event in the Wyoming mantle subsequent to the Archean event. As a result, the arc-like geochemical patterns of the Wyoming Cretaceous to Quaternary volcanic rocks and thus that of their mantle source reflect late-Archean subduction-related metasomatism.

The very negative ϵ_{Nd} , unradiogenic Pb and moderately radiogenic Sr isotopic compositions of the Wyoming Cretaceous to Quaternary volcanic rocks indicate that the Wyoming sub-continental mantle has isotopic ratios that reflect low time-integrated Sm/Nd, U/Pb and Rb/Sr ratios. Such geochemical features were most likely superimposed on the mantle source during Archean subduction.

The geochemical characteristics of lower crustal xenoliths from the Leucite Hills show time-integrated low Sm/Nd, U/Pb and Rb/Sr (Section 3.6). This, combined with the similarity in the trace element patterns and the correlation in Pb-Pb and Sr-Nd isotopic diagrams between data from the lower crustal xenoliths and the volcanic rocks from the Wyoming craton suggest that lower crustal materials were probably involved in metasomatizing the Wyoming sub-continental lithosphere. The significance of lower crustal material in metasomatizing the mantle source has been proposed by some authors. For example, the minettes from Schirmacher Oasis, East Antarctica, have characteristically enriched trace element patterns and high $^{207}\text{Pb}/^{204}\text{Pb}$ ratios (Hoch et al., 2001). The geochemical characteristics of these rocks can best be explained by partial melting of a lithospheric mantle source that was enriched by metasomatic agents derived from recycled lower crustal rocks (Hoch et al., 2001). However, the problem with the involvement of lower crustal materials with the mantle source of the Leucite Hills lamproites is that the lower crustal rocks themselves formed as a result of late-Archean subduction (Chapter 3) and it seems unlikely that they could be subducted again into the mantle at about the same time.

Among crustal materials, ancient pelagic sediments are also characterized by low Sm/Nd and U/Pb and variable Rb/Sr ratios (Othman et al., 1989). Subduction of such materials could have left behind a metasomatised mantle source with low Sm/Nd, U/Pb and Rb/Sr ratios. Alternatively, slab dehydration and/or partial melting, could also affect the U/Pb and Rb/Sr ratios of sediment being recycled into the mantle (White and Dupré, 1986). U and Rb extraction would lower the U/Pb ratios and Rb/Sr ratios in the slab. Such ratios could also have been lowered during a Late-Archean metamorphic event resulting from the continental collision that led to the formation of granulite in the lower crust.

As an alternative to subduction, Wooden and Mueller (1988) suggested that μ values of the Wyoming mantle could be lowered by a collisional event at 2.8 Ga. Fraser et al. (1985) attributed the Pb isotopic composition of the lamproites from the Wyoming Province to U depletion of the mantle during the late Archean.

Based on the present-day isotopic variations of the Cretaceous to Quaternary volcanic rocks emplaced in the Wyoming craton and the assumption that the Wyoming sub-continental lithosphere behaved as a closed system to Rb, Sr, Sm, Nd since late-Archean metasomatism, it is possible to calculate Sm/Nd and Rb/Sr ratios of the metasomatised Wyoming sub-continental mantle. Depending on whether a depleted (DM) or unfractionated (CHUR or BE) mantle was used as a reference source, different Sm/Nd ratios can be calculated for the metasomatised, Wyoming sub-continental lithosphere. Calculations show that if late Archean subduction affected a DM source, the Sm/Nd ratios of the metasomatised Wyoming sub-continental lithosphere would lie between 0.180 to 0.288, and the Rb/Sr ratios between 0.038 to 0.052. In the case of the CHUR and BE sources, the Sm/Nd and Rb/Sr ratios of the mantle source vary from 0.201 to 0.269 and 0.036 to 0.052, respectively. Figs. 5.24-25 illustrates the Nd and Sr isotopic evolution paths for volcanic rocks emplaced in the Wyoming craton.

There is little geochemical and geological evidence to support a mid-Proterozoic enrichment event or Eocene subduction of the Farallon Plate. The lack of evidence for post-Archean major subduction indicates that the metasomatised portions of the Wyoming sub-continental lithosphere mantle have remained relatively undisturbed from Archean until recent plume activity. Some geochemical features support the involvement

Figure 5.24 (a) Calculated Nd isotopic evolution paths for the Cretaceous to Quaternary volcanic rocks during late Archean metasomatism of a DM source. The Wyoming sub-continental lithosphere (WSCL) would attain Sm/Nd ratios ranging from 0.180 to 0.288, (b) If the late-Archean metasomatism affected a CHUR-like reservoir, the WSCL would attain Sm/Nd ratios of 0.201 to 0.269. The measured Sm/Nd ratios of the Wyoming volcanic rocks range from 0.12-0.17. DM curve constructed assuming $^{143}\text{Nd}/^{144}\text{Nd} = 0.51315$ and $^{147}\text{Sm}/^{144}\text{Nd} = 0.214$ and CHUR curve constructed assuming $^{143}\text{Nd}/^{144}\text{Nd} = 0.51263$ and $^{147}\text{Sm}/^{144}\text{Nd} = 0.1967$ ((DePaolo, 1988).

Figure 5.25 (a) Calculated Sr isotopic evolution paths for the Cretaceous to Quaternary volcanic rocks during late Archean metasomatism of a DM source. The WSCL would attain Rb/Sr ratio ranging from 0.038 - 0.052, (b) if late-Archean metasomatism affected a BE-like reservoir, the Rb/Sr ratios of WSCL range from 0.030-0.044. The measured Rb/Sr ratios of the Wyoming volcanic rocks range from 0.029-0.111. DM curve constructed assuming $^{87}\text{Sr}/^{86}\text{Sr} = 0.7025$ and $^{87}\text{Rb}/^{86}\text{Sr} = 0.052$ and BE curve constructed assuming $^{87}\text{Sr}/^{86}\text{Sr} = 0.7045$ and $^{87}\text{Rb}/^{86}\text{Sr} = 0.6989$ (Faure, 1986).

of a recent enrichment event. The LOI contents (up to 9 wt%) of the Leucite Hills lamproites indicate that a volatile-rich component was involved in metasomatizing the mantle source. The volatiles have to be dominantly H₂O-rich since hydrous-bearing phases form important minerals in these lamproites. In addition, primary carbonate phases have yet to be found in lamproites (Mitchell, 1995b). As an example, the higher LOI, as well as Ba contents of the madupitic lamproites relative to the phlogopite lamproites (Fig. 5.26) indicate the importance of volatile addition to the source.

Edgar (1983) has shown that a negative correlation exists between the K/K+Na ratio as well as Ba content of ultrapotassic rocks from the western U.S.A. and distance from the Yellowstone hotspot. Those volcanic rocks situated closest to the hotspot (i.e., the Leucite Hills lamproites) contain the highest K/K+Na ratio and Ba values. Edgar (1983) proposed that K was derived from the Yellowstone plume and subsequently carried by aqueous fluids known to be efficient agents for transporting K₂O. The relatively low Sr isotopic composition of the Leucite Hills lamproites suggests that insignificant amounts of K were added to the mantle source during late-Archean subduction. It thus seems reasonable to assume that the plume supplied most of the K to the source. The K, Ba and LILE addition to the Wyoming mantle must have affected large parts of the mantle. It seems reasonable to assume that the mantle underlying the Leucite Hills and Smoky Butte lamproites has been metasomatised by K- and volatile-rich agents to a greater extent than elsewhere in the Wyoming lithosphere.

The comprehensive model proposed here is based on Sr, Nd and Pb isotopic composition of lower crustal xenoliths and lamproites from the Leucite Hills, as well as the previously published data from Cretaceous to Tertiary volcanic rocks emplaced in the Wyoming craton.

5.6 Summary

The two main groups of lamproites in Leucite Hills, the madupitic lamproites and the phlogopite lamproites, cannot be related to each other by either fractional crystallization or crustal contamination. The variation in the chemical composition of the madupitic lamproites and the phlogopite lamproites is thus a result of mantle source heterogeneity.

Figure 5.26 The position of data from the madupitic lamproites and phlogopite lamproites on the LOI versus Ba diagram show that a volatile-rich component was involved in metasomatising the source mantle.

The mantle source of the Leucite Hills lamproites has clearly undergone both depletion and enrichment events. The major depletion event in the Wyoming Archean Province can be attributed to the formation of the continental crust between 3.5 and 2.5 Ga ago. The first enrichment event most likely occurred during late Archean subduction.. The depletion in Nb and other HFSE suggests the involvement of subducted materials into the mantle source of the Leucite Hills lamproites. The Archean subduction is also supported by the Sr, Nd and Pb isotopic data from the Leucite Hills lamproites, the Wyoming Cretaceous to Tertiary volcanic rocks, and the lower crustal xenoliths. Due to the nature of the subducted materials (pelagic sediments?) and the high P-T conditions associated with the subduction, the metasomatised regions of the mantle source had low U/Pb and Rb/Sr ratios and high Sm/Nd ratios. The mantle source behaved as a closed system until it was metasomatised by volatile-rich components derived from a plume or mantle upwelling at more recent time. These components provided additional K₂O, LILE and volatile components to the metasomatised continental lithosphere. Those portions of the mantle mainly affected by the recent plume-related metasomatism, were the most appropriate sites for producing ultrapotassic magmas.

CONCLUSIONS

The main conclusions drawn from the study of the Leucite Hills lamproites and their lower crustal xenoliths are as follows:

1. Based on the petrography and whole rock major and trace element geochemistry, lamproites from the Leucite Hills can be grouped into phlogopite lamproite and madupitic lamproite. In addition, in all of the isotopic ratio diagrams, the phlogopite lamproites and the madupitic lamproites form two distinct groups. The olivine-bearing phlogopite lamproites contain higher MgO, Ni and Cr concentrations than the phlogopite lamproites. However, if the contribution of the olivine xenocrysts is taken into account, the two lamproite types have similar major and trace element concentrations. The similarity in the Sr, Nd and Pb isotopic ratios of the olivine-bearing phlogopite lamproites with those of the phlogopite lamproites indicate that they are genetically related and originated from the same or similar sources.
2. A LAM-ICPMS study of separated apatite, clinopyroxene and perovskite grains from the madupitic lamproites shows that the incompatible trace element budget of the whole rock is mainly controlled by perovskite and apatite and to a lesser extent by clinopyroxene. The trace element patterns of these minerals are characterized by incompatible trace element enrichment and Nb - Ta depletion.
3. The results of leaching experiments, and isotopic analyses of mineral separates indicate that isotopic ratios of the mineral phases from the madupitic and transitional madupitic lamproites in the Leucite Hills best estimate the isotopic composition of the magma.
4. Bulk assimilation modeling shows that the major and trace element contents and isotopic ratios of the Leucite Hills lamproites would hardly be affected by the assimilation of the lower crust, average upper crust and sedimentary enclaves. In addition, crustal assimilation by the phlogopite lamproite cannot produce the madupitic lamproite and vice versa. The elemental and isotopic variation between the phlogopite

lamproites and the mafic lamproites suggest that these magmas were derived from distinct sources within the mantle.

5. The variations in the major and trace elements as well as the isotopic compositions of the Cretaceous to Quaternary igneous rocks emplaced in the Archean Wyoming craton, indicate widespread heterogeneity in the Wyoming subcontinental mantle. Such variations can also exist within individual volcanic centers indicating the presence of heterogeneities on the scale of a few hundred meters.

6. On the basis of the Sr and Nd isotopic ratios, it was previously suggested that EMI is present in the source region of the Leucite Hills lamproites. However, the widespread deviation of Pb isotopic compositions of the Wyoming volcanic rocks relative to the EMI component suggests that EMI does not exist in the Wyoming sub-continental lithosphere.

7. Isotopic studies of lower crustal xenoliths from the Leucite Hills lamproites show that they are Archean in age and formed by metamorphism of a differentiated basaltic melt. The approximate Nd initial ratio suggests that the original magma was probably derived either from a CHUR reservoir or a slightly depleted mantle metasomatized during late Archean subduction.

8. The model favored for the origin of the Leucite Hills lamproites is one in which a late-Archean subduction metasomatized the lithosphere below the Wyoming craton. The metasomatized lithosphere remained isolated and behaved as a closed system until it underwent metasomatism by a recent mantle plume. Subsequent melting of the metasomatized lithosphere led to the generation of the Leucite Hills lamproitic magmas.

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APPENDIX A

SAMPLE PREPARATION

A.1 Whole Rock

Altered portions of the whole rock samples were removed and the samples were split into small pieces using a hydraulic jack. The pieces were manually crushed into <1 cm fragments using a steel crusher and ground to fine powder (<200 mesh) in a chrome-steel pulverizer. The crushing equipment was rinsed with distilled water, dried with ethanol and precontaminated between each stage.

Three gm of sample were sent to Activation Laboratories Ltd, Ancaster, Ontario for major and trace element analysis. Approximately 100 - 200 mg were used for isotopic measurements.

A.2 Mineral Separation

Whole rock specimens were trimmed and then broken into small pieces using a hydraulic jack. The broken pieces were crushed manually into powder using a steel crusher. Using plastic sieves, the powder was sieved into three sizes: >250 μm , 250-149 μm , <149 μm . The sieves were cleaned with compressed air, distilled water and ethanol between each sample. The 250-149 μm fraction was washed with distilled water in a plastic petri-dish and dried under a heating lamp. A hand magnet wrapped in a Kimwipe tissue was used to remove the strong magnetic minerals. Heavy liquids were then used to separate the minerals on the basis of their specific gravity. In methylene iodide ($d = 3.29$), perovskite ($d = 4.03$) settled whereas apatite ($d = 3.1$), clinopyroxene ($d = 3.2$) and other light minerals floated. In bromoform ($d = 2.87$), clinopyroxene and apatite settled whereas less dense minerals floated. Two fractions, one of perovskite and the other of clinopyroxene - apatite were rinsed with acetone and dried under a heating lamp. The mineral fractions were further purified using a Frantz isodynamic separator. The final mineral fractions were obtained by hand-picking under a binocular microscope.

The mineral fractions were then placed in an ultrasonic bath, and vibrated for 10 to 20 minutes, which abraded the mineral grains and removed any attached particles. The

abraded grains were washed in ethanol and nanopure water and were then leached with diluted HCL. Because the concentration of Sr, Nd and Pb had already been determined by LAM-ICP-MS, it was possible to estimate the quantity of minerals required for isotopic analysis. This was also based on the assumption that for Sr, Nd, and Pb analysis, a minimum of 1 μg Sr, 500 ng Nd and 250 ng Pb were needed to analyze a sample on the mass spectrometer.

The process of digesting the minerals is similar to that for the whole rock samples and has been described in the section below.

A.3 Sample Digestion

Samples were digested in ultrapure HF and HNO₃ using a ratio of 2:5 and evaporated to dryness. Five ml 6M HCl was added to the samples. After digestion, the samples were evaporated to near dryness and were added 5 ml of 2.5M HCl.

For Pb isotopic analysis, a portion of solution was weighed, transferred to a small disposable beaker and evaporated to dryness. The evaporate was dissolved in 1M HBr and centrifuged for 10 minutes. This solution was used for column chemistry.

Another portion of the solution was weighed and evaporated to dryness. The evaporate was dissolved in 2ml of 2.5M HCl and centrifuged in a centrifuge tube for 10 minutes prior to Rb, Sr and REE column separation. To prevent contamination and high blanks, unspiked samples were put through the columns prior to the spiked samples.

A.4 Isotope Dilution

Some samples were spiked to determine their Rb, Sr, Sm and Nd concentrations. Spikes available for isotopic dilution were enriched in ⁸⁷Rb, ⁸⁴Sr, ¹⁴⁵Nd and ¹⁵²Sm. The Sr, Sm and Nd spikes were calibrated by the author in 1998 and 1999 (Table A.1). The new concentration for Rb spike calibrated by A. Rukhlov in 2001 (personal communication) has also been included in Table A.1.

Prior to spiking, estimates were made about how much spike should be added to the solution in order to prevent over- or under-spiking of the samples. Spike and sample

Table A.1 Concentration and isotopic composition of spikes

	Isotopic ratios	Concentration	
$^{85}\text{Rb}/^{87}\text{Rb}$	0.008	Rb	0.037981 $\mu\text{mol/g}$
$^{86}\text{Sr}/^{88}\text{Sr}$	3.03418	Sr	0.01376 $\mu\text{mol/g}$
$^{87}\text{Sr}/^{88}\text{Sr}$	0.39076		
$^{84}\text{Sr}/^{88}\text{Sr}$	17.991		
$^{146}\text{Nd}/^{144}\text{Nd}$	1.4252	Nd	0.100204 $\mu\text{mol/g}$
$^{145}\text{Nd}/^{144}\text{Nd}$	35.8640		
$^{143}\text{Nd}/^{144}\text{Nd}$	0.26182		
$^{149}\text{Sm}/^{147}\text{Sm}$	266.2561	Sm	0.12081 $\mu\text{mol/g}$
$^{149}\text{Sm}/^{152}\text{Sm}$	251.8454		

solutions were weighted accurately and the mixture was then evaporated to dryness. The evaporated sample was dissolved in 2ml of 2.5M HCl and centrifuged for 10 minutes.

The precision of the isotopic dilution method is $\pm 0.5\%$ for Sr, Sm and Nd and $\pm 1\%$ for Rb.

Rb, Sr and REE column procedure

Rb, Sr and REE were separated using ion exchange chromatography. The procedure is based on HCl elution through 20 cm teflon columns of 1.1 cm internal diameter filled with 13 ml of Bio-Rad AG50W-X8 cation exchange resin (200-400 mesh). The detailed column procedure is summarized in Table A.2. Rb, Sr and REE aliquots were collected in disposable beakers and evaporated at a temperature of about 100°C.

The dried REE portion was dissolved in 0.2 ml of 0.5M HCl. Nd and Sm were separated from each other and from the other REE using reverse phase chromatography. The procedure is based on HCl elution through 20 cm Bio-Rad Vycor (quartz) columns of 6 mm diameter filled to a height of 7 cm with teflon powder coated with di-hexyl orthophosphoric acid. The column procedure is summarized in Table A.3. Details regarding the Rb, Sr and REE column construction and calibration are given in Wen (1985) and Pollock (1987).

The final solutions collected from each column were evaporated in a disposable beaker and transferred into a teflon beaker for mass spectrometry analysis.

Pb column procedure

Pb was separated using ion exchange chromatography and carried out using HCl, HBr and nanopure H₂O elution through 0.2 and 0.5 ml columns that were filled with AG1-X8 cation exchange (100-200 mesh). Table A.4 gives the detailed procedure for Pb extraction.

A.5 Blank Values

The average values of total procedure blanks obtained by the author during the period of August 1997 to December 1999 are given in Table A.5.

Table A.2 Elution chart for Rb - Sr column.

	ml	M HCl
Load Sample		
Wash	10	2.5
Wash	30	2.5
Collect Rb	10	2.5
Wash	25	2.5
Collect Sr	15	2.5
Wash	20	2.5
Wash	15	6.2
Collect REE	25	6.2
Clean	60	6.2
Condition	60	2.5

Table A.3 Elution chart for REE column.

	ml	M HCl
Load Sample		
Wash	0.2	0.15
Wash	0.2	0.15
Wash	1	0.3
Wash	1	0.3
Collect Nd	4	0.3
Wash	1	0.3
Wash	1	0.3
Wash	1	0.3
Collect Sm	4.5	0.5
Wash	3	0.5
Clean	25	6
Condition	25	0.15

Table A.4 Elution chart for 0.2 ml Pb column.

Wash	1	6 M HCl
Wash	1	H ₂ O
Wash	1	1 M HBr
Load Sample		
Wash	1	1M HBr
Wash	0.8	H ₂ O
Collect Pb	1	6 M HCl

Elution chart for 0.5 ml Pb column.

Wash	2	6 M HCl
Wash	2	H ₂ O
Wash	2	1M HBr
Load Sample		
Collect U	4	1M HBr
	1	1 M HCl
Collect Pb	3	6 M HCl

Table A.5 Procedural blank values.

	ng
Rb	0.4
Sr	1.5
Sm	0.04
Nd	0.26
Pb	1

APPENDIX B

Analytical Methods

B.1 Electron Microprobe

Major element abundances of mineral grains in polished thin sections or mineral mounts were determined by a wavelength-dispersive technique (WDS) using a Camebax MBX electron microbeam in the Department of Earth Sciences, Carleton University. Operating conditions were 15 kV accelerating voltage, 20 nA beam current and counting times of about 30 seconds or 40,000 counts. X-ray lines were chosen to minimize or eliminate possible elemental interferences. Calibrated natural or synthetic minerals and compounds were used as calibration standards to determine the contents of elements in the minerals. Raw data were converted to elemental wt% by the Cameca PAP matrix correction program. Analyses are accurate to 1-2% of the quoted values for major elements and 3-5% for minor elements.

B.2 Laser Ablation Microprobe - Inductively Coupled Plasma - Mass Spectrometry (LAM-ICP-MS)

Minerals were mounted on a Teflon plate and covered with epoxy resin. The grain mounts were then polished with corundum until they were exposed at the surface. The polished minerals were then cleaned with ethanol to remove any material from surfaces.

The trace element concentrations of the minerals were determined using Laser Ablation Microprobe - Inductively Coupled Plasma - Mass Spectrometry (LAM-ICP-MS) at the Memorial University of Newfoundland (MUN). In this highly sensitive analytical technique, a pulsed laser beam ablates the solid material in a cell and a portion of the ablated material is carried in a continuous flow of argon to the ICP where it is atomized and ionized. The mass spectrometer then measures the ion-beam intensity of the charged masses produced in the plasma.

The LAM at MUN consists of Q-switched Nd:YAG laser that can provide beams with intensities from 0.25 to 20 mJ/pulse and wavelengths from 1064 (IR) to 266 nm (UV) (Fryer et al. 1995). With careful focusing of the laser beam, sample pit sizes (both diameter and depth) as low as 10 μm may be produced. However, in order to achieve a maximum

sensitivity (i.e., lower detection limits), the sampling rate (mass ablated per unit time) has to be optimized as does the incident energy and the resolution of the beam (e.g., 100 μm). In other words, there is a trade off between obtaining smaller pits and achieving lower detection limits. For example perovskite, known for its enrichment in REE, can be ablated with low energy beam (0.1 mJ/S) and 10 μm pits. But these conditions cannot be applied to minerals with very low concentrations of REE such as olivine. Instead, a beam size of 100 μm and an energy beam of 10 mJ/S were used to detect the REE at reasonable detection limits.

The ICP-MS is a VG-PQII+™ machine, an enhanced sensitivity quadrupole ICP-MS developed by Fisons. A forward power of 1350 kW and plasma flows of 1 L/min was maintained during the analyses.

Calibration of in situ LAM-ICP-MS analyses requires the use of well-established external reference materials in conjunction with an internal standard to correct for variable ablation yields between sample and reference material. The external standard used at MUN is the synthetic glass NBS 610 that was analyzed twice before and after each run.

The concentrations of the internal standards were obtained at Carleton University by electron microprobe analysis on the minerals to be analyzed by LAM-ICP-MS. Since Ca or Si is abundant in many of the minerals analyzed, they have been used as internal standards. In addition, NBS-610 seems to produce more accurate results when Ca and Si are used as internal standards (Frey et al., 1995).

Two analyses of NBS-610 were made prior to running selected samples. The sample was then analyzed for a maximum of 150 seconds, with 50 seconds measurement of background followed by the ablation event. Ablating was usually continued until the laser beam hit the glass or glue, in which case an inflection in the signal intensity was observed.

The raw data were downloaded to a PC to be processed for the next stage. Software LAM-TRACE (version 1.76) was employed to pre-integrate the raw count rates and to create time-resolved signals. For each analysis, the time-resolved signals of elements were examined to determine the background and ablation intervals for integration. Careful inspection of signals at this stage is necessary as they may furnish significant additional data such as chemical zonation or the presence of inclusions in the crystal.

In order to calculate the precision and accuracy of LAM-ICP-MS data, BCR-2 was used as a standard analyzed during each run. The precision of data is presented in Table B.1.

Table B1 Calculated precision and accuracy of standard BCR-2 standard used for the LAM-ICP-MS analysis.

	Mean (n=19) wt%	Std Dev	Precision C.V. (%)	BCR-2 Certified wt%	Accuracy diff. (%)
MgO	5.46	5.16	95	3.48	-57
Al ₂ O ₃	14.93	1.24	8	13.64	-9
SiO ₂	58.53	5.38	9	54.06	-8
CaO	6.85	0.30	4	6.95	1
TiO ₂	1.95	0.16	8	2.24	13
MnO	0.17	0.00	1	0.177	2
FeO	12.52	2.79	22	13.41	7
	ppm			ppm	
V	374.9	22.35	6	407	8
Sc	30.27	14.04	46	32.6	7
Cr	18.63	0.36	2	16	-16
Co	39.76	2.59	7	37	-7
Rb	50.33	6.02	12	47.51	-6
Sr	304.9	24.13	8	337	10
Y	31.21	2.98	10	32.51	4
Zr	165.78	16.43	10	184.4	10
Nb	10.64	0.71	7	13.14	19
Cs	1.21	0.15	13	1.19	6
Ba	566.5	53.40	9	684	17
La	25.08	1.80	7	25.3	1
Ce	49.41	3.30	7	53.6	8
Pr	6.2	0.66	11	6.83	9
Nd	25.03	2.35	9	28.63	13
Sm	6.02	0.97	16	6.67	10
Eu	1.92	0.31	16	2	4
Tb	0.92	0.15	17	1.04	12
Gd	5.52	0.79	14	6.8	19
Dy	5.75	0.84	15	6.38	10
Ho	1.18	0.13	11	1.29	8
Er	3.11	0.52	17	3.66	15
Tm	0.43	0.07	15	0.54	21
Yb	3.04	0.61	20	3.34	9
Lu	0.48	0.09	19	0.51	5
Hf	4.09	0.61	15	5.17	21
Ta	0.58	0.07	11	0.78	26
Pb	13.15	2.44	19	10.34	-27
Th	5.5	0.69	13	5.98	8
U	1.69	0.57	34	1.7	0

B.3 Inductively Coupled Plasma (ICP) and Inductively Coupled Plasma - Mass Spectrometry (ICP-MS)

The whole rock major elements and trace elements were analyzed using ICP and ICP-MS facilities, respectively, at Activation Laboratories Ltd at Ancaster, Ontario. The description below is taken directly from the laboratory manual and more information can be found at <http://www.actlabs.com>:

“Samples are taken into solution using fusion techniques in which 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 and mixed continuously until completely dissolved (30 minutes).

For major oxides, samples were run on a simultaneous/sequential Thermo Jarrel-Ash ENVIRO II ICP. Calibration is performed using 7 certified USGS and Canmet certified reference materials. One of 7 standards is used during the analysis for every group of ten samples. In the ICP method, the solution is then introduced into a radio frequency excited plasma (8000°K). Each element in the solution produces a characteristic spectrum. The intensity of the spectral lines is directly proportional to the quantity of the element present. The advantage of this technique includes: 1. ICP is a multi-element technique and the major elements and some trace elements can be determined simultaneously to sensitivities better than x-ray fluorescence, 2. elements determined by ICP are very complimentary to the Instrumental Neutron Activation Analysis (INAA) method.

For the trace elements analyses, samples prepared from the previous stage are spiked with internal standards to cover the entire mass range and are further diluted and introduced into a Perkin Elmer SCIEX ELAN 6000 ICP-MS. ICP-MS technology couples a standard ICP torch used for the ICP methodology to a quadropole mass spectrometer to measure mass rather than emission line. The main advantage includes the very low sensitivities as well as multi-element nature of the technique.”

Accuracy and precision values, calculated based on the repeated runs of MRG1 standard, are given in Table B.2.

Table B.2 Reported precision and accuracy of standard MRG used for ICP analysis.

	Mean (n=9) wt%	Std Dev	Precision C.V. (%)	Accuracy diff. (%)	MRG Certified wt%
SiO ₂	38.74	0.31	0.80	-0.90	39.09
Al ₂ O ₃	8.43	0.06	0.68	-0.35	8.46
Fe ₂ O ₃	17.69	0.19	1.06	-1.34	17.93
MnO	0.166	0.00	0.76	-2.35	0.17
MgO	13.56	0.12	0.92	0.07	13.55
CaO	14.52	0.17	1.14	-1.29	14.71
Na ₂ O	0.73	0.01	0.82	-1.35	0.74
K ₂ O	0.19	0.02	8.73	5.56	0.18
TiO ₂	3.825	0.04	0.95	1.46	3.77
P ₂ O ₅	0.08	0.00	0.00	0.00	0.08

Table B.2 Reported precision and accuracy of standard MRG used for the ICP-MS analysis.

	Mean (n=1109) ppm	Std Dev	Precision C.V. (%)	Accuracy diff. (%)	MRG Certified ppm	Detection Limits ppm
V	514.700	33.23	6.46	-2.14	526.00	5
Cr	443.000	18.26	4.12	3.15	430.00	10
Co	86.300	4.48	5.19	-0.76	87.00	0.5
Ni	185.000	19.89	10.73	3.97	193.00	10
Cu	123.000	19.53	15.77	7.58	134.00	10
Zn	176.000	22.06	12.48	7.43	191.00	10
Ga	18.600	0.87	4.66	9.43	17.00	1
Ge	1.500	0.46	313.37	45.35	1.00	1
As	0.000	0.00	0.00	0.00	0.73	5
Rb	7.480	0.77	10.25	-12.00	8.50	2
Sr	266.200	14.76	5.55	0.08	266.00	2
Y	13.125	0.81	6.18	-6.25	14.00	0.5
Zr	99.147	9.05	9.13	-8.20	108.00	1
Nb	19.809	2.21	11.15	-0.95	20.00	0.5
Mo	1.220	0.57	46.50	40.71	0.87	0.5
Ag	0.000	0.00	0.00	0.00	0.11	0.5
In	0.000	0.00	0.00	0.00		0.2
Sn	3.581	0.65	18.23	-0.53	3.60	1
Sb	0.770	0.32	37.41	-10.85	0.86	0.1
Cs	0.750	7.64	42.27	31.87	0.57	0.5
Ba	49.153	5.31	10.81	-19.42	61.00	3
La	9.531	0.76	7.93	-2.75	9.80	0.05
Ce	26.504	2.02	7.63	1.94	26.00	0.1
Pr	3.591	0.36	10.15	5.63	3.40	0.02
Nd	18.607	1.09	5.86	-3.09	19.20	0.05
Sm	4.594	0.24	5.25	4.40	4.40	0.02
Eu	1.426	0.08	5.64	2.63	1.39	0.005
Gd	3.998	0.32	8.10	-0.06	4.00	0.02
Tb	0.603	0.04	6.16	18.33	0.51	0.01

Table B.2 (continued)

	Mean (n=1109) ppm	Std Dev	Precision C.V. (%)	Accuracy diff. (%)	MRG Certified ppm	Detection Limits ppm
Dy	2.991	0.15	5.15	3.15	2.90	0.02
Ho	0.501	0.02	4.41	2.15	0.49	0.01
Er	1.265	0.07	5.65	12.95	1.12	0.01
Tm	0.146	0.01	8.23	32.61	0.11	0.005
Yb	0.810	0.09	11.21	34.92	0.60	0.01
Lu	0.114	0.01	7.62	-5.40	0.12	0.002
Hf	3.784	0.24	6.45	0.64	3.76	0.1
Ta	0.841	0.09	10.38	5.14	0.80	0.1
W	0.000	0.00	0.00	0.00	0.30	0.5
Tl	0	0.00	0.00	0.00	0.06	0.1
Pb	5.26	1.36	25.87	-47.70	10.00	5
Bi	0	0.00	0.00	0.00	0.13	0.2
Th	0.890	0.15	16.80	-4.25	0.93	0.05
U	0.283	0.06	19.91	17.80	0.24	0.05

B.4 Thermal Ionization Mass Spectrometry

Radiogenic isotopic analyses were performed using a Finnigan-MAT 261 multicollector solid source mass spectrometer at Carleton University. The mass spectrometer is a multi-collector machine and operates in static mode for Sr, Nd and Pb.

Samples are loaded on either Ta or Re filaments. The loading procedure is described in Table B.3. The conditions for running the samples such as optimum temperature, current and signal intensity are listed in Table B.4.

A standard is run prior to running the other 12 samples on the magazine. Based on numerous runs from September 1992 to February 2000, the following average values have been obtained for standards: NBS-987 $^{87}\text{Sr}/^{86}\text{Sr} = 0.710251 \pm 0.00003$, La Jolla $^{143}\text{Nd}/^{144}\text{Nd} = 0.511870 \pm 0.00003$, NBS 981 $^{206}\text{Pb}/^{204}\text{Pb} = 16.890 \pm 0.010$, $^{207}\text{Pb}/^{204}\text{Pb} = 15.429 \pm 0.013$, and $^{208}\text{Pb}/^{204}\text{Pb} = 36.498 \pm 0.042$. Uncertainties are given at the 2σ level.

Measured Sr and Nd isotopic ratios for the unspiked samples were corrected for fractionation to a $^{88}\text{Sr}/^{86}\text{Sr} = 8.3752$ and $^{146}\text{Nd}/^{144}\text{Nd} = 0.7219$. Sr and Nd isotopic ratios exceeding $\pm 1\%$ of the fractionation correction values (i.e., 8.291 and 8.458 for $^{88}\text{Sr}/^{86}\text{Sr}$ and 0.7147 to 0.7291 for $^{146}\text{Nd}/^{144}\text{Nd}$) were considered as unacceptable. An average fractionation correction of $0.12 \pm 0.01\%$ per mass unit was applied to all measured Pb isotopic ratios based on analysis of NBS 981.

Table B.3 Loading procedure for mass spectrometry analysis.

Element	Filament	Loading acid	Drying current
Rb	double Re	H ₃ PO ₄	1 A
Sr	single Ta	H ₃ PO ₄	1.5 A
Sm	double Re	1N HNO ₃	1.5 A
Nd	double Re	1N HNO ₃	2 A
Pb	single Re	H ₃ PO ₄ + silica gel	1.3 A

Table B.4 Running conditions for mass spectrometry analysis.

Element	Running temperature	Evaporation current	Ionization current	Minimum intensity
Rb	1300 °C	0.0 - 0.2 A	1.5 - 2.5 A	⁸⁷ Rb 0.2 V
Sr	1500 °C	2.5 - 3.5 A		⁸⁸ Sr 1.5 V
Sm	1800 °C	0.0 - 0.5 A	3.5 - 4.5 A	¹⁴⁹ Sm 0.15 V
Nd	1800 °C	0.5 - 1.5 A	4.5 - 5.5 A	¹⁴⁴ Nd 0.35 V
Pb	1300 °C	1.5 - 2.0 A		²⁰⁸ Pb 1.5 V