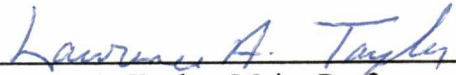


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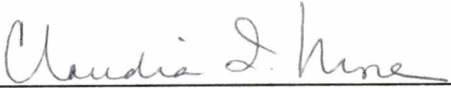
I am submitting herewith a thesis written by Stefan John Higgins entitled "Petrology and Geochemistry of the Early Proterozoic Burakovsky Layered Intrusion, Southern Karelia, Russia." I have examined the final copy of this thesis for form and content and recommend that it be accepted in partial fulfillment of the requirements for the degree of Master of Science, with a major in Geology.



Lawrence A. Taylor, Major Professor

We have read this thesis
and recommend its acceptance:





Accepted for the Council:



Associate Vice Chancellor
and Dean of the Graduate School

**PETROLOGY AND GEOCHEMISTRY OF THE EARLY
PROTEROZOIC BURAKOVSKY LAYERED INTRUSION
SOUTHERN KARELIA, RUSSIA**

A Thesis

Presented for the

Master of Science

Degree

The University of Tennessee, Knoxville

Stefan John Higgins

May, 1996

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ABSTRACT

The early Proterozoic (2449 ± 1.1 Ma) 700 km² Burakovsky layered intrusion of Southern Karelia, Russia is the largest mafic pluton in the Baltic Shield. The intrusion has been divided into a **layered series** and a **border group**. The layered series is further subdivided into five zones: (1) an ultramafic (largely dunites); (2) a pyroxenite; (3) a gabbro-norite; (4) a pigeonite-gabbro-norite; and (5) a magnetite-gabbro-norite. Mineral and whole-rock chemical variations in the ultramafic zone indicate little variation in chemistry over the entire sequence (e.g., in olivine, variations are from Fo₈₉ to Fo₈₂ over the upper 1700 meters). In contrast, the X_{Mg} values for orthopyroxenes from the gabbro-norite zone vary from 0.85 to 0.58 over an interval of 500 meters. On the basis of the current data, there are no apparent reversals in chemistry in the ultramafic zone, while there is evidence for several in the gabbro-norite zone. The monomineralic nature and absence of significant chemical variation in the ultramafic zone indicates a contrasting style of magmatism compared with the gabbro-norite zone, which contains well-defined phase and cryptic layering. Moreover, the lack of evidence for abundant magma influxes in the ultramafic zone contrasts with the chemical reversals observed in the gabbro-norite zone, suggesting that replenishment was a prevailing process in the upper half of the intrusion.

Samples from the ultramafic zone are LREE-enriched indicating that they formed from a LREE-enriched parental magma. The LREE-enrichment in the parental magma suggests either that the source region was LREE-enriched or that the melt was contaminated by crust in route to the magma chamber. Samples from the gabbro-norite zone are also LREE-enriched. Moreover, there are at least two parental magmas for the

gabbro-norite zone delineated by these REE characteristics. It is proposed here that the lower subzone of the gabbro-norite zone be broken into two additional zones, denoted by Group I and Group II magmas. Group I magmas, from the lower part of the gabbro-norite zone, have $(\text{Ce/Yb})_n$ ratios of 6.9 to 13.9 while Group II magmas, from the upper portion, have $(\text{Ce/Yb})_n$ ratios of 15.8 to 27.3. Volcanic rocks in Karelia that are coeval to the BLI and volcanic rocks of a similar age in other parts of the Baltic Shield are LREE-enriched. Furthermore, the BLI has an ϵ_{Nd} value of -2.0, and other layered intrusions in the Baltic Shield of similar age have ϵ_{Nd} values ranging from -2.3 to -2.0. All of these data are most consistent with a mantle that was LREE-enriched for a considerable length of time prior to the partial melting episode(s) which produced the magmatism in the Baltic Shield during the Archean-Proterozoic transition.

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INTRODUCTION

1. INTRODUCTION

Studies of layered mafic intrusions have added immeasurably to our understanding of the fractionation of magmas to form layers of cumulus minerals. Indeed, the well-developed cumulate stratigraphy preserved in these *fossil magma chambers* provides a detailed history of the conditions present in the melt as crystallization proceeded. In some intrusions the cumulate rocks provide evidence of the periodic injections of primitive magma, magma mixing, fractional crystallization, and the assimilation of surrounding country rock.

A relatively small number of layered mafic intrusions have provided the majority of the information available about the complex processes operating in mafic magma chambers. These include such notable localities as the Stillwater Complex in Montana (Hess, 1960); the Bushveld Complex in South Africa (Naldrett et al., 1986); and the Muskox Intrusion in Canada (Irvine and Smith, 1967). Despite the progress made towards understanding igneous fractionation as a result of studies of these intrusions, a considerable amount of information is yet to be discovered, perhaps through detailed studies of little known localities. One such area is the former Soviet Union where a potentially large body of data remains largely inaccessible to western scientists. The current work focuses on understanding the petrogenesis of the largest layered mafic intrusion of the Baltic Shield, the Burakovsky layered intrusion of southern Karelia, Russia.

Layered mafic intrusions of the eastern Baltic Shield (e.g., the Burakovsky, Lukkulaivaara, and Tsipringa intrusions) formed as a result of igneous activity that

occurred between 2.4 and 2.5 Ga. Layered intrusions are often coeval with major phases of rift-related magmatism, and thus, they provide reliable time markers for the various stages of continental rifting. Indeed, the episode of mafic magmatism that produced the layered mafic intrusions in the Baltic Shield is thought to mark the initial stages of continental rifting in this region and the outpouring of voluminous flood basalts.

The objectives of this study are: (1) to characterize the mineralogy, petrography, and structure of the Burakovsky layered intrusion of southern Karelia, Russia; (2) to determine the petrogenesis of the intrusion through careful study of the major- and trace-element geochemistry, and (3) to use these data to interpret the nature and style of magmatism in the Baltic Shield during the early Proterozoic.

This work is presented in two parts. The first part deals primarily with the mineralogy, petrography, structure, and major-element chemistry of a suite of rocks from the Burakovsky intrusion. The second portion of the thesis concentrates on interpreting the trace-element chemistry of these samples. Part I has been submitted, and Part II will be submitted, for publication as:

- (1) "Petrology of the early Proterozoic Burakovsky layered intrusion, southern Karelia, Russia: Part I. Mineral and whole-rock major-element chemistry" (submitted to *The Canadian Journal of Earth Sciences*).
- (2) "Petrology of the early Proterozoic Burakovsky layered intrusion, southern Karelia, Russia: Part II. Trace-element chemistry" (in preparation).

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PART 1

**PETROLOGY OF THE EARLY PROTEROZOIC BURAKOVSKY
LAYERED INTRUSION, SOUTHERN KARELIA, RUSSIA:
MINERAL AND WHOLE-ROCK MAJOR-ELEMENT
COMPOSITIONS**

1. ABSTRACT

The early Proterozoic (2449 ± 1.1 Ma) 700 km² Burakovsky layered intrusion of Southern Karelia, Russia is the largest mafic pluton in the Baltic Shield. The intrusion can be divided into a **layered series** and a **border group**. The layered series contains two main parts: (1) an ultramafic series (85% dunite) and (2) a mafic series (mostly gabbros). Mineral and whole-rock chemical variations in the ultramafic zone indicate little variation in chemistry over the entire sequence (e.g., in olivine, variations are from Fo₈₉ to Fo₈₇ over the upper 1600 meters). In contrast, the X_{Mg} of pyroxenes from the gabbro-norite zone vary from 0.85 to 0.58 over an interval of 500 meters. On the basis of the current data, there are no apparent reversals in chemistry in the ultramafic zone, while there is evidence for several in the gabbro-norite zone. Furthermore, there is a distinctive reversal to more primitive compositions in mineral and whole-rock chemistry at the top of the ultramafic zone. These differences indicate that either (1) the gabbro-norite zone crystallized from the residue left from formation of the ultramafic zone and the subsequent "flushing" of the more evolved portions to form the early Proterozoic Karelian volcanic fields, or (2) there were two separate magmas (i.e., for the ultramafic and gabbro-norite zones). Plagioclase crystals from the gabbro-norite zone are chemically zoned, providing evidence of the local processes affecting crystallization of this sequence. Plagioclase crystals from near the base commonly display relatively Na-rich cores that are believed to result either from supersaturation at the onset of nucleation or the prior crystallization of some other Ca-rich phase (e.g., clinopyroxene). Many plagioclase grains also contain reversely zoned rims, indicative of either (1)

exchange between trapped liquid and the main magma during times of replenishment, or (2) an increase in the CIPW-normative $D^{pl/L}(X_{An})$ partition coefficient due to an increase in the augite component of the melt. The monomineralic nature and absence of significant chemical variation in the ultramafic zone indicates a contrasting style of magmatism compared with the mafic series, which contains well-defined phase and cryptic layering. Moreover, the lack of evidence for abundant magma influxes in the ultramafic zone contrasts with the chemical reversals observed in the gabbro-norite zone, suggesting that replenishment was a prevailing process in the upper half of the intrusion.

2. INTRODUCTION

Much of our current understanding of the processes which operate in mafic magma chambers is a result of field, petrologic, and chemical studies of layered mafic intrusions. These include such notable localities as the Stillwater Complex (Hess, 1960; Jackson, 1961; Raedeke and McCallum, 1984), the Bushveld Complex (Naldrett et al., 1986; Eales et al., 1993), the Skaergaard Intrusion (Wager and Deer, 1939; Wager, 1960; Wager and Brown, 1968), the Kiglapait Intrusion (Morse, 1981; DePaolo, 1985), the Great Dyke of Zimbabwe (Wilson, 1982; Naldrett and Wilson, 1990); and the Muskox Intrusion (Irvine and Smith, 1967; Irvine, 1980a). These intrusions have provided an immense amount of information about the fractional crystallization, differentiation, mixing, and replenishment (by a new influx of magma) of mafic magmas to form layers of cumulus minerals. Layered intrusions of Precambrian age (e.g., Stillwater, Bushveld, The Great Dyke) are particularly important since they contain the

only known concentrations of PGE (Platinum-Group Elements). This suggests either that there may be fundamental differences between the composition of the mantle during the Archean/Proterozoic and today (Naldrett et al., 1990), or that younger, potentially PGE-rich mafic intrusions, have not had an appropriate amount of time to be exposed at the surface by tectonism and/or erosion. Regardless, the study of intrusions older than the Paleozoic provides a window through which the composition of the mantle at this important time in the Earth's history can be viewed. Although the majority of layered mafic intrusions occurring throughout the world (such as those mentioned above) have been well characterized, little is known in the West about those plutons that occur within the boundaries of the former Soviet Union. The need to assess these largely unexplored intrusions, perhaps gaining a greater understanding of the emplacement and differentiation of mafic magmas, has prompted a systematic study of the Burakovsky layered intrusion (BLI), the largest pluton in the Baltic Shield, indeed, in Europe, and located in western Russia.

Layered mafic intrusions occurring in the Baltic Shield (e.g., Burakovsky, Kivakka, Tsipringa, Lukkulajsvaara, Kivach, and Petusjärvi) are thought to have formed as a result of intrusive igneous activity that occurred between 2.4 and 2.5 billion years ago. U-Pb dating of zircons from the BLI yielded an age of 2449 ± 1.1 Ma (Amelin et al., 1995). The lower age of these intrusions is constrained by the Late Archean granites across which they cut (Rundqvist and Mitrofanov, 1993). The emplacement of these intrusions in the Baltic Shield during the Early Proterozoic is considered part of worldwide igneous activity evidenced by layered intrusions and mafic dyke swarms of similar

ages on other cratons. These include the Great Dyke in the Zimbabwe Craton (2461 ± 16 Ma; Wilson and Prendercast, 1989); the Hearst-Matachewan mafic dyke swarm ($2452 \pm 3/-2$), Copper Cliff Formation (2444 Ma), and East Bull Lake intrusion ($2480 \pm 10/-5$) in the Canadian Shield (Heaman, 1988; Krogh et al., 1984; respectively); and the Jimberlana intrusion in the Yilgarn Block in Australia (2420 ± 30 Ma; McClay and Campbell, 1976).

The Burakovsky layered intrusion was first discovered in 1950 through study of a geophysical anomaly (Sharkov et al., 1995). It was not until the work of Garbar et al. (1977), however, that the pluton was defined as a layered mafic intrusion. The initial investigations of the Burakovsky layered intrusion, from 1964-1983, were conducted to determine its economic potential, with a major emphasis on prospecting for copper-nickel, chromium, and titano-magnetite deposits. Beginning in 1984, the Karelian Geological Expedition, under the guidance of V.A. Ganin, N.G. Grinevich, and V.N. Loginov, began an integrated study of the BLI that involved not only prospecting work, but geochemical and petrological investigations as well. In addition, numerous investigations of the intrusion have been performed by geologists from Leningrad and Moscow State Universities, the Leningrad Mining Institute, and the Russian Academy of Science (Sharkov et al., 1995). Despite the significant number of studies, the BLI remains poorly understood, and little information has been disseminated to western scientists. With the present study, we hope to integrate the extensive collection of Russian work with new data on the mineralogy and petrology of this important intrusion.

3. TERMINOLOGY

The applicability of cumulus theory to describe the formation of layered mafic intrusions has been challenged considerably in the literature, most recently by McBirney and Hunter (1995). These authors suggest that many of the rocks in layered intrusions described as cumulates, and thus, formed from cumulus processes, may have been recrystallized. While the cumulate terminology may not adequately describe the processes involved in the formation of layered rocks, it is still useful in describing the textures. In this paper, we use the cumulate terminology of Wager (1960) and Irvine (1982) purely in the descriptive sense and not as a means for describing the origins of any of the rocks of the BLI. Rock nomenclature used in this paper is based upon the recommended classification of the IUGS, summarized in Streckeisen (1976).

4. GENERAL GEOLOGY

There are more than forty layered mafic intrusions known to exist over a widespread area of the northeastern part of the Baltic Shield in Finland, Sweden, and Russia. Most of these intrusions are located along contacts between Early Proterozoic rift zones (Figure 1) and the Archean basement (Balashov et al., 1993). The outcrops of these intrusions vary greatly in size, ranging from less than one square kilometer to several hundred square kilometers (Alapieti et al., 1990). They have been grouped into two extensive belts, a northern belt that trends north-south through the Kola Peninsula, and a southern belt that trends east-west through Finland and northern Karelia (Figure 2).

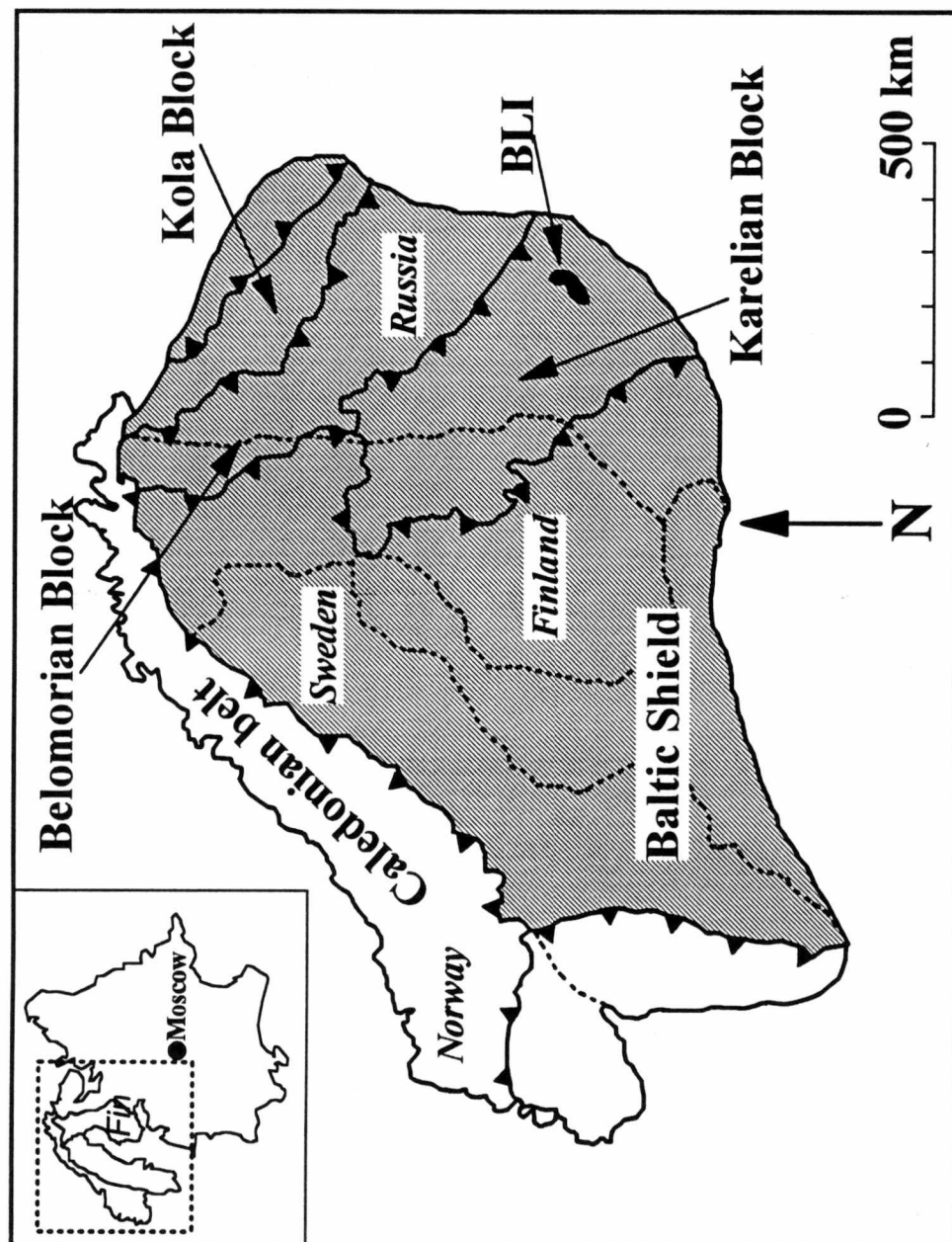


Figure 1. The tectonics of the Baltic Shield, modified after Zonenshain et al. (1990). The countries are outlined by dashed lines. The location of the Burakovsky layered intrusion (BLI) in relation to Moscow is shown.

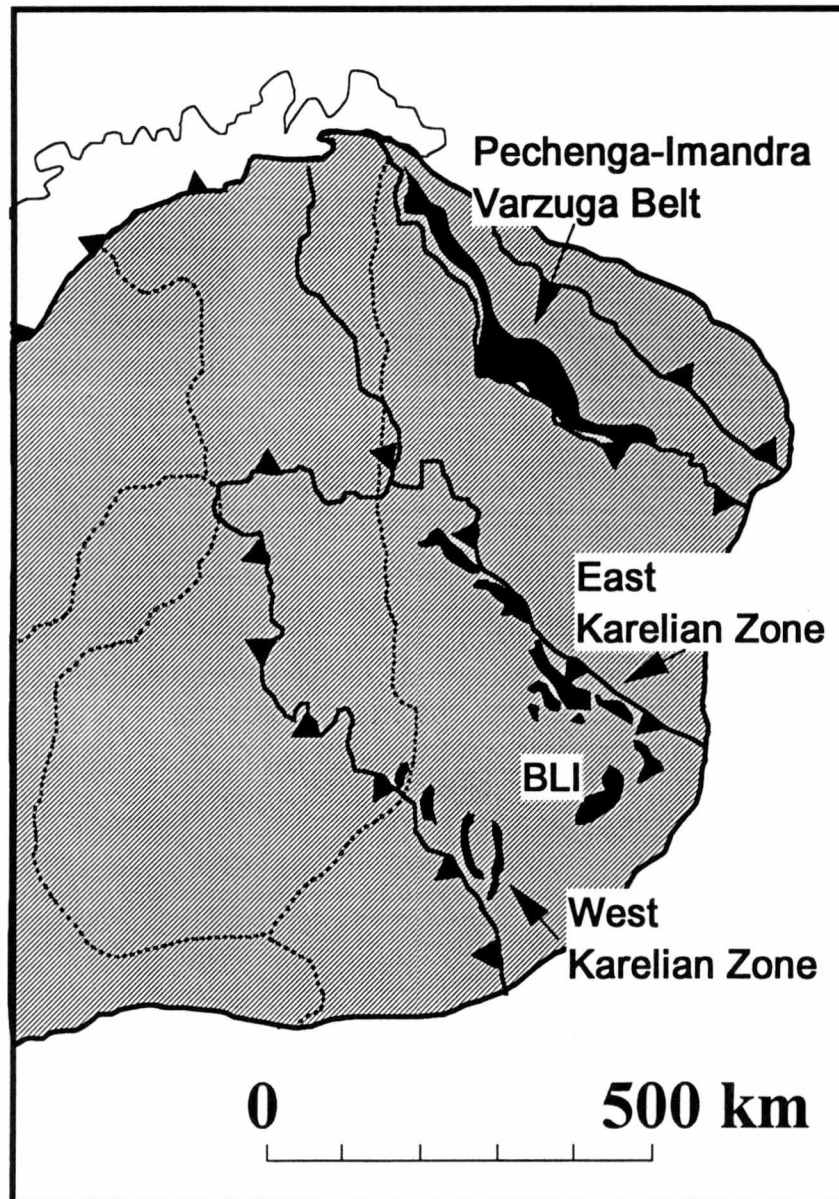


Figure 2. Map of the northeastern part of the Baltic Shield showing the two major belts: the Pechenga-Imandra-Varzuga (northern) and the Karelian (southern). The Burakovsky layered intrusion is also shown. The northern belt is a suture zone between the Kola and Belomorian blocks. The southern belt lies between the Belomorian and Karelian blocks (after Zonenshain et al., 1990).

The Pechenga-Imandra-Varzuga (PIV) supracrustal belt (northern belt) begins in eastern Finland and extends into western Russia (see Figure 2). In Russia, the belt is offset several hundred miles to the southeast before continuing to the eastern edge of the Kola Peninsula. The PIV belt is considered to be a suture zone between two Archean blocks, the Kola and Belamorsky (Zonenshain et al., 1990) (see Figure 1). The PIV supracrustal belt is split into two distinct subbelts, the Pechenga subbelt to the northwest, and 150 km to the southeast, the Imandra-Varzuga subbelt. A chain of Proterozoic age layered intrusions occur along the northeastern periphery of the Imandra-Varzuga subbelt within the Kola block.

The Burakovsky layered intrusion (BLI) occurs in the E-W trending Finno-Karelian supracrustal belt (southern belt). The BLI is contained within the most prominent suture zone on the Baltic Shield, the East-Karelian, which lies between the Karelian and Belomorian blocks (see Figure 2). The East-Karelian suture zone is comprised of the Sumii Formation which is formed by two types of rocks, volcanogenic and clastic (Voitovich, 1980). The Sumii Formation underwent folding and metamorphism during the Pre-Jatulian/Seletsky orogenic phase (2200-2300 Ma; Zonenshain et al., 1990). This folding is connected with the collision of the Belomorsky and Karelian blocks (Zonenshain et al., 1990).

The BLI is the largest mafic pluton on the Baltic Shield, occupying an area of over 700 km², with a thickness that measures between four and six kilometers. Drill core data have clarified the depth of the stratigraphic section to about 5 km (Alapieti et al., 1990). The intrusion is located to the northeast of Lake Onega within an Archean

granite-greenstone terrain and is almost completely overlain by Quaternary glacial sediments (ranging from 10-60 meters in thickness). Consequently, all data that exist about the extent, composition, and internal structure of the intrusion have been determined by geophysical means and from relatively shallow (200-500 m) drill cores (Sharkov et al., 1995). The massif has an irregular oval shape approximately 13-17 km wide and elongated 50 km to the northeast. The BLI has been broken into three distinct blocks by faulting: the Aganozersky, Shalozersky, and Burakovsky blocks. Relative to the Shalozersky block, the Burakovsky and Aganozersky blocks are downthrown and upthrown, respectively. Figure 3 is a geologic map of the BLI showing faults and drill core locations.

The Aganozersky block in the eastern part of the intrusion is formed mainly of ultramafic rocks and has received the most attention by Russian scientists. The block has a funnel-like shape, is slightly tilted to the west, and is sheared by submeridional faults along its margins (Sharkov et al., 1995). The Shalozersky and Burakovsky blocks are comprised primarily of mafic rocks. Geophysical and drill core data indicate that these blocks are shaped like lopoliths. All of the samples analyzed in this study are from the Aganozersky block.

5. STRATIGRAPHY AND PETROLOGY

The BLI is a complexly differentiated series of rocks that range from mafic to ultramafic in composition. Figure 4 is a schematic stratigraphic section of the BLI showing the major zones and subzones. Note the thickness of the Ultramafic Zone

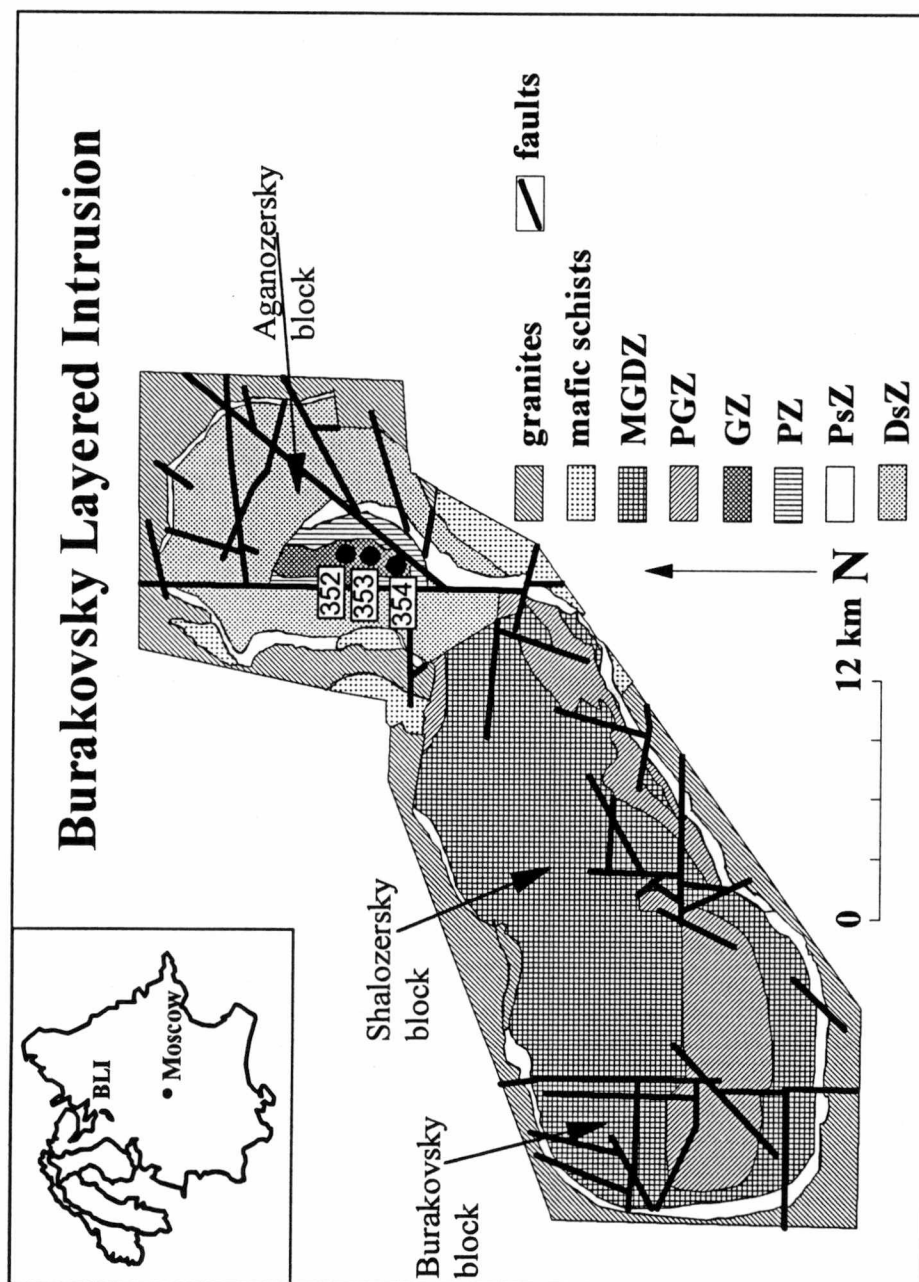


Figure 3. A geologic map of the Burakovsky layered intrusion with major faults shown by the solid lines. MGDZ: magnetite gabbrodiorite zone; PGZ: pigeonite gabbrodiorite zone; GZ: gabbrodiorite zone; PZ: pyroxenite zone; PsZ: peridotite subzone; DsZ: dunite subzone.

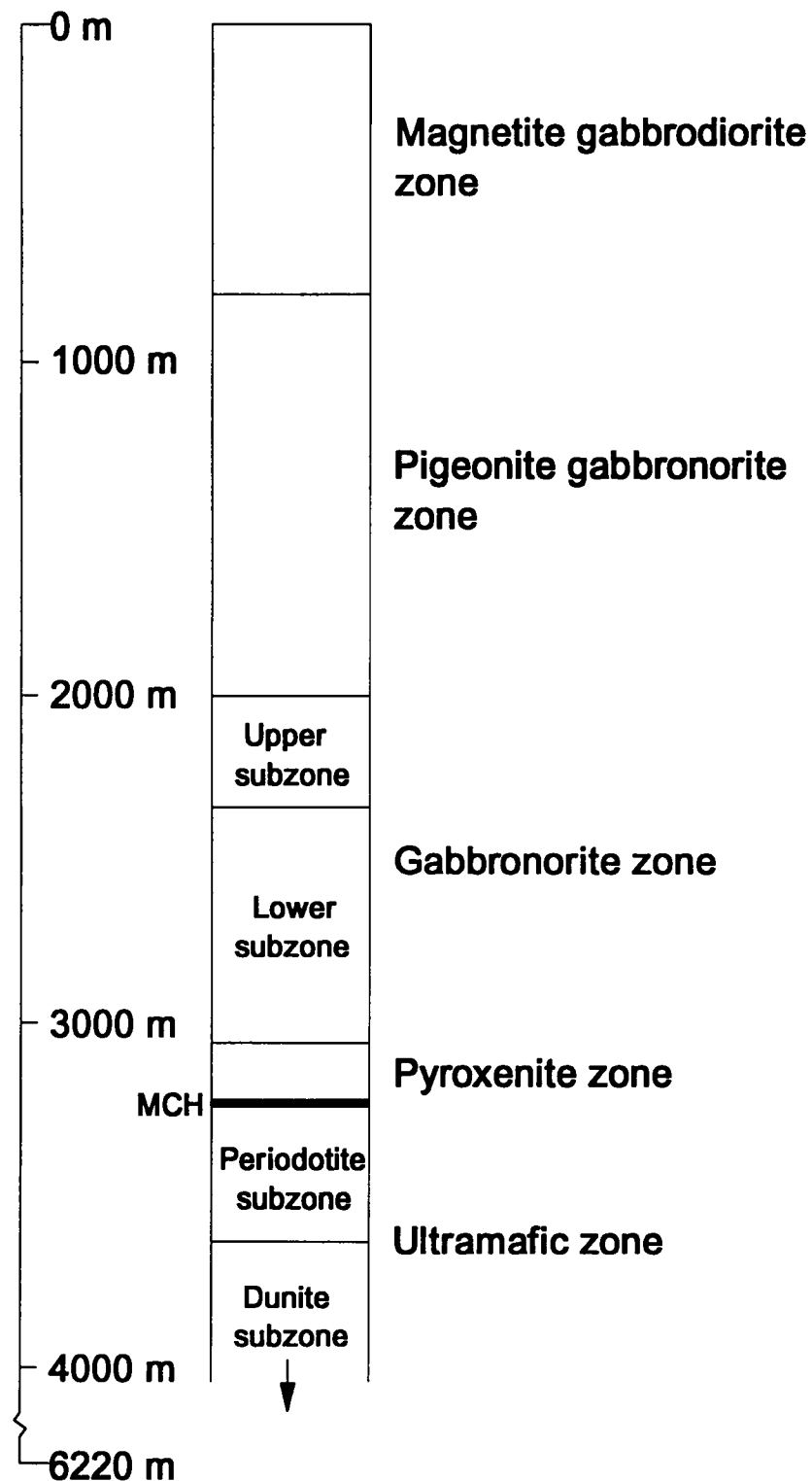


Figure 4. A schematic stratigraphic profile of the layered series of the BLI. Note the thickness of the ultramafic zone relative to the other portions of the layered series. In particular, the lower dunite subzone is 2600 meters thick. MCH = main chromite horizon.

relative to the rest of the intrusion. Stratigraphically, the intrusion can be divided into two major units, the **layered series** and the **border group**.

Layered Series

The Layered Series is broken into five different units, from bottom to top: a 3.0 km thick **ultramafic zone**, a 0.2 km thick **pyroxenite zone**, a 1.1 km thick **gabbronorite zone**, a 1.2 km thick **pigeonitic gabbronorite zone**, and a 0.8 km thick **magnetite gabbro-diorite zone**.

Ultramafic Zone

The ultramafic zone forms the lowest part of the layered series and consists of two subzones: (1) a lower-dunite sequence (olivine $\pm$ chromite) and (2) an upper-peridotite sequence (olivine + cpx $\pm$ chromite and rare chromite cumulates). The division between the upper and lower subzones is based upon the first appearance of cumulus pyroxene in the assemblage. The lower dunite subzone (~ 2600 meters thick) is formed primarily of serpentized dunites whereas the upper peridotite subzone (~ 400 meters thick) consists mostly of poikilitic peridotites. The main intercumulus phases found in the ultramafic zone are low- and high-Ca pyroxenes, with plagioclase, and occasionally phlogopite, occurring more rarely. The percentage of interstitial minerals varies from 5-10% in the dunite subzone and reaches as high as 40% in the peridotite subzone (Sharkov et al., 1995). The main chromite horizon of the BLI occurs at the top of the upper peridotite subzone near the contact with the pyroxenite zone (see Figure 4).

In addition to the main chromite horizon, there are several other chromitite-rich layers (containing 50-75% chromite) contained within the peridotite subzone. These layers range in thickness from 0.5 meters to 4 meters.

Pyroxenite Zone

The stable appearance of orthopyroxene and clinopyroxene cumulates and the disappearance of cumulus olivine marks the top of the ultramafic zone and the beginning of the pyroxenite zone. As mentioned above, this coincides with the main chromite horizon. The pyroxenite zone is composed of orthopyroxene $\pm$ chromite and orthopyroxene + clinopyroxene $\pm$ chromite $\pm$ olivine cumulates. The main rock types, however, are clinopyroxenites that lack cumulus textures. The main characteristic of these rocks is the replacement of earlier cumulus olivine and pyroxene with inverted pigeonite. A transitional layer, 5-10 meters thick, occurs at the base of the pyroxenite zone near the contact with the ultramafic zone. This layer contains peridotite cumulates that have been partially replaced by clinopyroxene aggregates. In addition, many of the clinopyroxene crystals contain oval-shaped inclusions of quartz and calcite 0.1-3 mm in size (Sharkov et al., 1995).

Gabbronorite Zone

The gabbronorite zone occurs in the central part of the Aganozersky block, but it is more widespread in the Burakovsky and Shalozersky blocks. The boundary between the gabbronorite zone and the underlying ultramafic zone is delineated by the stable

appearance of cumulus plagioclase and the upper limit is constrained by the disappearance of cumulus orthopyroxene and the appearance of cumulus pigeonite (Sharkov et al., 1995). The gabbronorite zone has been broken into two portions: a lower subzone (opx, opx + cpx $\pm$ crt, and pl + opx $\pm$ cpx cumulates) and an upper anorthositic subzone (pl + opx + cpx and pl cumulates).

The lower subzone is approximately 600 meters thick and is characterized by alternating websterites (cpx + opx $\pm$ ol $\pm$ crt), pyroxenites (cpx or opx $\pm$ crt), norites (pl + opx), gabbronorites (pl + opx + cpx), gabbros (pl + cpx), and anorthosites (pl). From bottom to top, the thickness of the gabbronorite layers increases with a gradual decrease in the orthopyroxene content. There are two "marking horizons", one, a 2-3 meter thick dunite-harzburgite layer, located near the bottom of the subzone at the base of the first sequence, and the other, a 2-3 meter thick layer of chromite-bearing bronzitites and websterites located in the upper part of the subzone. The upper anorthositic subzone consists of approximately 320 meters of interlayered gabbronorites and anorthosites.

Pigeonite Gabbronorite Zone

The pigeonite gabbronorite zone occurs in the Shalozersky and Burakovsky blocks and consists of gabbronorites containing cumulus inverted pigeonites, pigeonites, and augites. This zone has a poorly-defined layering composed of alternating leucocratic and mesocratic gabbronorites that vary in thickness from 1.5-10 meters. Veins of microcline-rich granite, 0.5-10 meters in thickness, often cut through the gabbronorite (Sharkov et al., 1995).

Magnetite Gabbro-Diorite Zone

The magnetite gabbro-diorite zone is the uppermost layer in the intrusion, occurring at the top of the layered series. It is located in the northern parts of the Shalozersky and Burakovsky blocks and is composed of Ti-magnetite + plagioclase + pigeonite + Cpx cumulates. The base of the magnetite gabbro-diorite zone is marked by a titanomagnetite-rich (up to 30 vol.%) gabbro layer, but the most abundant rock type is magnetite-gabbro-diorite. This rock consists of cumulus inverted pigeonite (10-30%), pigeonite (10-20%), titanomagnetite (3-12%), and plagioclase (40-75%). The plagioclase is oligoclase to andesine in composition. Apatite can be found, but is rare. In addition, minor amounts of biotite and quartz occur as interstitial material. A poorly defined layering is present, consisting of alternating mesocratic and melanocratic rock types, with 10-12% Ti-magnetite and 1-5% Ti-magnetite, respectively, is present (Sharkov et al., 1995).

Border Group

The **border group** occurs along the margin of the intrusion as a band 200-400 meters in thickness. It can be divided into an outer chilled zone and an inner zone. The stratigraphy of the border group is opposite to that of the layered series, with fine-grained mafic rocks of the chilled zone in contact with wall rocks, grading inward into pyroxenites and peridotites of the inner zone.

6. SAMPLING AND ANALYTICAL METHODS

Individual mineral and whole-rock analyses were performed on a host of samples taken from several drill cores located in the Aganozersky block of the BLI. Three samples of the ultramafic zone were analyzed, all from the upper peridotite subzone (drill cores 200, 333, and 248, respectively). Twenty-five samples of the gabbro-norite zone, taken from drill cores 353 and 354, were analyzed. These cores penetrate the gabbro-norite zone to a maximum depth of 473 meters (Figure 5). Samples are labeled using core number and depth in meters (e.g., core #/depth). In addition to these samples of the layered series, we have analyzed one sample from the border zone of the intrusion.

We are currently in the process of obtaining samples from the pyroxenite zone of the BLI. Results from analyses of these samples will be presented in a follow-up to this paper.

Mineral chemical compositions were determined with a fully automated CAMECA SX-50 electron microprobe at the University of Tennessee, Knoxville. Analytical conditions employed an accelerating voltage of 15 kV, a beam current of 30 nA for pyroxenes and olivines and 20 nA for feldspars, 20 seconds counting time for each element, and full ZAF correction. A beam size of 5 μm was used for all mineral analyses.

The presence of fine exsolution lamellae in both clinopyroxene (10-20 μm) and orthopyroxene (5-10 μm) crystals necessitated special care in order to obtain representative and reconstructed whole grain compositions. Microprobe traverses were conducted across individual grains and the analyses averaged to determine a bulk

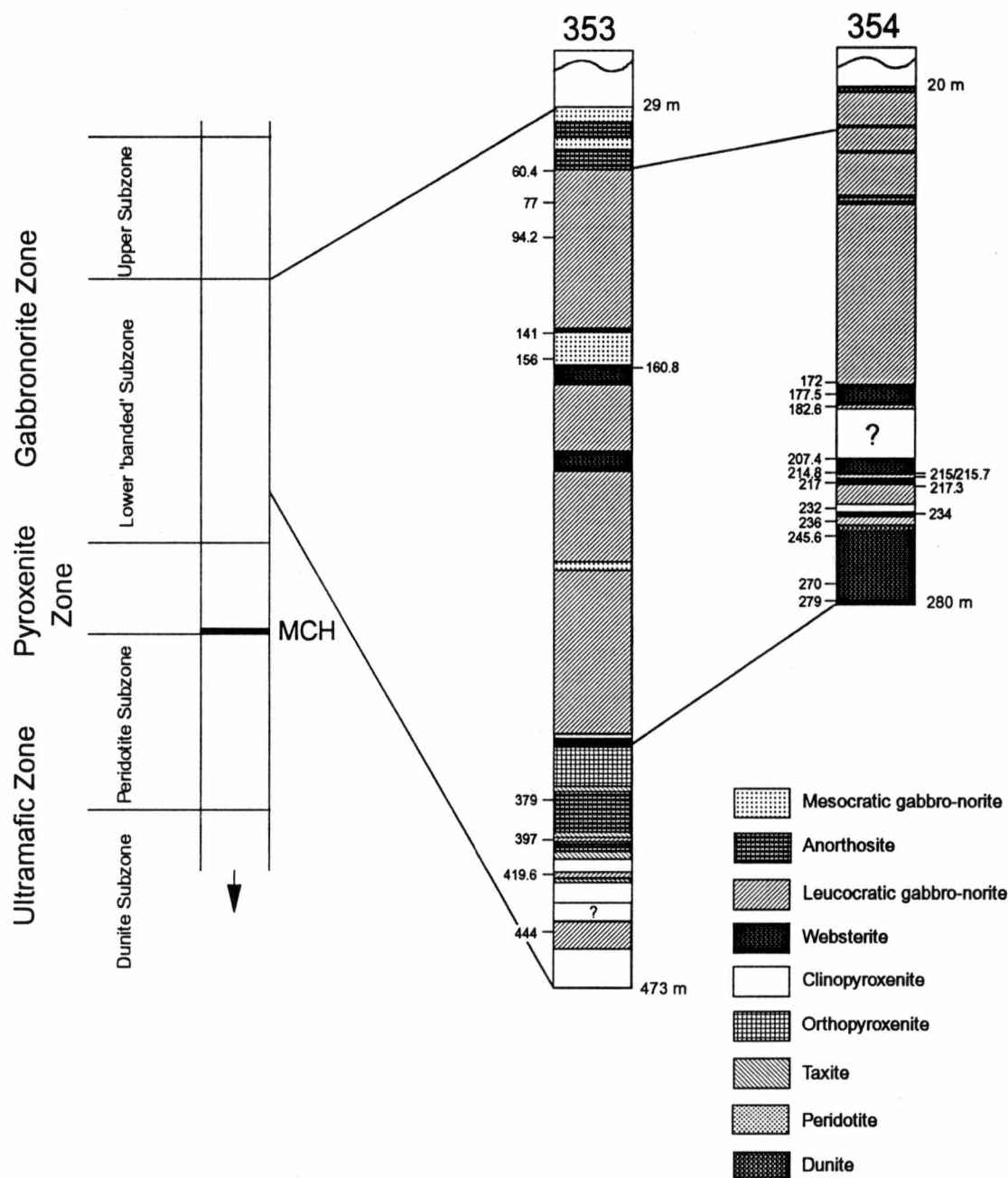


Figure 5. Schematic representations of lithologic units identified in cores 353 and 354 which penetrate the gabbro-norite zone of the BLI. The location of each sample is noted beside the column. A rough correlation of the two columns is also shown.

composition. To confirm the validity of this method, X-ray mapping, with an energy dispersive spectrometer mounted on the electron microprobe, was conducted on several grains. The results were remarkably similar to those obtained with mineral traverses, proving the effectiveness of this technique for determining bulk compositions of exsolved crystals.

Major and trace-element (Rb, Sr, V, Cr, Ni, Nb, Y, Zr) abundances for whole-rocks were determined at the University of Tennessee, using an EG & G Ortec TEFA energy-dispersive X-ray fluorescence spectrometer, with an internal standard checking process. Detection limits for trace elements were 5 ppm for Rb, 17 ppm for Sr, 46 ppm for V, 61 ppm for Cr, 33 ppm for Ni, 3 ppm for Nb, 4 ppm for Y, and 22 ppm for Zr. Four to five gram sample splits were made into pressed pellets utilizing a shatterbox, pellet die, and press. Each sample was ground in the shatterbox with a tungsten carbide mill for six minutes.

7. PETROGRAPHY AND MINERAL CHEMICAL COMPOSITIONS

Dunites

Three samples representing dunite layers were analyzed in this study. Of these samples, two were from the peridotite subzone of the ultramafic zone and one was from the border group (333/496, 200/444, and 28/223, respectively). Samples 333/496 and 28/223 are mesocumulates, while sample 200/444 is an orthocumulate. Cumulus phases present are olivine and chromite. Olivine is the primary phase and generally occurs as euhedral to subhedral crystals 1-2 mm in width. Table 1 reports major-element data for

Table 1. Representative major-element compositions for olivine crystals from the ultramafic and border zones of the BLI.

	28/223.8 ¹	200/444.7 ²	333/496.5 ²	248/190
SiO ₂	39.2 (2) ³	40.3 (2)	39.4 (1)	39.5 (2)
FeO	19.4 (3)	12.6 (1)	17.1 (2)	16.7 (2)
MnO	0.27(4)	0.17(3)	0.22(3)	0.21(3)
MgO	41.4 (2)	47.1 (2)	43.3 (2)	43.7 (2)
CaO	0.04(1)	0.04(1)	0.05(1)	0.03(1)
NiO	0.41(4)	0.54(3)	0.37(4)	0.40(3)
Total	100.7	100.8	100.4	100.5
Cations based on 4 oxygens				
Si	0.999	0.994	0.997	0.997
Fe	0.414	0.261	0.361	0.352
Mn	0.006	0.004	0.005	0.004
Mg	1.573	1.734	1.631	1.641
Ca	0.001	0.001	0.001	0.001
Ni	0.008	0.011	0.007	0.008
Total	3.001	3.005	3.002	3.003
Fo	79.2	86.9	81.9	82.3

1. Sample from the border group; first number = drill core, second = depth.
2. Samples from the upper subzone of the ultramafic zone.
3. Units in parentheses represent one standard deviation of replicate analyses of last digit cited.

olivine from these samples. Olivine compositions range from Fo 89 to Fo 82 over the upper 1700 meters of the ultramafic zone (Figure 6). The olivine is fractured and, in most cases, has undergone some alteration to serpentine. In a few of the samples, only relict olivine crystals remain, the bulk of the mineral having been entirely serpentinized. Chromite crystals, ranging from 0.08 to 0.2 mm in width, are mostly euhedral in form and generally occur within the intercumulus material. The Cr/(Cr+Al) varies between 0.67 and 0.80 for these samples. Clinopyroxene is the primary intercumulus phase in all samples and surrounds the olivine and chromite to form a poikilitic texture. Major-element compositions of clinopyroxene reveal variations from $\text{Wo}_{42.9}\text{En}_{51.1}\text{Fs}_{6.0}$ to $\text{Wo}_{40.0}\text{En}_{51.1}\text{Fs}_{8.9}$ upward in the peridotite subzone. In addition to clinopyroxene, plagioclase and occasionally phlogopite occur as interstitial minerals. The post-cumulus phases make up 5-10% of the mesocumulates and 25% of the orthocumulate. Figure 7 is a photomicrograph of a typical olivine and chromite crystal from the peridotite subzone.

Peridotites

One peridotite from the peridotite subzone (sample 248/190) was analyzed in this study. This rock is composed almost entirely of cumulus clinopyroxene and olivine, with minor secondary serpentine present as an alteration product of the olivine. On the basis of the IUGS classification (Streckheisen, 1976), the sample is defined as a wehlite. Olivine compositions vary from Fo 84 to Fo 82 over the 400 meter thick Peridotite

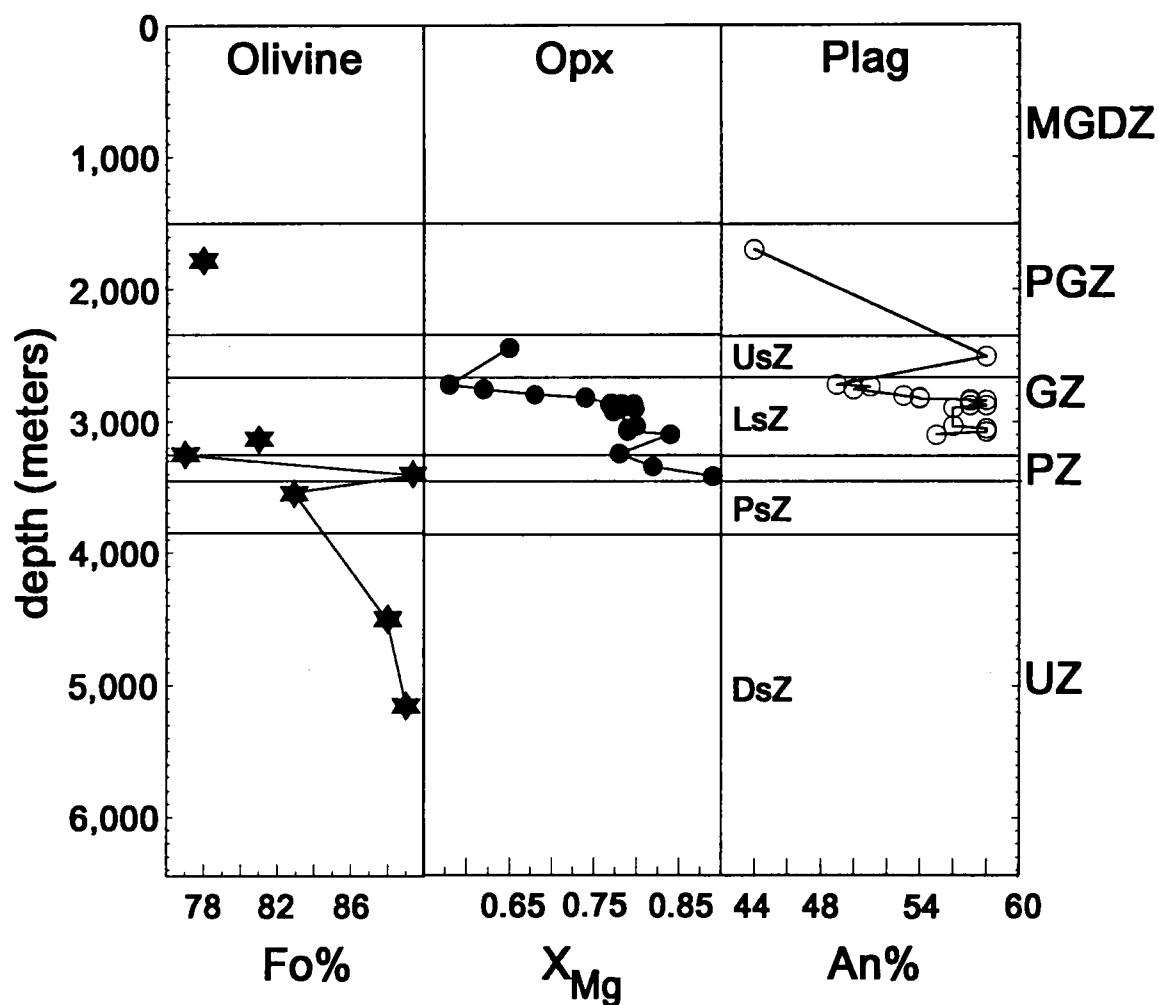


Figure 6. Variation in major-element compositions with stratigraphic height for olivine, orthopyroxene, and plagioclase. Plotted data are from this study and unpublished Russian analyses. Note that there are a number of reversals in magma composition within the gabbrozone. MGDZ: magnetite gabbrozone; PGZ: pigeonite gabbrozone; GZ: gabbrozone; PZ: pyroxenite zone; UZ: ultramafic zone.

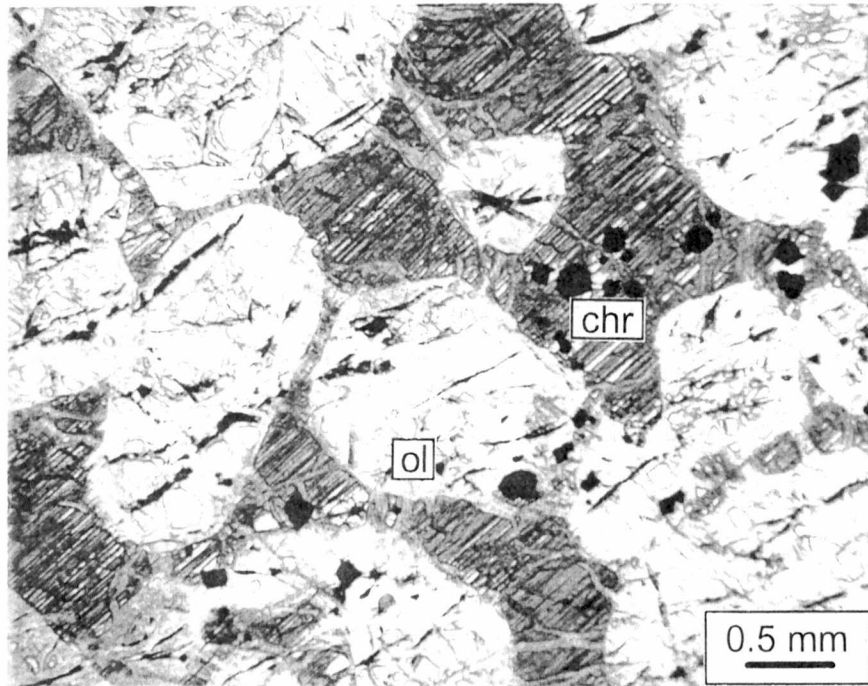


Figure 7. A transmitted light photomicrograph of a typical dunite (200/444) from the ultramafic zone. Olivine and chromite are shown. Note the euhedral form of both minerals.

subzone (see Figure 6), with olivine in sample 248/190 having an Fo value of 82.3 (see Table 1). Clinopyroxene occurs as 0.6 to 1.2 mm subhedral to euhedral crystals, with numerous parallel exsolution lamellae (generally 5-25 μm in width) and has a composition of $\text{Wo}_{40.7}\text{En}_{50.6}\text{Fs}_{8.71}$. Surrounding the cumulus phases is a matrix consisting largely of alteration products (i.e., serpentine and chlorite with minor phlogopite). Figure 8 is a photomicrograph showing the cumulus phases and textural relationships observed in a typical wehrlite from the ultramafic zone.

Gabbros

Several gabbros from drill cores 353 and 354 were analyzed in this study. These include seven samples from drill core 354 and ten samples from drill core 353. Petrographically, these samples show similar characteristics. All are coarse grained cumulates with no distinctive intercumulus phases observed petrographically, but chemically, there is evidence of post-cumulus growth (e.g., particularly in plagioclase). Clinopyroxene (0.6-2.5 mm in size) and plagioclase (0.5-3.0 mm in size) are present as cumulus phases in varying proportions. Moreover, many of the samples contain orthopyroxene (1.0 to 3.0 mm in size) as an additional cumulus phase. The plagioclase crystals are subhedral to euhedral in form and often lath-shaped.

Major-element compositions of cumulus pyroxenes from gabbro layers of the gabbro-norite zone are given in Tables 2 and 3. The X_{Mg} values of orthopyroxene decrease from 0.84 to 0.58 from bottom to top of the gabbro-norite zone. Orthopyroxene ranges from hypersthene to bronzite in composition and clinopyroxene

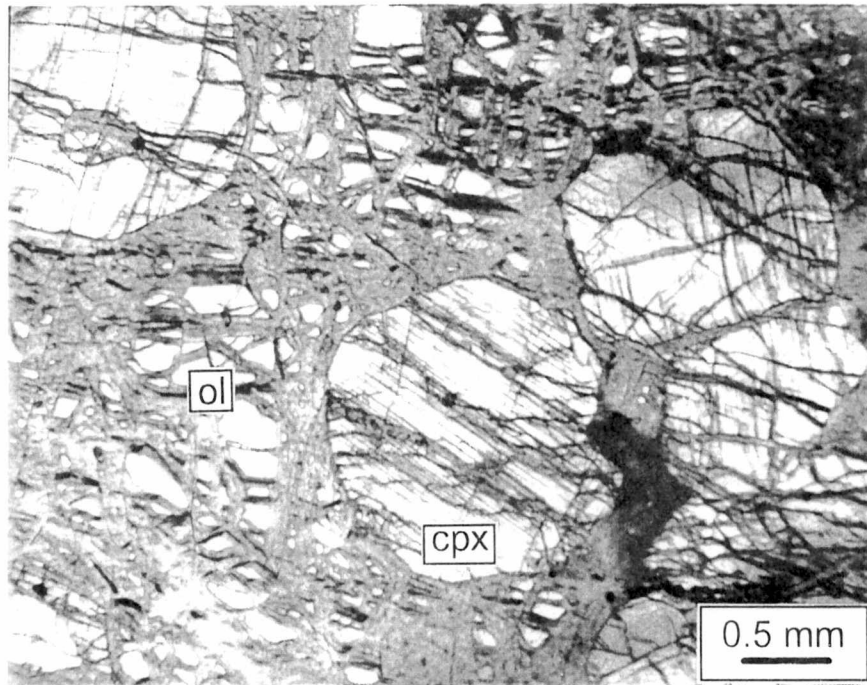


Figure 8. A transmitted light photomicrograph of a typical wehrlite (248/190) from the ultramafic zone. Cumulus phases shown are clinopyroxene and olivine.

Table 2. Reconstructed major element compositions of selected orthopyroxenes from the gabbro zone of the BLI

	353/60	353/94	353/161	354/214.8	354/215.7	354/217.3	354/234	354/245.6	354/270.2	353/444
	[7]	[23]	[16]	[20]	[15]	[15]	[19]	[20]	[16]	[32]
SiO ₂	51.1 (2)	52.2 (3)	53.8 (2)	54.9 (1)	54.4 (1)	54.3 (2)	53.8 (2)	54.3 (1)	54.0 (3)	55.5 (2)
TiO ₂	0.34(7)	0.31(8)	0.20(3)	0.13(2)	0.16(3)	0.18(2)	0.14(2)	0.14(3)	0.16(3)	0.16(1)
Al ₂ O ₃	0.85(7)	1.02(7)	1.04(2)	1.33(3)	1.32(3)	1.35(5)	1.50(6)	1.49(7)	1.38(10)	1.31(13)
Cr ₂ O ₃	<0.03	0.14(4)	0.03(2)	0.16(3)	0.15(3)	0.15(3)	0.40(3)	0.44(5)	0.26(3)	0.06(2)
FeO	25.5 (4)	20.8 (4)	16.7 (2)	12.9 (2)	14.3 (3)	14.6 (5)	13.6 (2)	12.7 (2)	14.3 (4)	10.1 (2)
MnO	0.48(9)	0.40(3)	0.33(3)	0.25(3)	0.28(3)	0.29(2)	0.25(3)	0.25(3)	0.28(4)	0.22(4)
MgO	19.5 (2)	22.8 (3)	26.0 (2)	28.5 (2)	27.6 (2)	27.3 (4)	27.6 (2)	28.3 (3)	27.2 (5)	29.9 (2)
CaO	1.50(44)	1.63(37)	2.03(28)	1.90(27)	1.85(31)	1.76(64)	2.04(28)	2.14(4)	2.10(75)	2.07(34)
Na ₂ O	<0.03	<0.03	<0.03	0.05(2)	<0.03	0.03 (2)	0.04(1)	0.07(2)	0.05(3)	<0.03
Total	99.4	100.2	100.1	100.2	100.1	99.9	99.4	99.8	99.7	99.4

Cations based on 6 oxygens

Si	1.956	1.955	1.956	1.958	1.957	1.957	1.946	1.948	1.953	1.970
Ti	0.010	0.009	0.006	0.004	0.004	0.005	0.004	0.004	0.004	0.004
Al ^{IV}	0.039	0.044	0.044	0.042	0.043	0.043	0.054	0.052	0.047	0.030
Al ^{VI}	0.000	0.000	0.000	0.014	0.013	0.014	0.010	0.011	0.012	0.025
Cr	0.001	0.004	0.001	0.005	0.004	0.004	0.011	0.013	0.007	0.002
Fe	0.816	0.650	0.508	0.386	0.431	0.441	0.410	0.380	0.431	0.301
Mn	0.016	0.013	0.010	0.008	0.008	0.009	0.007	0.007	0.008	0.007
Mg	1.115	1.270	1.411	1.518	1.476	1.465	1.489	1.512	1.465	1.581
Ca	0.062	0.065	0.080	0.073	0.071	0.068	0.079	0.082	0.081	0.079
Na	0.002	0.002	0.002	0.003	0.001	0.002	0.003	0.005	0.004	0.301
Total	4.016	4.012	4.020	4.009	4.009	4.008	4.014	4.013	4.012	3.998
X _{Mg}	0.58	0.66	0.74	0.80	0.77	0.77	0.78	0.80	0.77	0.84

1. Number in [] is the number of electron microprobe analyses.

2. Number in () is the standard deviation in terms of least units cited, as determined by replicate analyses. The value for many of the elements reflects exsolution which was accounted for by doing mineral traverses across selected grains and averaging the data.

3. X_{Mg} = molar Mg/(Mg + Fe)

Table 3. Reconstructed major element compositions for selected clinopyroxenes from the gabbro zone.

	353/60	353/94	353/161	354/172	354/177.5	354/214.8	354/215.7	354/217.3	354/232	354/234	354/245.6	354/279	353/444
	[25]	[12]	[21]	[20]	[23]	[30]	[10]	[15]	[20]	[25]	[20]	[25]	[25]
SiO ₂	50.8 (4)	51.1 (3)	52.2 (5)	52.1 (3)	51.9 (5)	53.1 (3)	52.3 (1)	52.5 (2)	52.6 (3)	52.4 (5)	51.9 (3)	52.9 (4)	53.1 (3)
TiO ₂	0.50(13)	0.62(19)	0.49(3)	0.38(4)	0.47(7)	0.24(3)	0.34(3)	0.32(4)	0.35(5)	0.22(4)	0.36(11)	0.24(3)	0.26(4)
Al ₂ O ₃	1.85(28)	1.86(14)	1.89(23)	2.12(9)	2.25(25)	2.24(17)	2.21(6)	2.26(12)	2.23(19)	2.28(21)	2.43(21)	2.35(17)	2.10(15)
Cr ₂ O ₃	0.08(3)	0.30(5)	0.06(2)	0.13(4)	0.11(3)	0.34(6)	0.28(4)	0.28(4)	0.70(8)	0.76(11)	0.92(15)	0.50(6)	0.10(3)
FeO	13.0 (26)	10.2 (10)	7.61(48)	7.02(21)	6.75(18)	7.55(147)	6.73(32)	6.89(74)	7.96(169)	7.88(235)	7.05(138)	6.57(145)	5.33(55)
MnO	0.29(5)	0.23(2)	0.20(3)	0.16(2)	0.16(3)	0.18(3)	0.17(3)	0.18(3)	0.18(3)	0.18(4)	0.17(3)	0.17(3)	0.15(3)
MgO	13.8 (10)	14.4 (4)	15.7 (4)	15.7 (2)	15.6 (4)	17.3 (18)	16.0 (3)	16.1 (8)	16.9 (21)	17.8 (31)	16.9 (20)	18.1 (24)	17.6 (10)
CaO	18.4 (34)	20.0 (13)	21.3 (8)	21.1 (5)	21.3 (7)	18.8 (33)	21.4 (6)	21.0 (16)	18.3 (37)	17.5 (55)	18.7 (32)	18.5 (40)	20.2 (17)
Na ₂ O	0.31(6)	0.42(4)	0.34(2)	0.33(2)	0.34(3)	0.35(7)	0.31(2)	0.33(3)	0.50(11)	0.41(14)	0.59(12)	0.33(7)	0.26(4)
Total	99.0	99.2	99.8	99.0	98.9	100.1	99.7	99.9	99.8	99.4	98.9	99.7	99.1
Cations based on 6 oxygens													
Si	1.936	1.930	1.937	1.941	1.935	1.947	1.935	1.938	1.941	1.936	1.928	1.939	1.953
Ti	0.014	0.018	0.014	0.011	0.013	0.007	0.010	0.009	0.010	0.006	0.010	0.007	0.007
Al ^{IV}	0.064	0.070	0.063	0.059	0.065	0.053	0.065	0.062	0.059	0.064	0.072	0.061	0.047
Al ^{VI}	0.009	0.013	0.020	0.034	0.034	0.043	0.031	0.036	0.038	0.035	0.034	0.041	0.043
Cr	0.002	0.009	0.002	0.004	0.003	0.010	0.008	0.008	0.020	0.022	0.027	0.015	0.003
Fe	0.415	0.323	0.236	0.219	0.211	0.232	0.208	0.213	0.245	0.243	0.219	0.201	0.164
Mn	0.009	0.007	0.006	0.005	0.005	0.006	0.005	0.006	0.006	0.006	0.005	0.005	0.005
Mg	0.782	0.810	0.870	0.873	0.870	0.945	0.883	0.886	0.930	0.979	0.935	0.989	0.964
Ca	0.753	0.810	0.848	0.842	0.852	0.739	0.847	0.831	0.724	0.692	0.743	0.727	0.796
Na	0.023	0.031	0.025	0.024	0.025	0.025	0.022	0.024	0.036	0.030	0.043	0.023	0.019
Total	4.018	4.021	4.020	4.012	4.013	4.005	4.015	4.012	4.008	4.012	4.016	4.008	4.002
Wo	38.4	41.5	43.2	43.4	44.0	38.3	43.6	42.9	38.1	36.1	39.1	37.9	41.3
En	39.9	41.5	44.4	45.0	44.9	49.2	45.4	45.8	48.8	51.0	49.1	51.4	50.0
Fs	21.7	17.0	12.4	11.6	11.1	12.5	11.0	11.3	13.1	12.9	11.8	10.7	8.74
X _{Fe}	0.65	0.72	0.79	0.80	0.81	0.80	0.81	0.81	0.79	0.80	0.81	0.83	0.86

1. Numbers in [] refer to number of electron microprobe analyses for each sample.

2. Units in O represent one standard deviation of replicate analyses in terms of last digit cited.

lamellae are common. Clinopyroxenes are augitic in composition, ranging from $\text{Wo}_{41.3}\text{En}_{50.0}\text{Fs}_{8.74}$ to $\text{Wo}_{38.4}\text{En}_{39.9}\text{Fs}_{21.7}$ towards the top, and generally contain exsolution lamellae of orthopyroxene. Figure 9a and 9b are photomicrographs of typical pyroxenes from the gabbronorite zone. Note the distinctive parallel exsolution lamellae in the clinopyroxene crystal (Figure 9b). Figure 10 plots pyroxene compositions from this study on the pyroxene quadrilateral. For comparison, the trend of crystallization in the Bushveld Complex is also shown (Wager and Brown, 1967; Atkins, 1969).

Compositions of plagioclase crystals from different stratigraphic levels of the gabbronorite zone are given in Tables 4 and 5. Many of the plagioclase crystals are strongly zoned. Figure 11 is a photomicrograph showing typical cumulus plagioclase crystals from the gabbronorite zone. Figures 12a-12f are plots of mol.% An versus distance (μm) for the plagioclase compositions given in Tables 4 and 5.

In general, plagioclase crystals near the base of the gabbronorite zone (e.g., 353/444) have small (~ 0.4 mm) sodic cores and large ($\sim 0.8 - 1$ mm) calcic rims (Table 4; Figure 12a). Plagioclase crystals from the middle portions of the gabbronorite zone (e.g., 354/215.7, Figure 12e) display small ($100-200 \mu\text{m}$) cores (e.g., An_{58}) and thin ($50-100 \mu\text{m}$) rims. In addition, these crystals often have reversely zoned rims, as evidenced by increases in An%. For example, in sample 354/215.7, the variation is from An_{47} to An_{57} (Table 4 and Figure 12e). Plagioclase from the upper portions of the gabbronorite zone (e.g., sample 353/77) have large ($400-600 \mu\text{m}$) calcic cores and normally zoned rims (e.g., An_{53} and An_{49} , respectively).

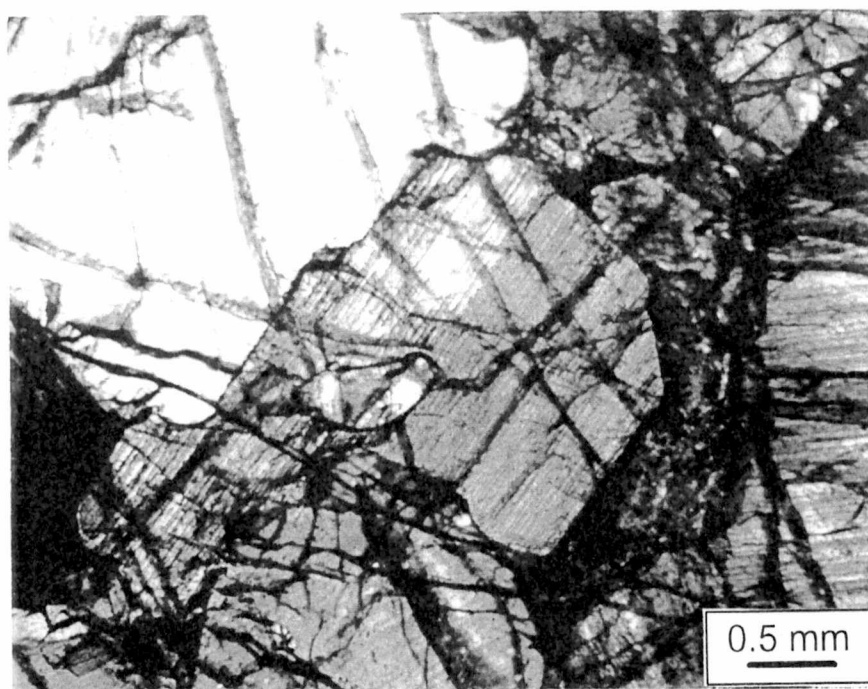
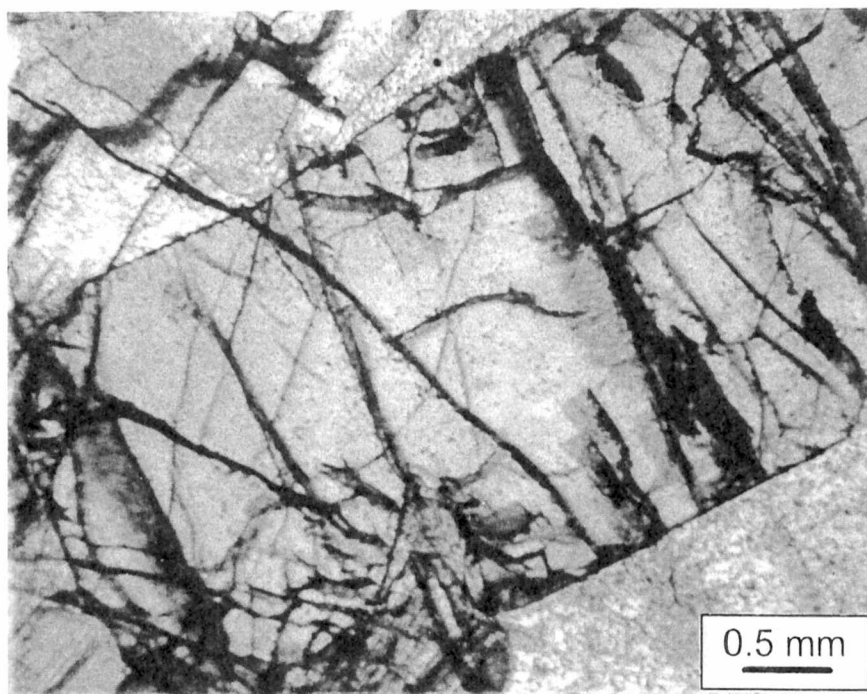


Figure 9. Photomicrographs of typical pyroxenes from the gabbro-norite zone. [A] an orthopyroxene crystal taken in transmitted light, [B] a clinopyroxene crystal, showing exsolution, taken in polarized light.

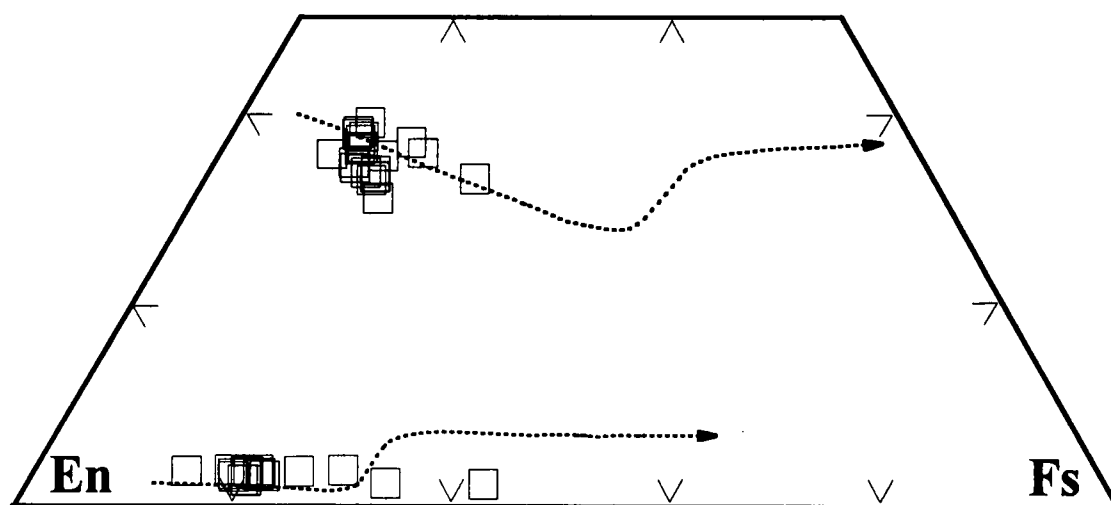


Figure 10. Crystallization trend of Ca-rich and Ca-poor pyroxenes in the gabbro-norite zone of the BLI. The trend of Bushveld pyroxenes is shown for comparison, represented by dashed lines (after Wager and Brown, 1967).

Table 4. Representative plagioclase major-element compositions for core 354, which penetrates the gabbro norite zone.

354/172			354/177			354/215.7		
	<u>core</u>	<u>rim</u>	<u>m'</u>	<u>core</u>	<u>rim</u>	<u>core</u>	<u>rim</u>	<u>m'</u>
SiO ₂	53.3	53.7	55.0	53.5	53.9	52.4	53.1	55.8
Al ₂ O ₃	28.7	28.8	27.8	28.2	28.3	29.0	28.7	26.8
FeO	0.35	0.37	0.36	0.40	0.36	0.36	0.36	0.28
CaO	11.5	11.4	10.5	11.4	11.2	11.9	11.6	9.60
Na ₂ O	4.71	4.81	5.31	4.90	4.89	4.65	4.69	5.77
K ₂ O	0.15	0.23	0.23	0.17	0.22	0.16	0.20	0.29
Total	98.7	99.4	99.2	98.6	98.8	98.5	98.6	98.5
Cations based on 8 oxygens								
Si	2.442	2.445	2.499	2.455	2.464	2.411	2.437	2.547
Al	1.549	1.546	1.491	1.528	1.524	1.576	1.551	1.442
Fe	0.014	0.014	0.014	0.015	0.014	0.014	0.014	0.011
Ca	0.566	0.558	0.510	0.560	0.548	0.588	0.571	0.470
Na	0.419	0.425	0.468	0.436	0.434	0.415	0.418	0.511
K	0.009	0.013	0.014	0.010	0.013	0.009	0.012	0.017
Total	4.999	5.001	4.996	5.004	4.997	5.013	5.003	4.998
An%	57.0	56.0	51.4	55.6	55.1	58.1	57.1	47.1

1. Unusually sodic region located in the margin of the crystal just inside the rim.

Table 5. Representative major-element plagioclase compositions for core 353, which penetrates the gabbro norite zone.

353/77				353/161				353/444	
	core [12]	rim	core [12]	rim	m ²	mx ³	core	rim	
SiO ₂	54.6 (2)	55.5	53.6 (2)	54.2	54.9	53.5	57.8	53.1	
Al ₂ O ₃	28.1 (2)	27.3	28.5 (1)	27.8	27.5	28.5	26.1	29.1	
FeO	0.30(5)	0.33	0.37(14)	0.26	0.26	0.33	0.53	0.30	
CaO	10.9 (1)	10.0	11.5 (1)	10.9	10.5	11.8	8.38	11.9	
Na ₂ O	5.15(7)	5.59	4.71(7)	5.18	5.30	4.60	6.30	4.57	
K ₂ O	0.40(2)	0.44	0.46(6)	0.50	0.54	0.43	0.18	0.12	
Total	99.4	99.2	99.1	98.9	99.0	99.1	99.3	99.1	
Cations based on 8 oxygens									
Si	2.480	2.524	2.449	2.481	2.504	2.445	2.608	2.424	
Al	1.506	1.461	1.534	1.501	1.479	1.537	1.386	1.567	
Fe	0.011	0.013	0.014	0.010	0.010	0.013	0.020	0.011	
Ca	0.530	0.489	0.564	0.533	0.513	0.576	0.405	0.585	
Na	0.454	0.493	0.418	0.459	0.469	0.408	0.551	0.405	
K	0.023	0.026	0.027	0.029	0.031	0.025	0.010	0.007	
Total	5.005	5.006	5.006	5.013	5.006	5.004	4.980	4.999	
An%	52.6	48.6	55.9	52.2	50.6	57.1	41.9	58.7	

1. Units in parentheses represent one standard deviation of replicate analyses of last digit cited.

2. Unusually low An.

3. Unusually high An.

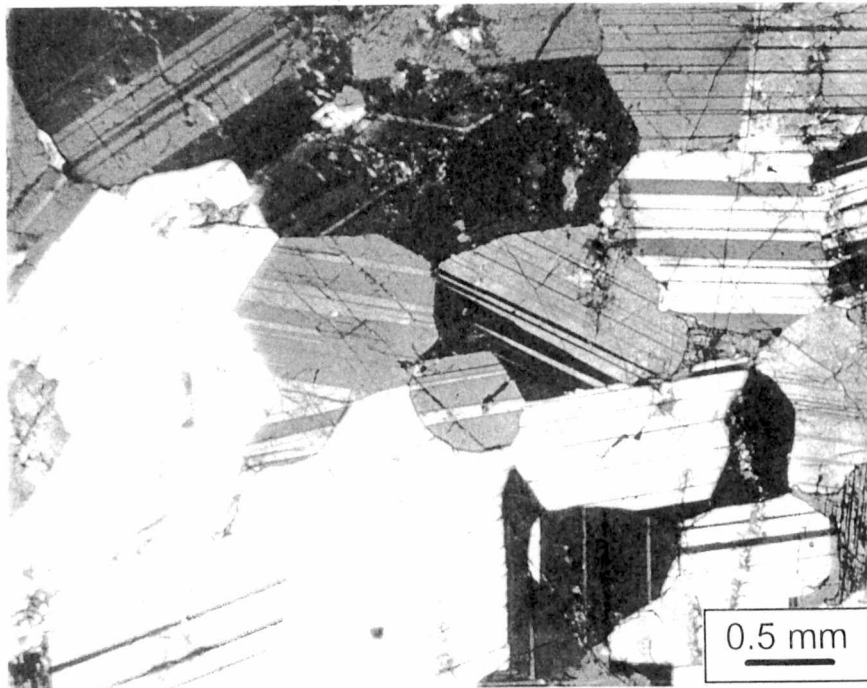


Figure 11. A transmitted light photomicrograph of typical plagioclase grains from the gabbronorite zone (sample 354/215.7). Note the distinctive euhedral nature of the plagioclase crystals. The texture is typical of cumulates.

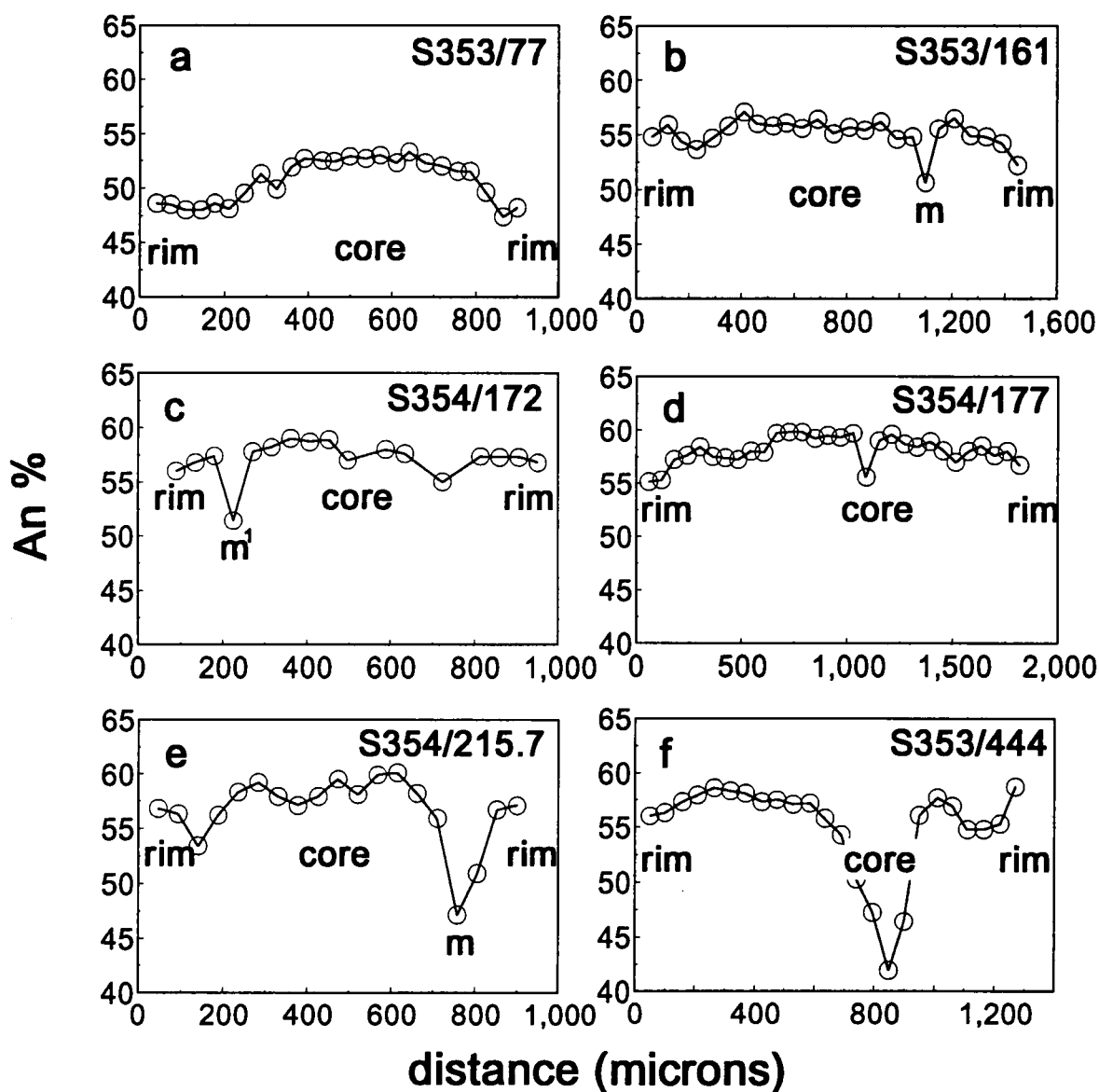


Figure 12. Zoning profiles of individual plagioclase crystals from selected samples of the gabbro-norite zone. Note the Na-rich cores in grains from the lower and middle portions of the intrusion (e.g., d and f) and the reversally zoned rims of the middle portion (e.g., c and e). Also note the variations in the sizes of the crystals as reflected by differences in the x-axis scales.

Pyroxenites

Pyroxenites (a total of 8 samples) from rock layers of differing stratigraphic height in the gabbro-norite zone were analyzed in this study. Six of these samples are websterites and two are clinopyroxenites. All of these samples are coarse grained mesocumulates. In the websterites, cumulus phases are clinopyroxene (0.6-3.0 mm in size) and orthopyroxene (1-3.0 mm in size) embedded in an intercumulus matrix, consisting primarily of plagioclase, that varies from 10-20 volume percentage of the samples. Clinopyroxene and orthopyroxene crystals in these samples are subhedral to euhedral with well developed parallel exsolution lamellae (10-20 μm in width). The X_{Mg} values of these minerals vary from 0.79 to 0.83 and 0.77 to 0.80, respectively. The narrow range in X_{Mg} values reflects the sampling interval, which is only about 100 meters for the pyroxenite layers. Representative major-element chemical compositions of individual pyroxenes are given in Tables 2 and 3. It should be emphasized that the reported compositions are reconstructed average chemistry for exsolved pyroxenes. End-member components of these pyroxenes are plotted in Figure 10. One of the websterites, sample 354/234, contains olivine as an additional cumulus phase. The two clinopyroxenites, 354/232 and 354/279, differ in one regard, in that sample 354/279 contains cumulus olivine. In samples 354/234 and 354/279 olivine occurs as subhedral to euhedral crystals that range in sizes of 7-12 mm and 1-1.5 mm, respectively. Moreover, many of the olivine grains are either completely altered, or exhibit only relict rims. Olivine compositions in these samples are Fo 77.2 and Fo 78.7, respectively.

Whole-rock chemical compositions

Table 6 gives whole-rock major-element compositions for samples from the gabbronorite zone. Figure 13 shows the variation in X_{Mg} with depth in the gabbronorite zone. In general, the X_{Mg} of these rocks varies from 0.72 to 0.42 towards the top of the gabbronorite zone. This trend towards more evolved compositions is consistent with the fractional crystallization of a mafic magma. In addition, Figure 13 indicates several reversals in the magma chemistry. These reversals likely record the replenishment of the residual magma with a more primitive one.

Whole-rock major-element compositions for samples from the ultramafic zone and the border group are given in Table 7. To account for low totals in the samples (probably due to alteration), LOI (loss on ignition) was determined. In general, the X_{Mg} of these samples varies from 0.91 to 0.80 towards the top of the ultramafic zone. The most primitive X_{Mg} values are from the lowest part of the ultramafic zone (i.e., the dunite subzone). The border group sample has a X_{Mg} value of 0.79.

8. DISCUSSION

The Burakovsky layered intrusion is a complexly differentiated series of rocks with many characteristics that are similar to other layered mafic intrusions occurring around the world. In particular, comparisons of the BLI and the Bushveld and Stillwater Intrusions have been made (Sharkov et al., 1995). While there are many similarities between the mafic sequence of the BLI (i.e., gabbronorite zone, pigeonite-gabbronorite zone, and the magnetite-gabbrodiorite zone) and those found in these two classic

Table 6. Whole-rock major-element (wt.%) concentrations for samples from the gabbronorite zone as determined by XRF analyses.

	353/60	353/94	353/141	353/156	353/161	354/172	354/177	354/182	354/207	354/214	354/215
SiO ₂	51.9	51.0	53.1	50.8	51.4	53.0	52.2	52.9	54.0	53.3	52.0
TiO ₂	0.39	0.38	0.15	0.24	0.17	0.16	0.16	0.14	0.32	0.25	0.07
Al ₂ O ₃	19.5	14.1	24.8	14.0	18.6	17.0	16.4	17.2	3.88	4.26	19.1
FeO	7.05	8.92	2.62	7.13	5.10	4.42	5.01	4.80	9.72	9.18	4.62
MnO	0.12	0.18	0.02	0.16	0.08	0.07	0.08	0.08	0.26	0.25	0.10
MgO	5.07	8.10	3.16	9.29	6.97	8.19	9.48	8.23	18.4	18.0	8.05
CaO	9.48	11.2	9.58	12.4	10.7	12.9	11.8	11.6	11.8	13.1	12.2
Na ₂ O	3.86	2.54	4.38	1.64	3.12	2.71	2.45	2.56	0.71	1.02	2.88
K ₂ O	0.44	0.27	0.47	0.11	0.33	0.12	0.12	0.11	0.06	0.06	0.08
P ₂ O ₅	0.10	0.11	0.06	0.12	0.06	<0.06	0.06	<0.06	0.15	0.12	<0.06
Total	98.0	97.1	98.4	96.0	96.7	98.6	97.8	97.7	99.3	99.5	99.1
X _{Mg}	0.56	0.62	0.68	0.70	0.71	0.77	0.77	0.75	0.77	0.78	0.76
	354/215.7	354/217.3	354/234	354/236	354/245.6	354/270.2	354/272	353/372	353/397	353/419.6	353/444
SiO ₂	52.0	54.9	54.1	53.9	53.2	54.4	53.5	52.1	51.0	50.0	51.9
TiO ₂	0.12	0.48	0.26	0.21	0.15	0.39	0.22	0.04	0.06	<0.03	<0.03
Al ₂ O ₃	17.3	3.88	4.14	16.1	5.51	3.72	2.57	21.7	21.4	22.8	21.3
FeO	4.95	8.65	7.88	4.83	10.7	9.27	7.90	3.03	3.29	2.97	2.71
MnO	0.10	0.25	0.27	0.10	0.27	0.27	0.25	0.05	0.06	0.03	0.04
MgO	9.02	15.5	16.3	7.48	21.4	17.5	18.8	6.16	6.06	5.82	6.63
CaO	12.0	15.7	15.7	10.8	4.99	11.8	15.0	11.4	12.1	12.0	11.1
Na ₂ O	2.63	0.94	0.86	1.90	1.00	0.91	0.59	3.37	3.24	3.37	3.28
K ₂ O	0.10	0.08	0.06	0.28	0.05	0.09	0.05	0.12	0.12	0.17	0.10
P ₂ O ₅	<0.06	0.12	0.07	0.06	0.21	0.16	0.08	<0.06	<0.06	<0.06	<0.06
Total	98.3	100.5	99.6	95.7	97.5	98.5	99.0	98.1	97.5	97.3	97.2
X _{Mg}	0.77	0.76	0.79	0.73	0.78	0.77	0.81	0.78	0.77	0.78	0.81

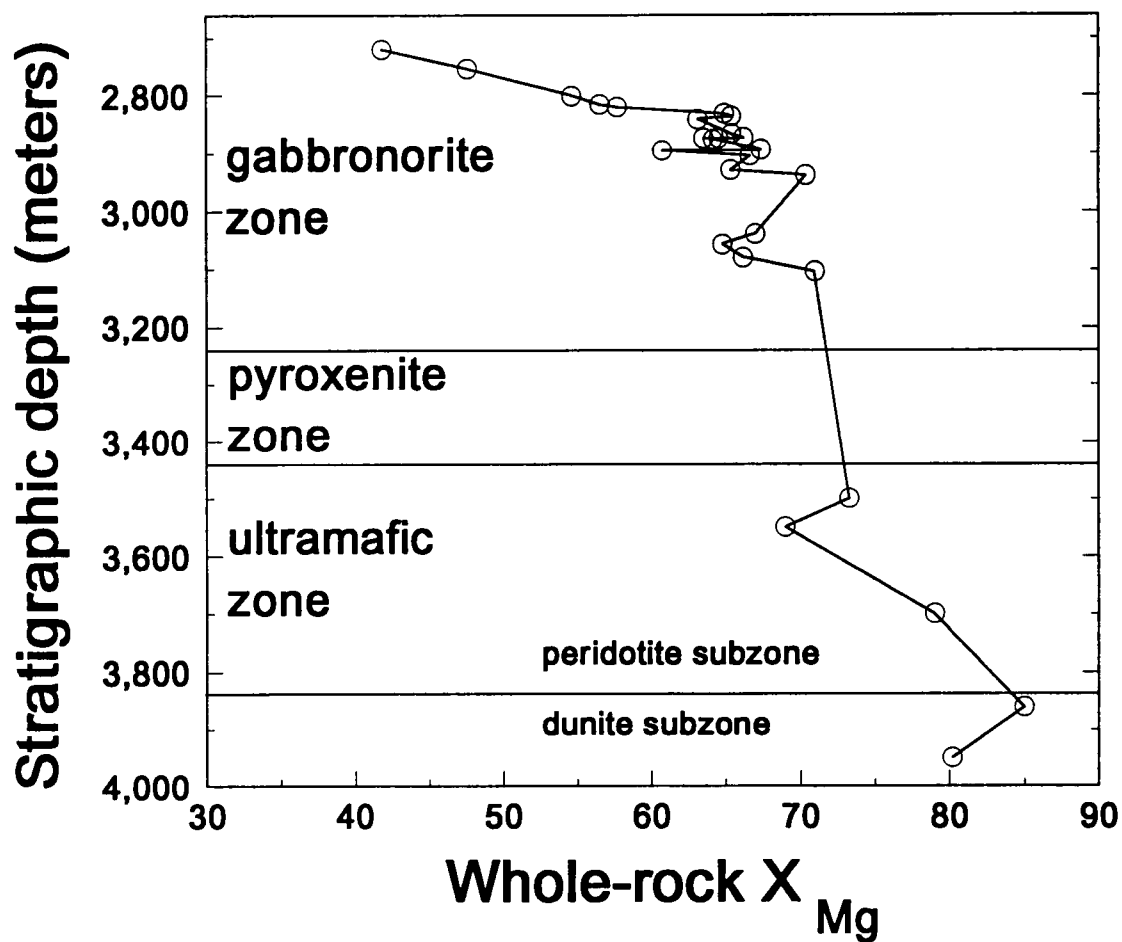


Figure 13. Variation in whole-rock major-element chemistry with stratigraphic depth in the BLI as reflected by X_{Mg} values. Only the top of the dunite subzone is shown due to a lack of samples from greater depths. Note the numerous reversals in magma composition recorded by samples from the gabbronorite zone.

Table 7. Whole-rock major-element (wt.%) concentrations for samples from the ultramafic zone and border group as determined by XRF analyses.

	248/190	333/496	200/444	15/152	28/223 ¹
SiO ₂	43.5	40.9	41.8	42.6	42.5
TiO ₂	0.15	0.22	0.11	0.03	0.36
Al ₂ O ₃	1.05	2.62	0.66	0.33	4.18
Cr ₂ O ₃	0.79	1.75	2.19	2.86	0.73
NiO	0.28	0.39	0.54	0.56	0.43
FeO	10.9	13.5	9.18	6.70	13.6
MnO	0.16	0.19	0.13	0.07	0.19
MgO	29.9	30.4	34.5	37.8	28.1
CaO	6.91	2.20	4.1	<0.15	3.49
Na ₂ O	<0.27	<0.27	<0.27	<0.27	0.49
K ₂ O	<0.07	<0.07	<0.07	<0.07	0.32
P ₂ O ₅	0.15	0.10	0.08	<0.06	0.14
LOI ²	6.25	8.70	9.42	7.77	4.82
Total	100.2	101.0	102.8	98.8	99.3
X _{Mg}	0.83	0.80	0.87	0.91	0.79

1. Border Group sample.

intrusions (i.e., relative cumulate stratigraphy, presence of PGE's, chromite-rich layers), the ultramafic sequences do not compare. First, the ultramafic zone of the BLI is composed of approximately 2600 meters of olivine and olivine + chromite cumulates overlain by about 400 meters of peridotite cumulates. These thick sequences of dunite are not found in either the Stillwater or the Bushveld complexes. In these intrusions, dunite primarily occurs as thin layers (i.e., on the scale of 50 meters) at the base of macro-rhythmic units comprised of dunite-harzburgite-orthopyroxenite (Hess, 1989; Raedeke and McCallum, 1984; Wager and Brown, 1967). In addition, both the Stillwater and the Bushveld complexes contain a thick sequence of almost monomineralic orthopyroxenite (~ 1000 meters and ~ 1800 meters, respectively), a feature not found in the BLI. Indeed, in the BLI, the only similar layer is the 200 meter thick pyroxenite zone (see Figure 4), but it is dominantly clinopyroxene not orthopyroxene. Therefore, the differences between the ultramafic zone of the BLI and the ultramafic sequences found in the Stillwater and Bushveld complexes indicate that another explanation is needed to describe its origin.

The Muskox intrusion in Canada provides a possible analog to the ultramafic zone of the BLI. This intrusion is approximately 2200 meters thick and can be divided into (1) a feeder dyke, (2) two marginal zones, and (3) a layered series (Irvine, 1970). The layered series is the major zone and consists of 2000 meters of rocks of ultramafic composition. There is no overlying mafic sequence, in contrast to the Stillwater and Bushveld complexes (Irvine, 1970). The rocks of the layered series are largely olivine-rich cumulates, ranging from dunites to olivine clinopyroxenites to olivine gabbros.

There are a few thin units composed entirely of pyroxene (forming websterites and orthopyroxenites). On the basis of these rock types, the layered series of the Muskox intrusion has been divided into 25 "cyclic" units (Irvine and Smith, 1967; Irvine, 1970). Although present data on the BLI are insufficient to delineate possible cyclic units in the UZ, the abundance of olivine in both intrusions is remarkably similar. Thus, the processes which formed the layered series of the Muskox may be similar to those that formed the ultramafic zone of the BLI. Consequently, it is useful to review the proposed origin for the layered series of the Muskox intrusion.

Irvine and Smith (1967) originally proposed, and Irvine (1970) elaborated upon, an origin for the thick sequence of ultramafic rocks that comprise the layered series of the Muskox intrusion. This involved the repeated addition of "new", undifferentiated, magma during the accumulation process. They suggested that after the formation of each cyclic unit, the remainder of the magma was flushed from the system by an influx of new magma, with a volume approximating the entire chamber. The magma flushed from the chamber then made its way to the surface further south as volcanic eruptions (Irvine, 1970). This process of flushing of the "old" magma by a large volume of "new" magma is a possible mechanism by which the ultramafic zone of the BLI could have formed. In the case of the BLI, the "old" magma may have been displaced from the chamber before clinopyroxene came on the solidus, causing only olivine-cumulates to form. Indeed, the variation in X_{Mg} over the upper 1600 meters of the ultramafic zone is only from 0.89 to 0.83 (see Figure 6), and, in fact, it is only in the top quarter (i.e., the upper peridotite subzone) that the Fo content begins changing demonstrably. This suggests that repeated

additions of a more primitive magma, frequent enough to prevent the crystallization of pyroxene until late in the sequence, could have occurred throughout the crystallization of the ultramafic zone. Another possibility is that the ultramafic zone crystallized from a volume of magma large enough that any variations in chemistry caused by fractionation of olivine would be relatively insignificant. The current data from the ultramafic zone indicate no definitive reversals in magma chemistry (as would be expected with repeated additions), and, therefore, the second possibility seems more likely at this time. As with the Muskox Intrusion, the “old” magma “flushed” from the magma chamber of the BLI likely migrated upward to produce extensive volcanic fields (i.e., the Karelian volcanic fields; 2.44 Ga old, Nykänen et al., 1994). These are intriguing possibilities that require further quantification through trace-element analyses of the current samples and detailed study of the Karelian volcanic fields.

The mafic rocks of the BLI provide strong evidence that fractional crystallization was the primary mechanism by which they formed. This is reflected in the well defined cryptic layering (e.g., X_{Mg} 0.81 to 0.56 towards the top) present in samples from the gabbro-norite zone. Moreover, these variations occur over an interval of only 500 meters, in contrast to the ultramafic zone (i.e., Fo 89 to Fo 83 over ~ 1600 m). The rapid change in magma chemistry over such a small interval suggests that the gabbro-norite zone formed from a relatively small volume of magma. The origin of this magma, however, is more problematic. One possibility is that the magma may have been the residue left after crystallization of the ultramafic zone and the subsequent “flushing” of the evolved material from the system. This more primitive residual magma (i.e., more

primitive than that which was flushed), now of a much smaller volume, would then crystallize to form the clinopyroxene-rich pyroxenite zone followed by formation of the gabbronorite zone. A second possibility is that the gabbronorite zone formed from an entirely different magma than the ultramafic zone. If this were the case, then the "old" magma left from crystallization of the ultramafic zone would have been completely "flushed" from the chamber, either before or after formation of the pyroxenite zone. To elaborate further upon the relationship of these two zones, trace-element data are needed, and that is beyond the scope of the current study.

Zoning in Plagioclase of the Gabbronorite zone

Plagioclase in the gabbronorite zone commonly have cores rich in Ab % compared to their rims. These cores are, in most cases, flanked by a broad, normally zoned, main portion (Figure 12f). The Na-rich cores are inconsistent with the expected products of crystallization from a primitive melt. Indeed, the Na-rich cores imply that some other process must have affected the magma composition prior to crystallization of plagioclase. One possibility is that the magma initially crystallized some other Ca-rich phase (e.g., clinopyroxene), depleting the liquid in Ca relative to Na. A possible product of this crystallization could be the 200 meter thick pyroxenite zone which underlies the gabbronorite zone, and this is consistent with the observation that the Na-rich cores occur primarily in plagioclases from the lower levels (i.e., close to the boundary between the two units). Another possibility, suggested by Morse and Nolan (1984) for similar zoning patterns observed in plagioclase crystals from the Kiglapait intrusion, is that the

Na-rich cores are a result of supersaturation at the onset of nucleation, and that the normally zoned main portion reflects steady state growth, consistent with the hypothesis of oscillatory nucleation.

Rim reversals in many of the plagioclase crystals from the BLI are observed, with differences in An of up to 12 mol.% (see Figure 12c and 12e). Morse and Nolan (1984) studied rim reversals, with differences in An of up to 32 mol.%, in cumulus plagioclase from the Kiglapait intrusion and determined that the origin must be from intercumulus liquid. They concluded that the reversals were due to an elevated partition coefficient (i.e., the CIPW-normative An partition $D^{pl/L}(X_{An})$) for An between crystals and liquid as a result of the enrichment of the intercumulus liquid in the augite component, *di*. Morse and Nolan (1984) further concluded that the local system may have been open to plagioclase exchange between pore liquid and main magma, but effectively closed to *di*. They showed, however, that this process will only contribute to small reversals, and that enrichment of the intercumulus liquid in *di* is needed in order to get the large reversals observed in plagioclase from the Kiglapait.

In the BLI, rim reversals in plagioclase are not nearly as pronounced as in the Kiglapait, at least not in the samples analyzed (differences in An of 12 mol.% versus 32 mol.%, respectively). Moreover, the rim reversals are in only a few samples (e.g., sample 354/215.7 - Figure 12e) from near the middle of the gabbro-norite zone. Furthermore, the fact that plagioclases from the upper portion of the gabbro-norite zone contain normally zoned rims (i.e., rather than reversely-zoned rims) is also noteworthy. This may indicate that communication between the pore liquid and main magma was

effectively closed off in the upper portion of the intrusion. One method by which this may be accomplished is through an increase in crystallization rate. If the rate of crystallization of plagioclase were increased, the rate of accumulation would also increase, and this would result in more liquid being trapped. Moreover, in a case like this, since there are numerous crystals forming and fractionating, it is more likely that the channels of communication between the pore liquid and the main magma will be cut off. This would result in closed-system crystallization of the pore liquid, which would form the normal zoning patterns found in samples from the upper portions of the gabbro-norite zone (e.g., 353/77, Figure 12a).

Pyroxenes

A major difference between the BLI and other layered complexes, besides the thick monomineralic UZ, involves the 200 meter thick Pyroxenite (transition) Zone. This zone, which is composed primarily of inverted-pigeonite with distinctive non-cumulus textures, is not known to occur in any other early Proterozoic layered mafic intrusion (Sharkov et al., 1995). These textures result from the replacement of earlier formed cumulus phases by non-cumulus clinopyroxene. Sharkov et al. (1995) suggested that this replacement process took place at high temperature, subsolidus conditions, within rocks recently solidified. In addition, the presence of oval-shaped quartz-carbonate inclusions, both within individual minerals and as veins in the rock, suggests that subsolidus metasomatism may have been an important process in the formation of this clinopyroxenite zone (Snyder et al., 1995; Sharkov et al., 1995).

There is a definitive chemical reversal to higher X_{Mg} in pyroxene and olivine at the top of the ultramafic Zone (i.e., at ~ 3000 meters) (see Figure 6). X_{Mg} values of orthopyroxene increase from 77 to 85 while the Fo content of olivine increases from 83 to 90. This major reversal provides evidence that the magma which formed the pyroxenite and gabbronorite zones, if the same as that which crystallized the ultramafic zone, must have undergone significant change, either through "flushing" of the evolved magma from the system, replenishment, or a combination of the two. Of course, as already mentioned, the change in chemistry at this level may simply be due to separate magmas forming the ultramafic and mafic sequences. To better constrain these ideas, it is helpful to consider the whole-rock chemical variations of samples from these two major zones.

Whole-rock chemical variations

Whole-rock chemical variations in rocks from the ultramafic and gabbronorite zones indicate that they formed from an evolving magma(s), as evidenced by distinctive cryptic layering (see Figure 13). In addition, Figure 13 provides further evidence (i.e., other than the mineral chemistry) that there were repeated reversals throughout the crystallization of the gabbronorite zone. It is likely that these reversals record times at which a new, more primitive magma was injected into the chamber. A similar trend is not apparent in whole-rock chemical variations found in the ultramafic zone (or in variations in olivine chemistry, see Figure 6), however, suggesting that there are differences between the two parent magmas or styles and/or rates of replenishment.

The whole-rock compositions of samples from the ultramafic are consistent with the previously expressed idea that the magma from which they formed was different from that which produced the gabbro-norite zone. This becomes even more apparent when attempting to reconstruct the origin of these two zones by modeling using the fractional crystallization program, MAGFOX (John Longhi, personal communication), with a parent composition represented by sample 28/223 from the border group and a pressure estimate of 2.9 kb (Koptev-Dvornikov, 1995). Sample 28/223, since it represents the chilled margin of the BLI, provides a possible parent liquid composition for the entire intrusion. When the composition is entered into the MAGFOX program, however, the results indicate that only the ultramafic zone rocks could have formed from a liquid of similar composition. Moreover, the liquid compositions at various solid percentages, when plotted on Harker diagrams of weight percentage oxide versus X_{Mg} (Figure 14), indicate that olivine was the primary crystallizing phase, consistent with the observed 2600 meter thick dunite-rich lower subzone. In addition, these results provide further evidence that the magma was either repeatedly replenished by more primitive material, or that the volume of magma was large enough, such that crystallization of a phase other than olivine (and minor chromite) did not occur until almost all of the sequence had formed. The current results (both whole-rock and mineral) indicate a consistent trend to more evolved compositions, however, rather than the periodic injection of new material (see Figure 6 and 13). Thus, as with the mineral chemical compositions, the whole-rock data are more consistent with the second possibility. In order to further clarify these

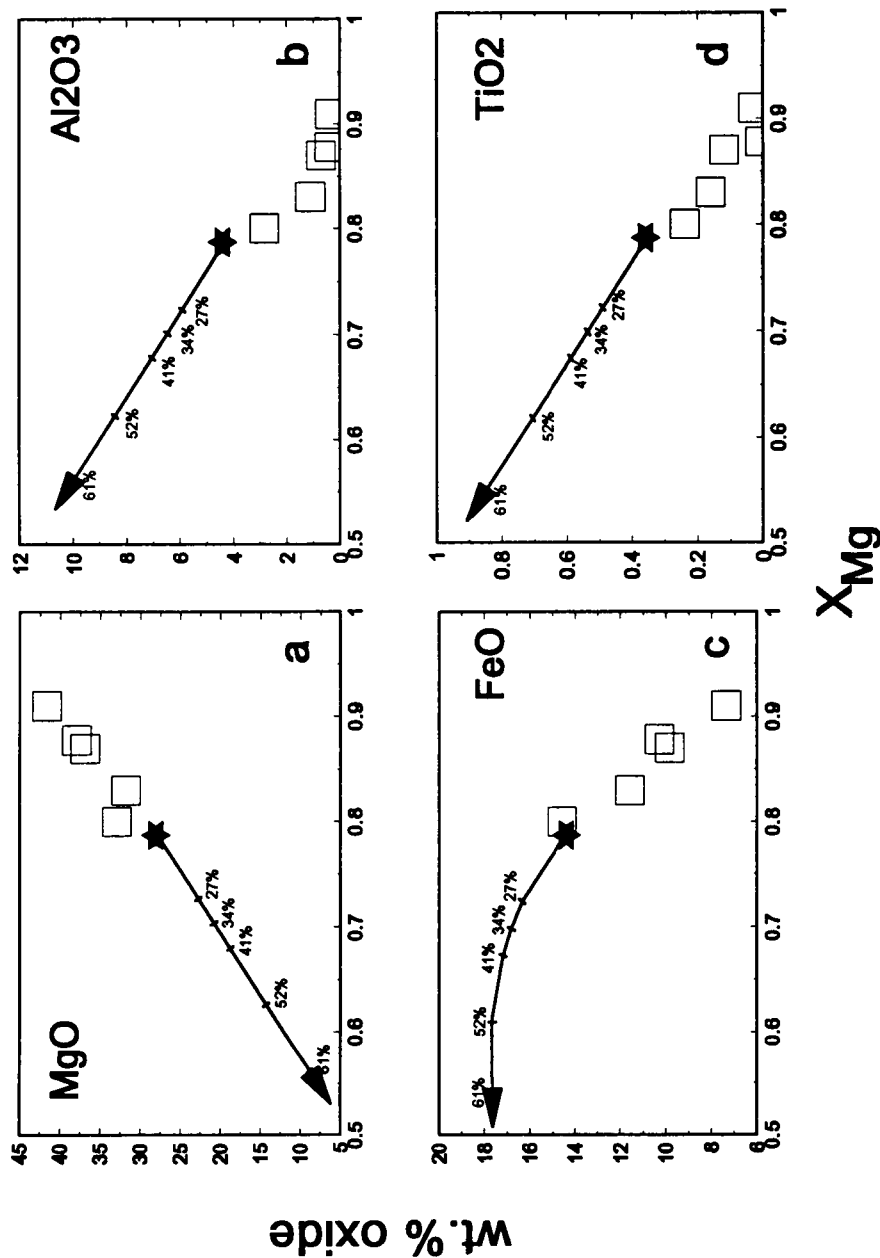


Figure 14. Harker diagrams plotting weight percentage oxide versus magnesium number for samples from the ultramafic zone. Model liquid compositions determined using the program MAGFOX (Longhi, personal communication) are also plotted. These diagrams show > 60% crystallization of olivine prior to the addition of another phase, consistent with the observed stratigraphy.

ideas, however, it is necessary to evaluate the trace-elements, for they may show reversals not seen in the major-element chemistry.

9. SUMMARY AND CONCLUSIONS

Major-element whole-rock and mineral compositions of rocks from the BLI indicate a progression towards more evolved compositions with increasing stratigraphic height. Olivines from the ultramafic zone indicate increasing Fe-enrichment upward in the intrusion (i.e., ranging from Fo 89 to Fo 83). Orthopyroxene cumulates from the gabbronorite zone reveal similar trends, with X_{Mg} decreasing from 0.84 to 0.58 upward. In the ultramafic zone, the variations occur over a much wider interval than in the gabbronorite zone (i.e., 1600 meters compared with 500 meters, respectively). This suggests that the rate of replenishment of the residual magma with new magma was much higher in the ultramafic zone than in the gabbronorite. Another possibility is that the volume of magma from which the ultramafic zone formed was large enough to prevent significant variation in the chemistry. These two ideas are also consistent with the stratigraphy, which shows that the ultramafic zone is composed almost entirely of monomineralic (dunites) rocks, while the gabbronorite zone contains several different lithologies (e.g., gabbro, pyroxenite).

Plagioclases from the gabbronorite zone are zoned. Crystals near the base contain Na-rich cores flanked by broad, more calcic rims. Towards the top of the gabbronorite zone, the rims become more sodium-rich, probably reflecting post-cumulus growth. The observed Na-rich cores are perhaps due to supersaturation at the onset of

nucleation, and the normal zoning of the main part due to steady state growth, consistent with the hypothesis of oscillatory nucleation. Another possibility is that these plagioclase crystals nucleated from a liquid which had been previously depleted in Ca by crystallization of some other Ca-rich phase (such as the clinopyroxene which forms the transition zone).

Plagioclases from near the middle of the gabbronorite zone have reversly-zoned rims. These rims may be the result of an elevated partition coefficient (i.e., the CIPW-normative partition $D^{pl/L}(X_{An})$) between crystals and liquid as a result of the enrichment of the liquid in augite component. Another possibility is that these rims resulted from the exchange of plagioclase components of the main magma with the intercumulus liquid surrounding the plagioclase during times of replenishment. This exchange would then cause the enrichment of the intercumulus liquid in An components (e.g., Ca), resulting in the formation of the observed An-rich rims. The lack of reverse zoning in plagioclase from near the top of the gabbronorite zone suggests that these crystals may have formed under different conditions. Indeed, it is likely that the conditions were such that the trapped liquid was essentially cut-off from communication with the main magma body. This allowed the pore liquid to crystallize in a closed system, resulting in normally zoned rims (as observed).

Fractional crystallization modeling using the MAGFOX program (Longhi, personal communication), with the whole-rock chemistry of the border group sample entered as the parent composition, will produce the entire sequence of ultramafic zone samples through crystallization of mostly olivine (as expected based upon the observed

stratigraphy). It is not possible, however, to produce the sequence of rocks observed in the gabbro-norite zone. This again suggests either that these zones formed from two different magmas, or in the least, that the border group does not represent a chilled margin for the mafic rocks. Moreover, if the ultramafic and gabbro-norite zones were formed from the same magma, some significant change in composition must have occurred following crystallization of the ultramafic rocks.

Future work will focus upon the trace-element chemistry of the BLI. In particular, it is hoped that the trace-element data will enable us to better clarify what processes acted to form the ultramafic and gabbro-norite zones, and, in addition, they should aid us in determining the parent magma for the mafic sequence of the BLI.

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PART 2

**PETROLOGY OF THE EARLY PROTEROZOIC BURAKOVSKY
LAYERED INTRUSION, SOUTHERN KARELIA, RUSSIA:
TRACE-ELEMENT GEOCHEMISTRY**

1. ABSTRACT

The early Proterozoic Burakovsky layered intrusion (2449 ± 1.1 Ma) is the largest mafic pluton in the Baltic Shield with a surface area of over 700 km². The first part of this study (Higgins et al., 1996) presented major-element evidence for the formation of the ultramafic and gabbro-norite zones. In this study, trace-element data for samples from these two zones are presented. Samples from the ultramafic zone are LREE-enriched indicating that they formed from a LREE-enriched parental magma. Indeed, a calculated parental magma for the ultramafic zone has a (Ce/Yb)_n ratio of 2.6, a (Nd/Sm)_n ratio of 1.1, and a (Dy/Yb)_n ratio of 1.6. The LREE-enrichment in the parental magma suggests either that the source region was LREE-enriched or that the melt was contaminated by crust in route to the BLI magma chamber. Samples from the gabbro-norite zone are also LREE-enriched. Moreover, there are at least two parental magmas for the gabbro-norite zone delineated by these REE characteristics. Thus, it is proposed here that the lower subzone of the gabbro-norite zone be broken into two additional zones, denoted by Group I and Group II magmas. Group I magmas, from the lower part of the gabbro-norite zone, have (Ce/Yb)_n ratios of 6.9 to 13.9 while Group II magmas, from the upper portion, have (Ce/Yb)_n ratios of 15.8 to 27.3. Volcanic rocks in Karelia that are coeval to the BLI and volcanic rocks of a similar age in other parts of the Baltic Shield are also LREE-enriched. Furthermore, the BLI has an ϵ_{Nd} value of -2.0, and other layered intrusions in the Baltic Shield of similar age also have negative ϵ_{Nd} values (e.g., -2.3 to -2.0). All of these data are most consistent with a LREE-

enriched mantle, probably as a result of metasomatism, during the Archean-Proterozoic transition.

2. INTRODUCTION

The 700 km² Burakovsky layered intrusion (BLI) of southern Karelia, Russia is the largest mafic pluton in the Baltic Shield. It lies within the east-west trending Finno-Karelian supracrustal belt and is associated with the most prominent suture zone on the Baltic Shield, the East Karelia (Figure 1). U-Pb dating of zircons indicate an age of 2449 ± 1.1 Ma for the BLI (Amelin et al., 1995), consistent with the idea that the BLI and other layered intrusions of the Baltic Shield (Figure 2) formed as a result of intrusive igneous activity that occurred between 2.4 and 2.5 billion years ago. Emplacement of these layered intrusions occurred as two distinct magmatic episodes at 2505-2501 Ma in the Kola Peninsula and 2449-2441 Ma in Karelia and the Kola Peninsula (Amelin et al., 1995). Intrusions which mark the first event include the Monchegorsk and Pana tundra plutons (2504.4 ± 1.5 Ma and 2501.5 ± 1.7 Ma, respectively). The second magmatic event is evidenced by such plutons as the Burakovsky of southern Karelia and the Lukkulaivaara (2442.1 ± 1.4 Ma) and Tsipringa (2441.3 ± 1.2 Ma) intrusions of the Olanga complex in northern Karelia (Figure 2). Both of these events are related to continental rifting in zones of pre-existing lithospheric weaknesses (Amelin et al., 1995). Finally, the emplacement of these intrusions during the early Proterozoic coincides with world-wide intrusive igneous magmatism as evidenced by the widespread occurrence of layered plutons and

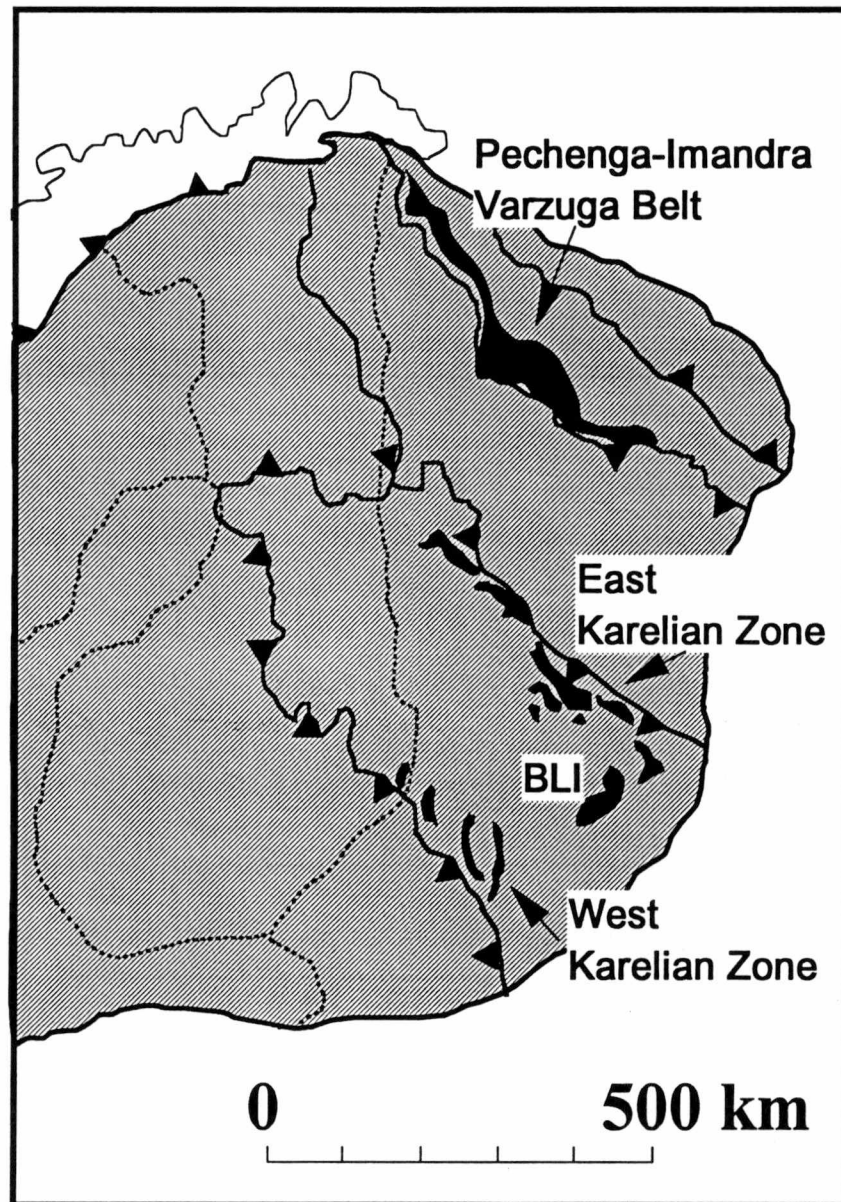
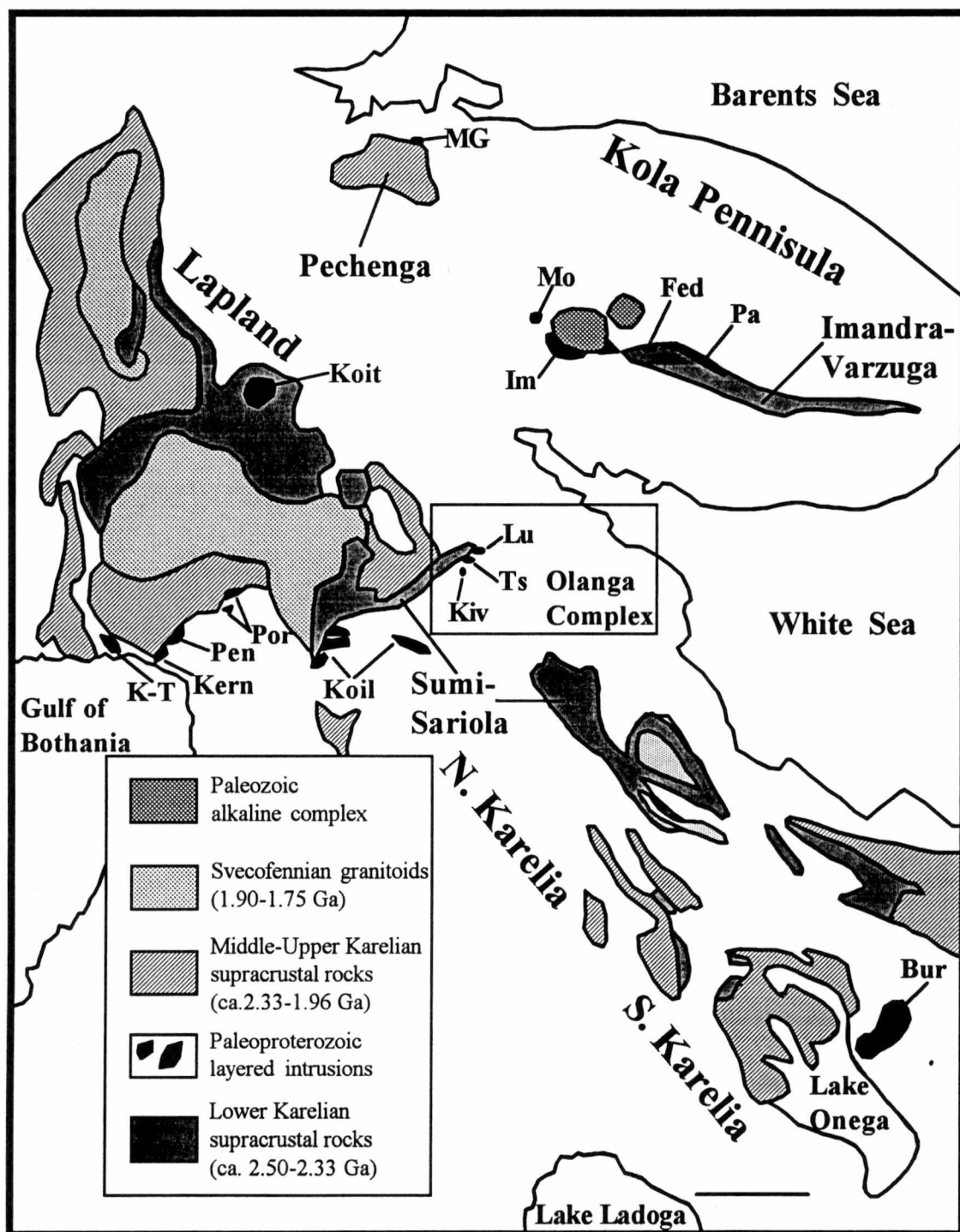


Figure 1. A tectonic map of the Baltic Shield showing suture zones, the two main supracrustal belts (Imandra-Varzuga and Finno-Karelian), and the location of the BLI (modified after Higgins et al., 1996). The Finno-Karelian belt is broken into the East Karelian and West Karelian Zones.

Figure 2. A schematic geological map of the eastern Baltic Shield showing the location of Proterozoic layered intrusions and volcanic rocks, modified after Amelin et al. (1995). The location of the Burakovsky is also given. Layered intrusions shown, in alphabetical order, include: Bur = Burakovsky; Fed = Fedorova tundra; Kem = Kemi; Kiv = Kivakka; Koil = Koillismaa complex; Koit = Koitelaninen; K-T = Kukkola-Tornio; Lu = Lukkulaivaara; MG = Mt. General'skaya; Mo = Monchegorsk; Pa = Pana tundra; Pen = Penikat; Por = Portimo complex; and Ts = Tsipringa.



mafic dyke swarms of similar ages on other cratons. Noted among these is the Great Dyke in the Zimbabwe Craton (2461 ± 16 Ma; Wilson and Prendercast, 1989) and the Kimberlana intrusion in the Yilgarn Block in Australia (2420 ± 30 Ma; McClay and Campbell, 1976).

The lithology of the BLI is dominated by layered ultramafic and mafic rocks, and it has been divided into three major blocks by faults: the Aganozersky, the Shalozersky, and the Burakovsky (Figure 3). An extensive description of the mineralogy, petrology, and structure of this intrusion is given in Sharkov et al. (1995) and Higgins et al. (1996). These authors recognized five major stratigraphic subdivisions in the BLI. From the bottom upwards, these are: (1) an ultramafic zone, consisting of 2.6 km of dunites overlain by 0.4 km of peridotites; (2) a pyroxenite zone, consisting of 0.2 km of clinopyroxenites; (3) a gabbro-norite zone, consisting of a 0.7 km thick lower banded subzone of alternating plagioclase-rich and olivine/pyroxene-rich cumulates and a 0.32 km thick upper anorthosite subzone; (4) a 1.2 km pigeonite-gabbro-norite zone; and (5) a 0.8 km magnetite-gabbro-diorite zone (Figure 4). The top of the ultramafic zone is marked by the disappearance of olivine and the appearance of orthopyroxene and clinopyroxene as cumulus phases. The boundary between the pyroxenite and gabbro-norite zones is evidenced by the stable appearance of plagioclase as a cumulus phase.

The first part of this study discussed the major-element variations found in the BLI (Higgins et al., 1996). Higgins et al. (1996) used these data to propose two different origins for the ultramafic zone and gabbro-norite zone. In the first hypothesis, they suggested that the gabbro-norite zone formed from the residue left from formation

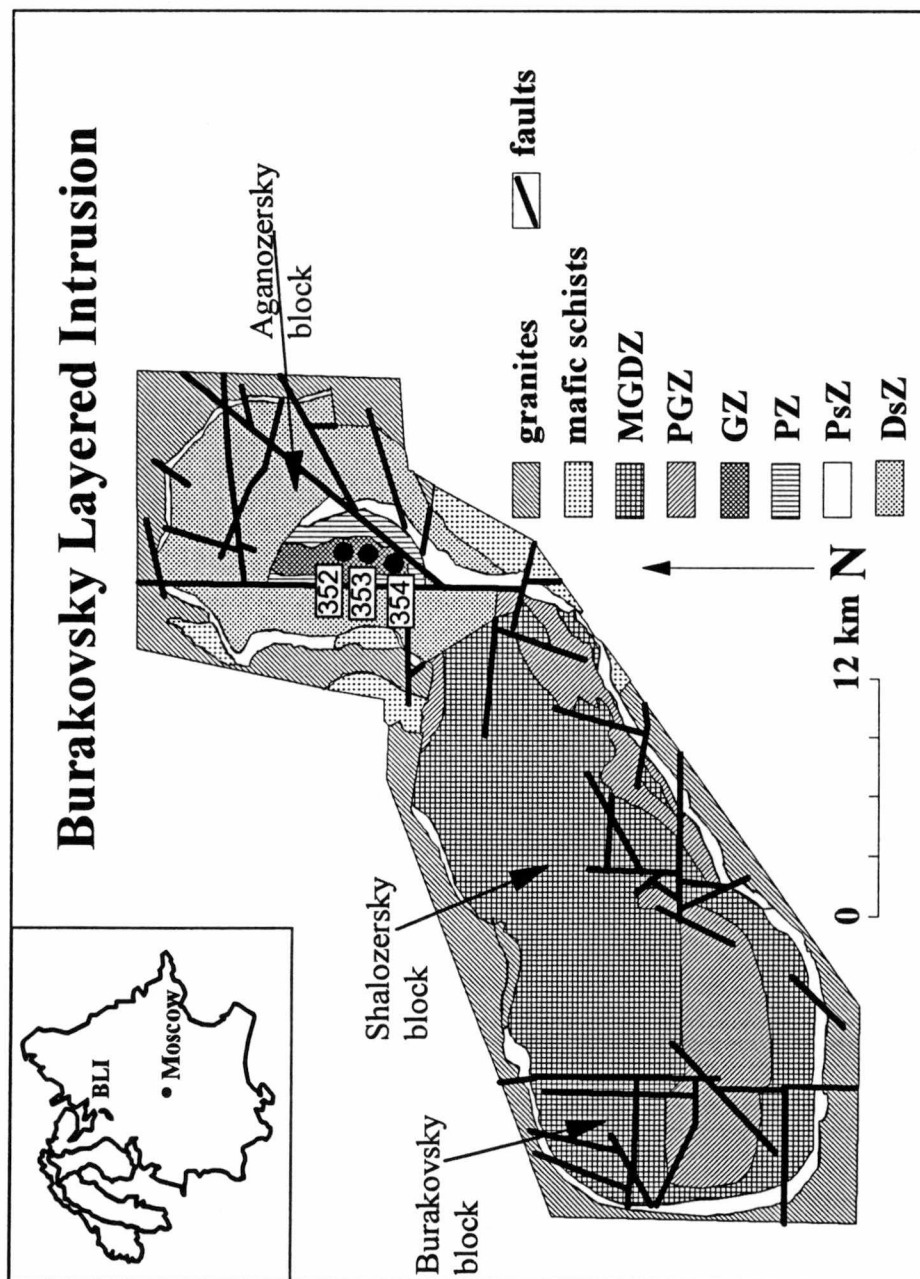


Figure 3. A geologic map of the Burakovsky layered intrusion with major faults shown by the solid lines (modified after Higgins et al., 1996). MGDZ: magnetite gabbrodiorite zone; PGZ: pyroxenite gabbro zone; GZ: gabbro zone; PZ: pyroxenite zone; PsZ: peridotite subzone; DsZ: dunitite subzone.

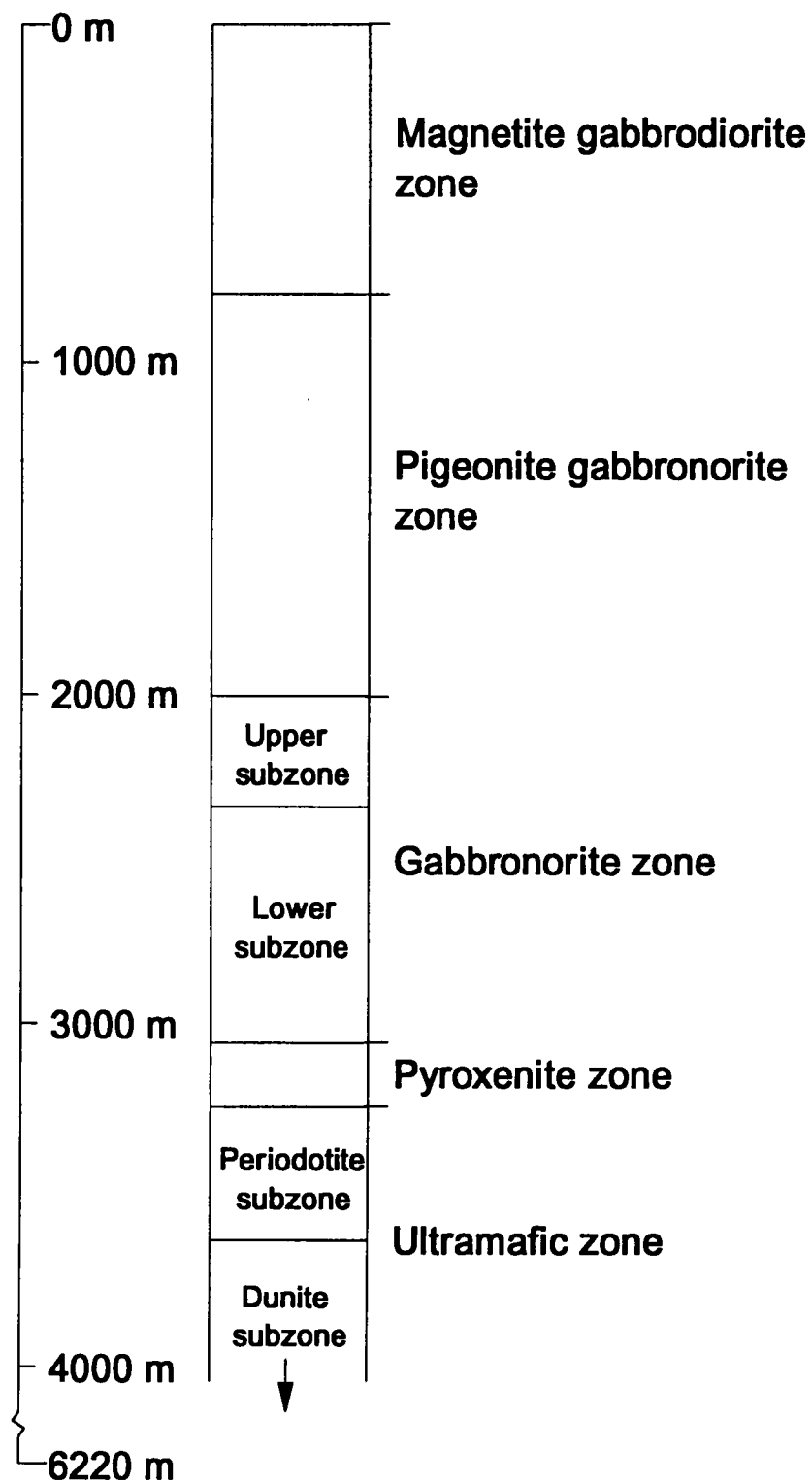


Figure 4. A schematic stratigraphic profile of the layered series of the BLI, modified after Higgins et al., 1996. Note that the ultramafic zone makes up almost half of the overall thickness of the layered series. MCH = main chromite horizon.

of the ultramafic zone by repeated injections of new material and the subsequent flushing of old material. The second proposed origin involved formation of these two zones from different magmas. In this paper, we will address these ideas by considering the variations in trace-elements within the two zones. Higgins et al. (1996) also determined that major-element variations in the gabbro-norite zone were consistent with the periodic injection of new magma. The trace-element data permit us to further elaborate upon the nature and style of these magma injections. In addition, these data will aid us in calculating a parent magma for the intrusion and in determining its relationship to the surrounding, coeval, Karelian volcanic fields. Finally, these data will help us to decide if these volcanic rocks are products of the "flushing" of old material from either the ultramafic or gabbro-norite zones. Subsequently, this may enable us to learn more about the nature of the mantle during this important time in Earth history.

3. SAMPLING AND ANALYTICAL METHODS

Whole-rock trace-element analyses were performed on fifteen samples from several drill cores which penetrate the gabbro-norite and ultramafic zones of the BLI (Figure 5). Of the samples analyzed, two were from the ultramafic zone and one from the border group. Samples are labeled using core number and depth in meters (e.g., core #/depth). An extensive major-element study was conducted on samples from drill cores 353 and 354 and these results are presented in Higgins et al. (1996). In this paper, we have also included trace-element analyses for samples from one other core,

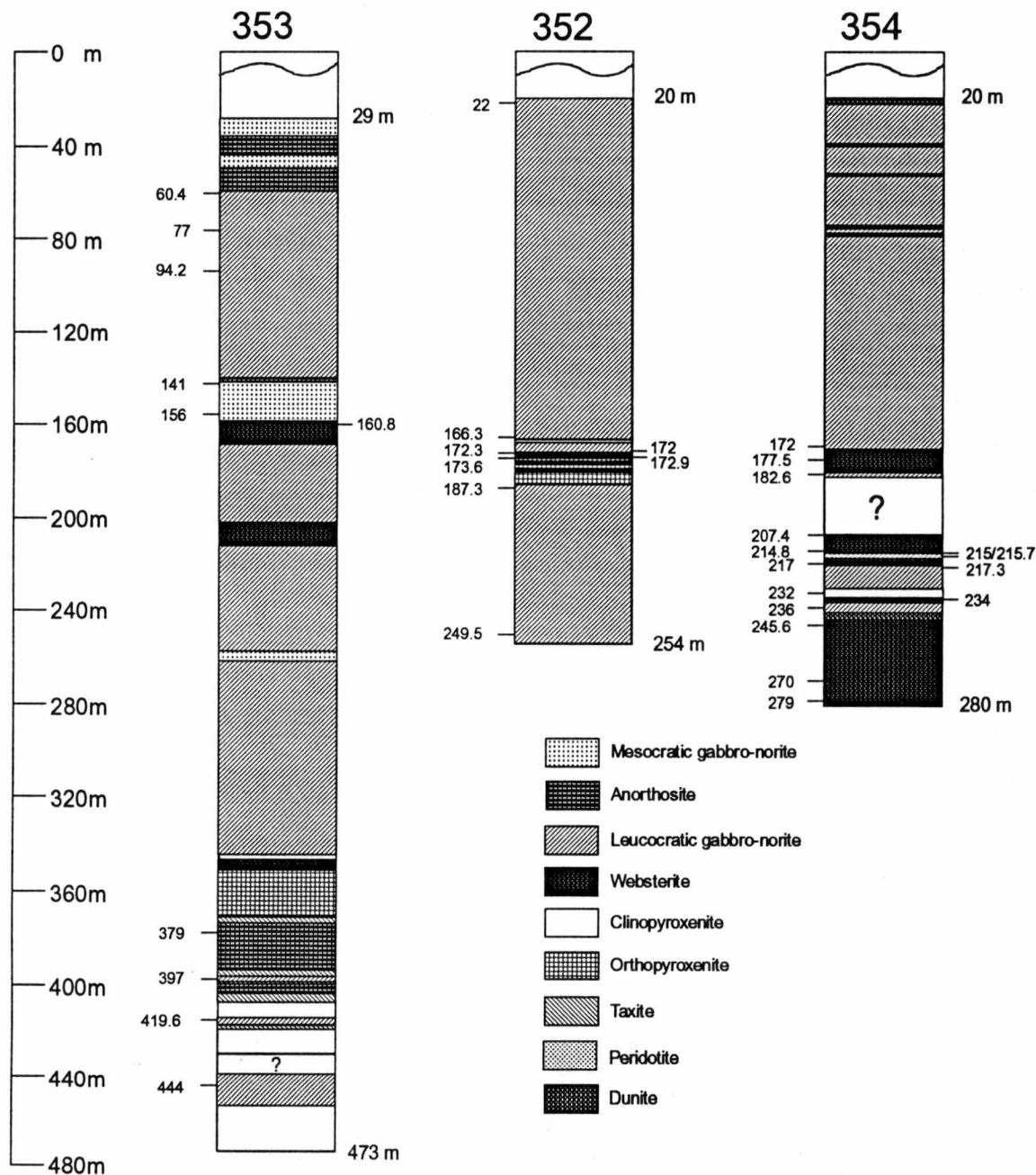


Figure 5. The stratigraphy of three of the drill cores which penetrate the gabbro-norite zone of the Burakovsky. The location of all of the samples that we have, including those analyzed in this study, are noted beside each column.

352, but the bulk of the data are for cores 353 and 354. The major-element data for these additional samples from core 352 will be presented in a future paper.

Trace-element concentrations for samples from the ultramafic zone and the gabbro-norite zone were determined with a FISIONS-VG PlasmaQuad II STE Inductively Coupled Plasma Mass Spectrometer at the University of Notre Dame. Detection limits for most elements that we are interested in (e.g., the REE's, Co, Ni, Cr, Hf, Li, Ba) are in the parts per billion (ppb) range. These are, in ppb, as follows: (1) Sc = 0.054; (2) V = 0.113; (3) Cr = 0.105; (4) Co = 0.005; (5) Ni = 0.008; (6) Cu = 0.206; (7) Rb = 0.150; (8) Sr = 0.207; (9) Y = 0.003; (10) Zr = 0.018; (11) Nb = 0.003; (12) Ba = 0.041; (13) REE = 0.03-0.024; (14) Hf = 0.011.

4. RESULTS

The analytical results for samples from the ultramafic zone are presented in Table 1. All of the trace element data are reported in parts per million. Major-element data for these samples can be found in Higgins et al. (1996). Chondrite-normalized REE plots of these data appear in Figure 6. In general, all of the samples have negative slopes and are LREE-enriched. This suggests that the parental magma for these samples was also LREE-enriched.

Whole-rock trace-element abundances, in parts per million, for samples from the gabbro-norite zone are given in Table 2. The major-element data for these same samples are given in Part I of this two part series (i.e., Higgins et al., 1996). The REE abundances, normalized to chondrite, are shown in Figure 7. As with the ultramafic zone, all of the samples are LREE-enriched and depleted in the HREE. Moreover, most

Table 1. Whole-rock trace-element abundances (ppm) for selected samples from the ultramafic zone. Included are the trace-element concentrations in a calculated parent for this zone.

Element	248/190	200/444	28/223 ¹	cal. parent
Sc	22.2	12.7	11.2	13.9
V	72.7	49.9	76.8	103.7
Cr	3379.1	6915.5	3744.4	2431.4
Co	104.8	102.1	124.9	60.0
Ni	1461.7	2359.4	1947.7	220.6
Cu	23.1	221.7	84.3	107
Rb	0.34	0.07	8.3	11.4
Sr	26.8	13.7	81.2	7.63
Y	3.04	1.79	5.61	7.63
Zr	6.96	2.8	40.4	55
Nb	0.27	0.1	1.88	2.58
Ba	3.03	8.12	115.7	36.8
La	0.92	0.49	5.8	2.22
Ce	2.4	1.39	13.2	6.31
Nd	1.76	1.02	6.64	4.63
Sm	0.54	0.3	1.47	1.36
Eu	0.17	0.1	0.39	0.45
Gd	0.53	0.31	1.27	1.74
Tb	0.09	0.05	0.2	0.27
Dy	0.6	0.33	1.13	1.55
Ho	0.11	0.07	0.2	0.27
Er	0.31	0.17	0.54	0.74
Tm	0.04	0.02	0.06	0.08
Yb	0.25	0.16	0.47	0.64
Lu	0.03	0.02	0.06	0.08
Hf	0.19	0.11	1.05	1.43

¹Sample from the border group.

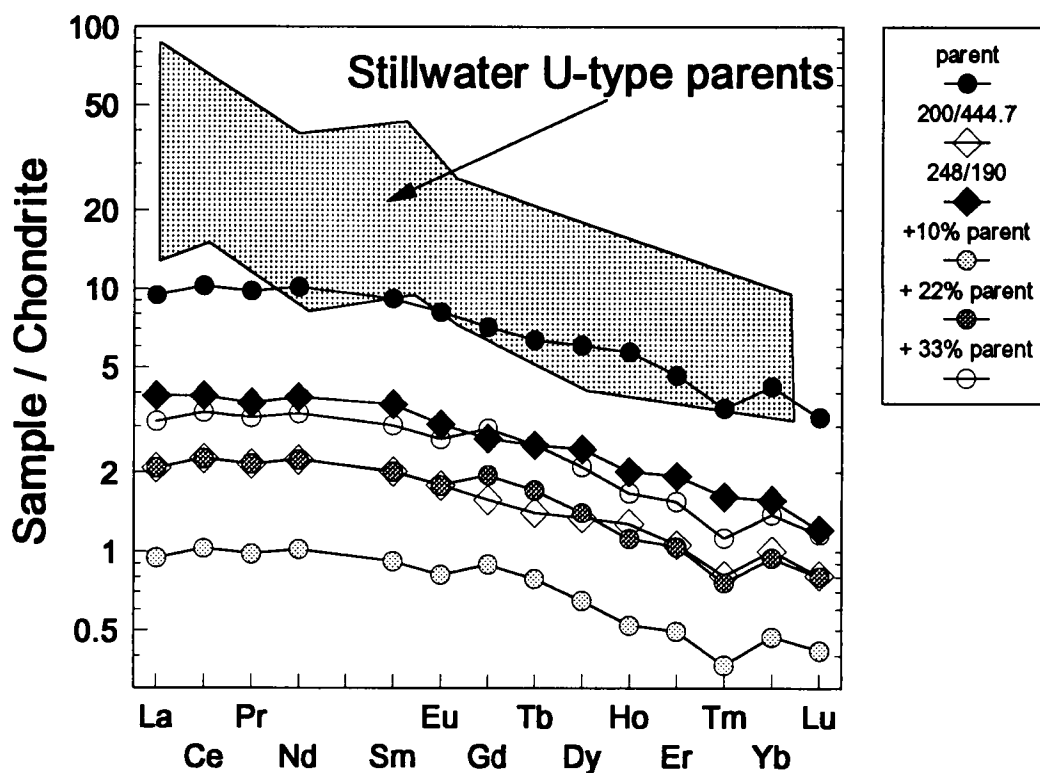


Figure 6. A chondrite-normalized REE plot of samples from the ultramafic zone of the Burakovsky. A calculated parental liquid (the method is outlined in the text) for the ultramafic zone is also shown. Note that it is possible to produce the REE characteristics of the samples with various combinations of the parent and the dunite. A field representing U-type parent magmas for the ultramafic series of the Stillwater Complex is shown for comparison (after Papike et al., 1995). Data from this study are presented in Table 1.

Table 2. Whole-rock trace-element abundances (ppm) for selected samples from the gabbro-norite zone.

Element	353/60	353/94	353/140	353/160	354/172	354/207	354/215	354/245	354/279	353/397	352/172	352/177
Sc	17.9	34.1	3.03	17.7	22.7	35.0	16.6	23.3	39.2	11.4	35.8	37.5
V	118.7	190.0	18.4	84.5	105.2	162.0	73.7	107.3	154.1	41.1	170.7	170.1
Cr	120.0	1510.0	27.6	103.0	317.8	924.7	541.7	3093.6	2974.7	466.5	1554.0	2943.6
Co	42.5	57.3	16.3	37.5	160.0	187.2	201.7	210.1	121.9	27.8	59.3	74.6
Ni	204.6	226.3	90.0	230.3	239.3	574.8	222.7	626.8	735.3	173.5	460.6	575.0
Cu	572.9	245.3	137.2	117.7	89.7	188.1	64.8	152.8	57.8	61.5	36.2	44.2
Rb	10.9	5.32	15.1	6.84	1.55	1.30	0.88	1.81	0.32	2.02	26.2	1.47
Sr	451.1	287.4	558.3	419.8	501.4	63.2	613.7	116.7	43.7	720.4	123.0	56.1
Y	6.25	8.15	2.03	3.41	3.77	5.98	2.49	3.58	4.62	2.04	9.68	5.57
Zr	31.0	24.8	127.2	15.4	8.06	18.7	4.46	10.6	4.69	4.17	70.2	10.4
Nb	1.68	0.81	2.08	0.80	2.63	3.68	6.55	3.84	1.78	0.16	5.04	0.17
Ba	180.6	109.9	207.2	116.4	52.5	18.0	45.3	55.2	6.00	51.5	83.3	15.6
La	5.59	4.08	6.09	3.40	1.43	1.76	0.92	1.51	0.48	1.20	40.5	2.0
Ce	12.1	9.58	11.6	6.77	3.46	4.58	2.05	3.57	1.77	2.74	23.1	5.59
Nd	5.31	5.49	3.82	2.97	2.30	3.29	1.43	2.31	2.00	1.60	10.0	3.11
Sm	1.19	1.44	0.59	0.65	0.67	0.97	0.42	0.60	0.76	0.41	2.17	0.87
Eu	0.57	0.52	0.50	0.38	0.36	0.26	0.27	0.24	0.25	0.27	0.58	0.26
Gd	1.22	1.48	0.57	0.65	0.71	1.03	0.47	0.59	0.81	0.40	1.99	0.86
Tb	0.20	0.25	0.06	0.10	0.12	0.19	0.07	0.10	0.16	0.06	0.34	0.16
Dy	1.22	1.61	0.38	0.63	0.76	1.18	0.47	0.68	1.00	0.38	2.03	1.06
Ho	0.22	0.30	0.07	0.12	0.14	0.21	0.09	0.13	0.18	0.07	0.37	0.19
Er	0.62	0.86	0.19	0.33	0.37	0.63	0.26	0.36	0.48	0.20	0.97	0.53
Tm	0.08	0.12	0.02	0.04	0.05	0.09	0.05	0.05	0.06	0.03	0.14	0.08
Yb	0.56	0.76	0.15	0.32	0.35	0.56	0.23	0.38	0.39	0.16	0.87	0.51
Lu	0.08	0.10	0.02	0.04	0.04	0.08	0.02	0.05	0.05	0.01	0.11	0.07
Hf	0.86	0.74	2.65	0.41	0.26	0.54	0.15	0.30	0.23	0.12	1.98	0.32

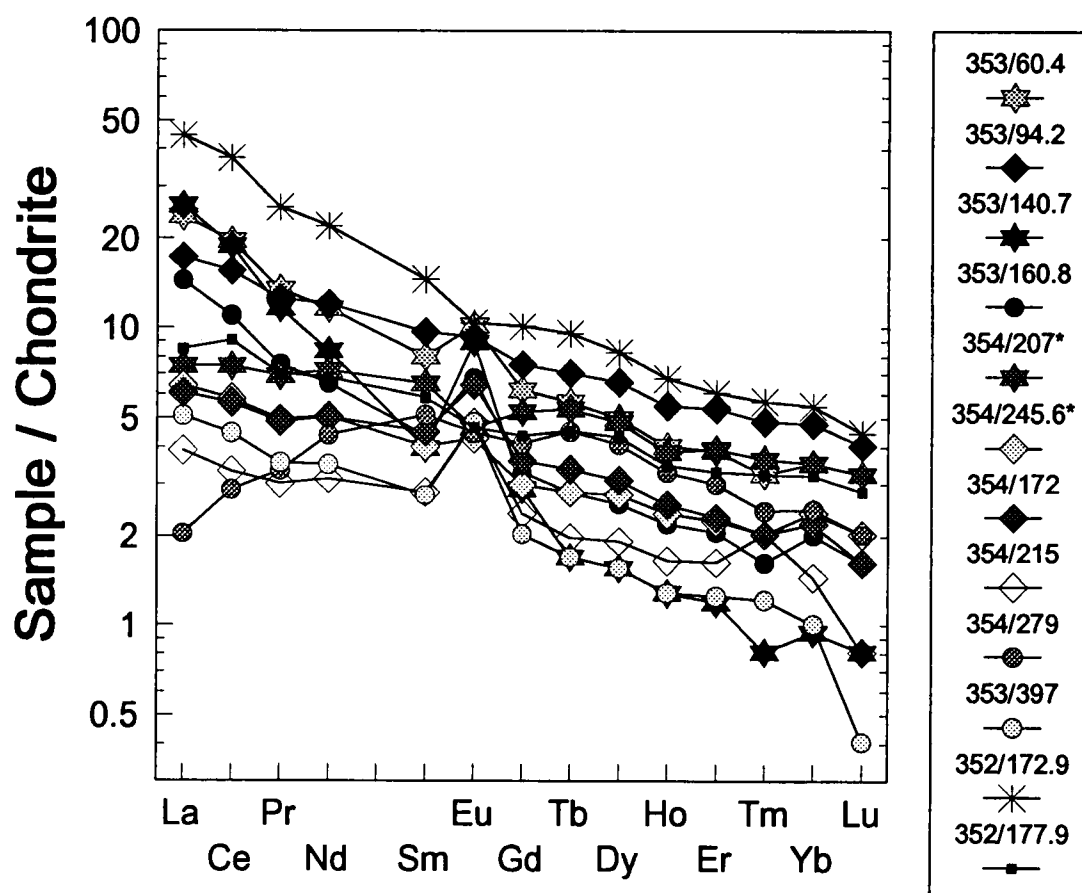


Figure 7. A chondrite-normalized REE plot of selected samples from the gabbro-norite zone. Note that, with the exception of 354/279 which is depleted in the LREE, the samples are all LREE-enriched. Note also that sample 353/140 has a much steeper slope than the other samples. Data are presented in Table 2.

of the samples have positive Eu anomalies, which reflects the presence of plagioclase. Note that sample 353/140 has anomalously low HREE values and that sample 354/279 is strongly depleted in the LREE relative to the other samples. This may suggest contamination, or, in the case of sample 354/279, formation from an ultramafic magma. These ideas will be considered in detail in the ensuing discussion.

5. TRACE-ELEMENT MODELING EQUATIONS

A trace element is said to be incompatible when its mineral-melt partition or distribution coefficient value is less than one. The distribution coefficient is defined as

$$K_d = C_i^{\text{Mineral}} / C_i^{\text{Melt}}$$

where C denotes concentration and i the element. This relationship is used in a variety of modeling equations in order to determine the concentration of a particular trace-element in the liquid or solid fraction.

Of particular importance is the Rayleigh equation, which governs trace-element partitioning during fractionation. This equation is defined as

$$C_L / C_o = F^{(D-1)}$$

where C_L is the concentration of the element in the liquid, C_o the original concentration, F is the fraction of melt remaining, and D is the bulk distribution coefficient for the particular element in the rock. Liquids in equilibrium with associated cumulates can be calculated using the simple, equilibrium partition-coefficient relation,

$$C_s = D * C_L$$

(where, C_s = concentration of the trace-element in the cumulate).

While the abundances of trace-elements can change appreciably during fractional crystallization, the relative proportions or ratios of various highly incompatible elements should not noticeably change until the latter stages of crystallization (Allègre and Minster, 1978). This idea provides a means of testing whether associated igneous rocks are fractionally crystallized derivatives of the same or different magmas. Moreover, cumulus minerals may be used as 'liquid analogs' since incompatible trace-element ratios in minerals are theoretically proportional to their ratios in the melts from which they fractionated (Lambert and Simmons, 1987, 1988). In this study, we have used these relationships to describe the petrogenesis of the ultramafic and gabbro-norite zones.

6. PARENTAL MAGMA CHARACTERISTICS

A major problem in modeling REE abundances in cumulus rocks and minerals concerns the calculation of the parent magma composition (e.g., Wager and Deer, 1939). In particular, this is difficult when working with adcumulates in which no intercumulus phases representative of the parental magma remain. Indeed, even in orthocumulates, the interstitial minerals probably do not truly reflect the closed-system crystallization of trapped intercumulus liquid (Irvine, 1980a). One way around this problem is to use a basal chill margin to estimate the parental magma characteristics (Wager and Deer, 1939). Theoretical modeling of REE partitioning in orthopyroxene and plagioclase (Simmons and Lambert, 1982) has demonstrated that cores of cumulus minerals may be used to estimate parental magma characteristics. Of course, this requires knowledge of the trace-element abundances in individual minerals, and these

data are not always available. In fact, as with the current study, whole-rock trace-element abundances are often the only available means for estimating the parent compositions. Regardless of the method, these data, when combined with mineral-melt partition coefficients (K_d values), provide important clues concerning the chemistry of the liquid from which the various cumulates crystallized.

An added complication that arises when attempting to model the characteristics of liquids in equilibrium with rocks and minerals concerns the K_d values that are used. Indeed, in practice, there is a wide range of values to integrate into the modeling particularly when dealing with phases such as olivine for which all of the incompatible elements have K_d values $\ll 0.01$ (e.g., REE's such as La and Nd have K_d values of 0.0001; McKay, 1986). Also, because K_d values may vary as a function of bulk chemistry and/or fO_2 , it is important to carefully peruse the available literature in order to pick the values most appropriate for the particular study in which they are to be used. Table 3 summarizes the partition coefficients used in this study.

Calculated parental liquid for the Ultramafic Zone

Major-element modeling with the program MAGFOX (John Longhi, personal communication) indicates that the border group of the BLI represents a chilled margin of the ultramafic zone (Higgins et al., 1996). On the basis of phase relations, however, this sample is too enriched in Mg relative to Fe to qualify as a parental magma (i.e., $X_{Mg} = 0.78$ where X_{Mg} is defined as $Mg/(Mg+Fe)$). Therefore, it was necessary to recalculate the parental liquid for the ultramafic zone using a mass balance approach. In this calculation, we assumed that the liquid composition represented by the border

Table 3. Mineral/Melt Partition Coefficients

Element	oliv ¹	augite ²	opx ³	plag ⁴	spinel ⁵	ilmenite ⁶
Sc	0.27	0.808	0.83 [*]	0.0071	1.5	0.55
V	0.039	1.81	1	2.24 [@]	38 ⁸	
Cr	3	1.66	2.6	0.0332	129 ⁸	
Co	5	1.2 ⁷	0.68 [*]	0.0202	2.55 ⁸	
Ni	30	2.5 ⁷	1.3 [*]	0.01	5.9 ⁸	
Cu [#]	0.12	0.57	0.21	0.11	0.098	
Rb	0.0001	0.026 ^a	0.004 ^b	1.75	0.0005	
Sr	0.0001	0.157	0.0034	1.61	0.0005	
Y	0.021	0.67 ^c	0.19	0.044 ^{**}	0.01	
Zr	0.008	0.195	0.021	0.0128	0.38	
Nb	0.0001	0.0081	0.027	0.005	0.01	
Ba	0.0001	0.0058	0.0063 ^c	0.811	0.0005	0.005
La	0.0001	0.0515	0.0019	0.0418	0.029	0.029
Ce	0.0001	0.108	0.0035	0.0302	0.038	0.038
Nd	0.0001	0.277	0.013	0.021 ^d	0.053	
Sm	0.0006	0.462	0.063	0.017	0.064	0.053
Eu	0.0007	0.458	0.059	0.166	0.061	0.02
Gd	0.001	[0.54]	0.069	[0.04]	[0.068]	
Tb	0.002	[0.6]	0.16	0.0095	0.078 ⁹	
Dy	0.003	0.711	0.15	[0.009]	[0.078]	
Ho	0.005	0.792 ^c	0.3	[0.0084]	[0.078]	
Er	0.008	0.66	0.24	[0.0078]	[0.077]	
Tm	0.022	[0.63]	0.48	[0.0073]	[0.077]	
Yb	0.019	0.633	0.39	0.0065	0.39	0.39
Lu	0.03	0.623	0.67	0.0068	0.0213 ⁹	0.47
Hf	0.013	0.223	0.19 [*]	0.0128	0.38	1.817

References: ¹McKay (1986); ²Hauri et al. (1994); ³Dunn and Senn (1994); ⁴Phinney and Morrison (1990); ⁵McKenzie and O'Nions (1991); ⁶Taken from Table 3 of Neal et al. (1989); ⁷McKay et al. (1986); ⁸Irving (1978); ⁹Nagasawa et al. (1980); ^{*}estimated from Weill and McKay (1975); [@]estimated from Dunn and Senn (1994); ^acalculated using Rb/Sr of McKenzie and O'Nions (1991); ^bcalculated using Rb/Sr of Arth (1976); ^cestimated from McKenzie and O'Nions (1991); ^dcalculated using Nd/Sm of Arth (1976); ^ecalculated using Ho/Er of Nicholls and Harris (1980); ^festimated from Dalpé and Baker (1993); [] represent D's determined by extrapolation from measured values of other REE's; [#]from Ewart and Griffin (1994); ^{**}estimated from Drake and Weill (1975); Many of the K_d values for spinel were taken from Table 3 of Neal et al. (1989).

group sample contained excessive olivine of a relatively primitive composition (i.e., the olivine was assumed to have an X_{Mg} value of 0.87, which equals the most primitive olivine analyzed in this study). In order to produce an X_{Mg} of 0.72, the maximum allowed by phase relations, it was necessary to remove 27% of this primitive olivine from the matrix. A similar mass balance approach was used to recalculate the trace-element concentrations. Table 1 gives the trace-element abundances of the border group sample and the calculated parent. Despite the recalculation of the parental magma for the ultramafic zone, the modeling results of Higgins et al. (1996) are not changed demonstrably. Indeed, when the new composition is entered into the program MAGFOX, olivine is still the first phase that crystallizes, followed by pyroxene and spinel. Moreover, there is greater than 40% crystallization of olivine prior to the addition of a new phase, consistent with the observed stratigraphy (Sharkov et al., 1995; Higgins et al., 1996).

The parental magma for the ultramafic zone is LREE-enriched and can be combined with a calculated dunite, with which it is in equilibrium (determined using the bulk D), to reproduce the observed trace-element variations in the samples (see Figure 6). Indeed, sample 200/444.7 can be produced by combining the calculated dunite with 22% of the parental liquid, and sample 248/190 can be produced by combining this dunite composition with 33% of the parental liquid. The percentage of parental liquid added is roughly the same as the modal abundance of intercumulus phases in each of these samples (i.e., 25% and 34%, respectively). A spidergram (Sun and McDonough, 1989) of these calculated values is shown in Figure 8. In general, these other trace-elements support the REE data. Note, however, that Sr, Zr, and Hf are

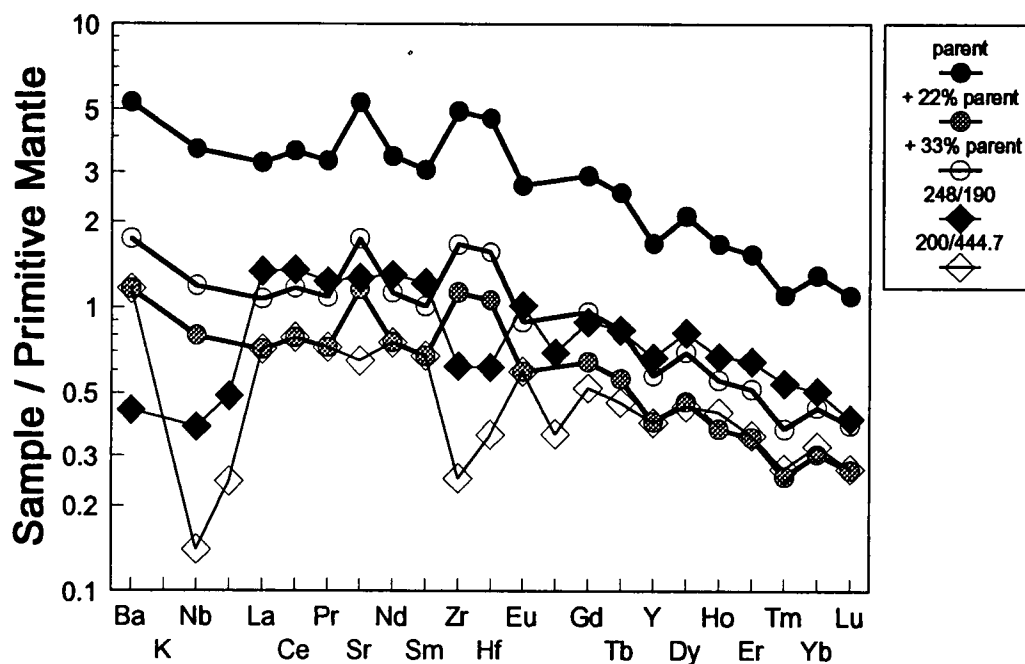


Figure 8. A spidergram (Sun and McDonough, 1989) showing the parental magma and selected samples for the ultramafic zone normalized to primitive mantle. Also shown are various combinations of the parent magma and a calculated dunite with which it is in equilibrium. Note that the dunite composition + 22% of the parent $\cong$ sample 200/444.7 and + 33% of the parent $\cong$ 248/190.

enriched in the parental melt relative to the samples. This reflects the fact that the border group sample (28/223) used to calculate the parent magma is slightly enriched in these elements.

The enrichment of sample 28/223 in Zr and Hf may be due to the presence of an oxide phase such as chromite or ilmenite. Chromite has K_d values > 0.2 , and ilmenite has K_d values > 1 , for both of these elements (Table 3). Indeed, sample 28/223 contains 4 wt.% chromite. Using a mass balance approach, 4 wt.% chromite was taken out, and the liquid recalculated. This did not affect the Zr and Hf levels demonstrably, however. The high K_d values for ilmenite indicate that relatively minor amounts could influence the Zr and Hf abundances in the calculated liquid. No ilmenite was found to be present in 28/223 petrographically, however, and a CIPW-norm calculation gives weight percent values < 1 . Regardless, even if two percent ilmenite is assumed to have fractionated from the liquid (and this is generous since the largest amount of ilmenite observed in samples from either portion of the BLI is only about one percent), the observed Zr and Hf enrichments of the calculated liquid cannot be erased. While we cannot state for certain why there is this enrichment in sample 28/223, this is the reason that Zr and Hf are high in the calculated parental liquid.

Calculated parent liquids for the gabbronorite zone

There is no chilled margin sample that was analyzed during this study that is representative of the parental magma(s) of the gabbronorite zone. Thus, the parent composition had to be estimated on the basis of the trace-element abundances in the samples. To accomplish this task, whole-rock trace-element concentrations were

combined with bulk distribution coefficients, calculated from modal analyses and K_d values (Table 3), in order to estimate the liquids in equilibrium with each of the samples (Figures 9 and 10). All of the calculated liquids are LREE-enriched, with depleted HREE giving the patterns a rather steep negative slope. The calculated liquid for sample 354/279 is more LREE-depleted than the other calculated melts, not unexpected since the sample itself is extremely LREE-depleted (see Figure 7). Note also that the calculated liquid for sample 353/140 is more LREE-enriched than those for the other samples, consistent with the earlier observation that this rock is anomalous. Sr abundances are relatively low in the calculated liquids (see Figure 10), and this is primarily due to the presence of plagioclase and augite in the samples (i.e., K_d values for Sr are 1.61 and 0.16, respectively, for these minerals). As to be expected, Sr abundances increase in the samples with increasing weight percentage plagioclase and augite. Nb is enriched in the liquids, and this is likely because it has very low K_d values (i.e., all are < 0.03 ; Table 3) for the main cumulus phases (augite, plagioclase, orthopyroxene). The low Ba values in the calculated liquids reflect a $K_d^{\text{min/melt}}$ of 0.8 for plagioclase. Finally, the high Zr and Hf values for the calculated liquid for sample 353/140 may reflect crustal contamination. This idea will be considered in detail below.

7. ORIGIN OF THE LAYERED SERIES

In part I of this study, Higgins et al. (1996) proposed an origin for the layered series based on major-element data which suggested that the gabbro-norite zone crystallized from the residue left after formation of the ultramafic zone. Unfortunately,

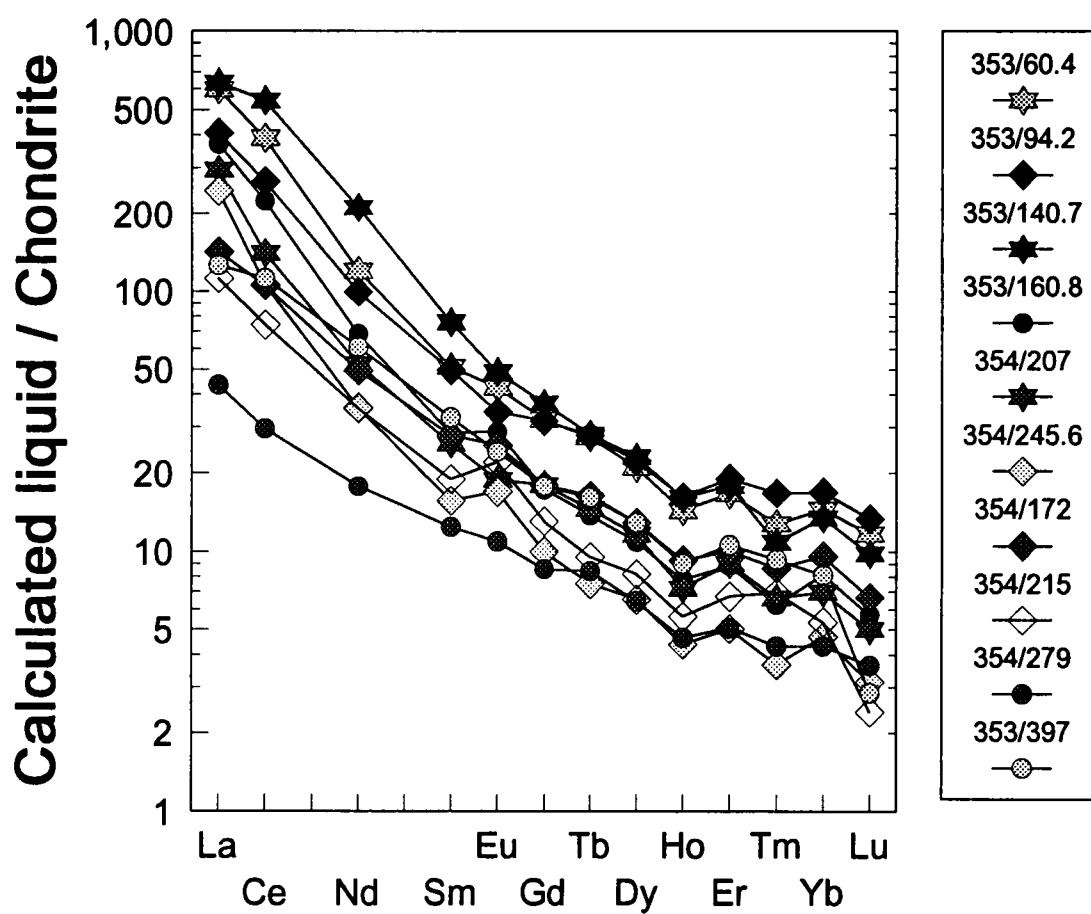


Figure 9. A chondrite-normalized REE plot of calculated liquids for selected samples from the gabbro-norite zone. Note that there are two groups of liquids delineated by different slopes for the LREE.

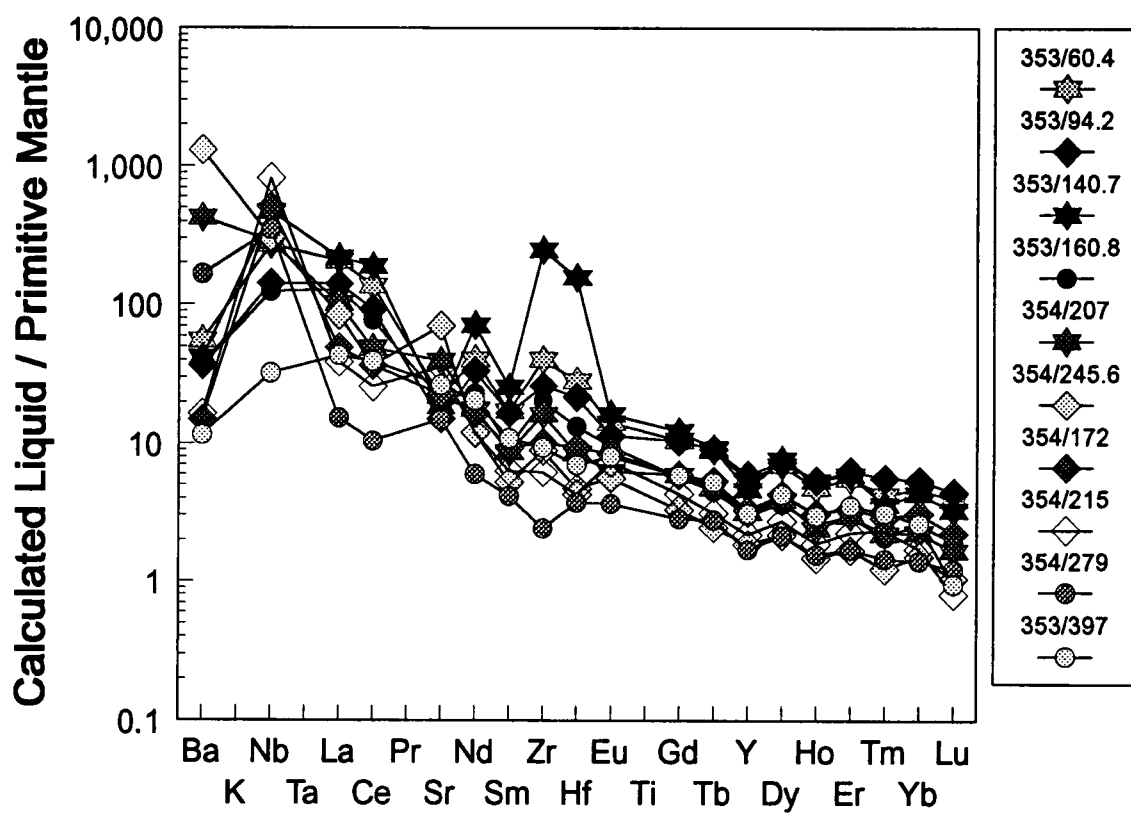


Figure 10. A spidergram (Sun and McDonough, 1989) showing the calculated liquids for selected samples from the gabbro-norite zone, normalized to primitive mantle. Note the high abundances of Zr and Hf in the parent liquid for sample 353/140.

the evidence provided by the major-elements to substantiate this scenario was weak. Another explanation given by Higgins et al. (1996) stated that the gabbro-norite zone may have formed from an entirely different magma than the ultramafic zone. REE characteristics of calculated liquids for both the ultramafic zone and gabbro-norite zone tend to support the second hypothesis. This can be appreciated by considering a plot of $(\text{La/Sm})_n$ versus $(\text{Dy/Yb})_n$ which shows that the ultramafic zone could not have formed from the same magma as the gabbro-norite zone (Figure 11). Indeed, the calculated gabbro-norite zone parental liquids form a distinct group from the ultramafic zone suggesting that the parent magma was not a fractionally crystallized derivative but was indeed of a different source region. The origins of these two zones are considered separately below.

Rare Earth Element evidence for origin of the ultramafic zone

Samples from the ultramafic zone are LREE-enriched and HREE-depleted. These types of REE patterns are not uncommon for layered mafic intrusions of early Proterozoic age. Indeed, Lambert and Simmons (1987) recognized similar patterns in samples from the Ultramafic Series of the Stillwater Complex. They suggested that the parental magmas of the Stillwater Ultramafic Series, which they termed U-type, were characterized by $(\text{Ce/Yb})_n$ ratios of 8 to 18, $(\text{Nd/Sm})_n$ ratios of 1.6 to 2.0, and $(\text{Dy/Yb})_n$ ratios of 1.7 to 1.9. In contrast, the parental magma calculated for the ultramafic zone of the BLI has a $(\text{Ce/Yb})_n$ ratio of 2.6, a $(\text{Nd/Sm})_n$ ratio of 1.1, and a $(\text{Dy/Yb})_n$ of 1.6. These data indicate that the BLI parental magma is neither as LREE-enriched as the U-type parent magmas of Lambert and Simmons (1987), nor is it as HREE depleted.

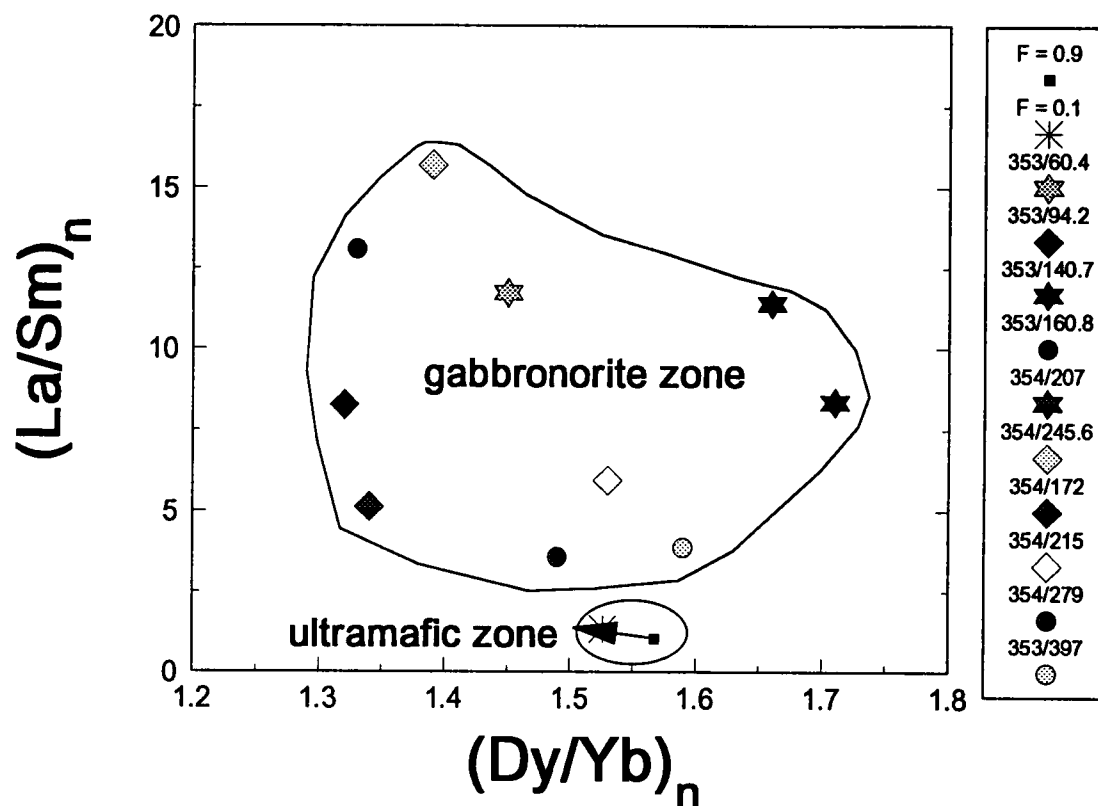


Figure 11. A plot of $(La/Sm)_n$ versus $(Dy/Yb)_n$ showing separate fields for calculated liquids of the gabbronorite zone and the ultramafic zone. The two values for the ultramafic zone represent the trace element abundances of the parental melt after 10% and 90% crystallization, respectively. Note that the trend of crystallization (denoted by the arrow) of this parental magma is not towards the field representing gabbronorite zone liquids.

However, Papike et al. (1995) performed SIMS analyses on orthopyroxene cores from the Stillwater Complex and determined that the LREE concentrations reported by Lambert and Simmons (1987) were too high. They believed this to be due to one or more contaminating phases (e.g., plagioclase, augite) being present in the orthopyroxene mineral separates used by Lambert and Simmons (1987) for isotope dilution analyses. Parental liquids, determined with the data of Papike et al. (1995), have $(\text{Ce/Yb})_n$ ratios of 2.9 to 13.8, $(\text{Nd/Sm})_n$ ratios of 0.8 to 1.3, and $(\text{Dy/Yb})_n$ ratios of 1.1 to 1.9. These calculated liquids are similar to the REE characteristics of the parent magma calculated for the ultramafic zone of the BLI (see Figure 6). Note, however, that the BLI parental magma plots at the least abundant REE portion of the U-type parental magma field. Regardless, the important observation to be made here is that the REE patterns for the parental magma of the ultramafic zone of the BLI and that of the Stillwater ultramafic rocks are both LREE enriched to similar degrees. This suggests that they may have formed by similar processes.

The LREE-enriched nature of the proposed parental magma for the ultramafic zone of the BLI suggests that it may have formed from partial melting of a large-ion lithophile element enriched mantle source. Alternatively, these LREE characteristics are a product of contamination of the parent magma by LREE-enriched crustal rocks in route to the BLI magma chamber. However, the ϵ_{Nd} of the BLI is -2.0 (Bayanova et al., 1993), suggesting that significant crustal contamination did not occur (e.g., typical Archean gneisses have ϵ_{Nd} values of -30 to -40; Collerson and McCulloch, 1982). Moreover, layered intrusions of the Baltic Shield of similar age to the BLI have ϵ_{Nd} values ranging from -2.3 to -2.0 (e.g., the Mt. General'skaya at 2453 ± 42 Ma and

Imandra plutons at 2444 ± 77 Ma, respectively; Balashov et al., 1993). If crustal contamination were the main cause of the LREE-enriched characteristics of these intrusions, than one would expect there to be more variability in the ϵ_{Nd} values over such a large region (i.e., the Baltic Shield). Thus, the negative ϵ_{Nd} values of the BLI are likely a product of derivation of the parent magma(s) from a mantle source that was enriched in LREE for a significant period of time, rather than a result of crustal contamination. The next question concerns the nature of the processes which could have enriched the mantle.

Lambert and Simmons (1987) proposed that the LREE-enrichment of the U-type magma for the Stillwater Complex may have been due to partial melting of an upper mantle containing veins enriched in LREE. The HREE characteristics of the U-type magma were explained by the presence of residual garnet throughout the melting process in the mantle (Lambert and Simmons, 1987). They further proposed that the melting of the source was a multi-stage, dynamic process. The similarity of the REE patterns for the BLI parent with the U-type parent of the Stillwater indicates that a comparable explanation may be offered. One problem, however, involves the multi-stage dynamic partial melting model proposed by Lambert and Simmons (1987). Multi-stage dynamic partial melting of a garnet-bearing source should produce REE patterns which fan out from Yb, with early melts showing the largest enrichment in LREE followed by progressively lower REE abundances in subsequent liquids (Loferski et al., 1994). In the ultramafic zone of the BLI, the REE patterns are subparallel (see Figure 6), suggesting formation through simple fractional

crystallization of a single melt, rather than as a product of multi-staged dynamic partial melting.

Metasomatism is another mechanism that can be argued to explain the origin of the LREE-enriched and HREE-depleted signatures in the parent magma of the ultramafic zone. This has been suggested to explain similar REE characteristics determined for samples from the Stillwater Complex (Loferski et al., 1994). Metasomatic fluids are normally enriched in LREE relative to HREE, primarily because the HREE are slightly more compatible. As such, the interaction of these fluids with a source region would result in LREE-enrichment. In addition, partial melts of this LREE-enriched source would have similar REE characteristics to the BLI magmas. Based upon this reasoning, metasomatism seems the most likely mechanism for enriching the mantle source region.

An origin proposed for the ultramafic zone, on the basis of major-element data (Higgins et al., 1996), involves an approach initially used to describe the petrogenesis of the Muskox Intrusion (Irvine and Smith, 1967; Irvine, 1970). This model entails the repeated addition of undifferentiated magma during accumulation. The residual magma is "flushed" from the system (effectively pushed out) by each addition of new magma. It was suggested in Higgins et al. (1996), on the basis of similar ages (i.e., 2500 ± 70 Ma versus 2449 ± 1.1 Ma for the BLI), that the Karelian volcanic fields which surround the BLI may be products of this "flushing". This idea can be more carefully evaluated by comparing chondrite-normalized REE plots for a suite of early Proterozoic Karelian komatiites (Bogatikov, 1988) with calculated liquids for the ultramafic zone (Figure 12). It is apparent from Figure 12 that these volcanic rocks are not related

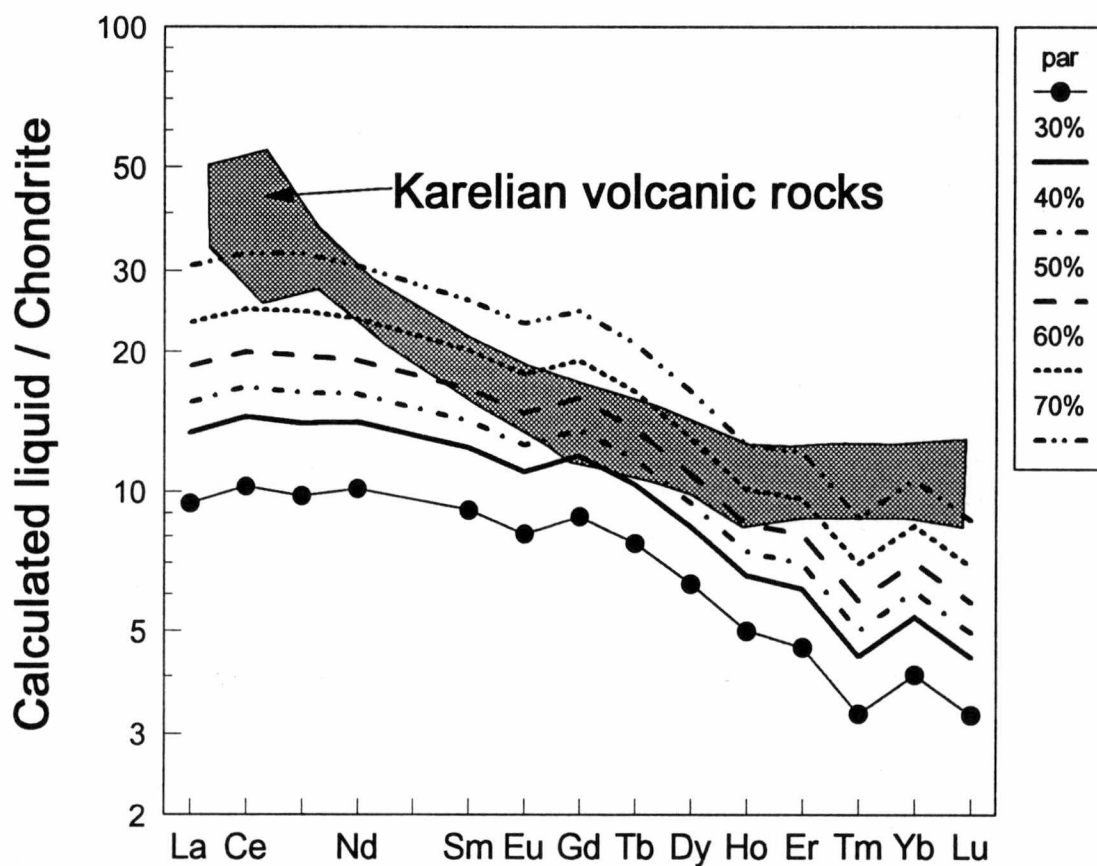


Figure 12. A chondrite-normalized REE plot comparing the ultramafic zone parent magma at various stages of crystallization with early Proterozoic Karel'ian volcanic rocks (shaded field; data are from Bogat'kov, 1988). Note that the field for the volcanic rocks cuts across the REE patterns for the calculated liquids indicating that the two are not related.

either to the calculated parent liquid of the ultramafic zone or to various values of F (i.e., the fraction of melt remaining) for this liquid. That is, the Karelian rocks are more enriched in LREE and are less depleted in the HREE relative to the ultramafic zone samples. On the basis of these data, it can be concluded that the early Proterozoic Karelian volcanic rocks are not related to the episode of magmatism which produced the ultramafic zone of the BLI. This does not imply that there are no volcanic rocks in Karelia related to this magmatism, but that none have been reported to date. Future work will entail a detailed petrologic study of volcanic rocks from Karelia with the possibility of finding a suitable match for the parent magma of the ultramafic zone.

Rare Earth Element evidence for origin of the gabbro-norite zone

Samples from the gabbro-norite zone have REE patterns with negative slopes (see Figure 7), implying that, as with the ultramafic zone, they crystallized from a magma that was LREE-enriched. The question that should be considered next is whether the gabbro-norite zone crystallized from one single magma or from several. If the gabbro-norite zone represents the crystallization product of a single magma, then the chondrite-normalized REE patterns for these liquids should be parallel, and there should be increasing REE abundances upward through the section. This is due to the fact that the mineral-melt partition coefficients for the REE in the major cumulus minerals of this zone (e.g., augite, orthopyroxene, plagioclase) are all less than one (Table 3). However, *the abundances of REE in samples from the gabbro-norite zone are inconsistent with simple fractional crystallization from a single magma.* In fact, there are at least two different parent magmas for the gabbro-norite zone as reflected in a plot

of $(\text{La/Nd})_n$ versus La (Figure 13). These parental magmas are denoted here as Group I and Group II. These two groups delineate calculated liquids with similar REE characteristics for samples from the lower and upper portions, respectively.

Parental liquids for samples from Group I are characterized by $(\text{La/Nd})_n$ ratios ranging from 2.1 to 3.2. In contrast, Group II parental melts are from higher up, stratigraphically, and have $(\text{La/Nd})_n$ ratios ranging from 4.1 to 6.9. In addition, Group I liquids have $(\text{Ce/Yb})_n$ ratios of 6.9 to 13.9, $(\text{Nd/Sm})_n$ ratios of 1.4 to 1.9, and $(\text{Dy/Yb})_n$ ratios of 1.3 to 1.6. Group II liquids have $(\text{Ce/Yb})_n$ ratios of 15.8 to 27.3, $(\text{Nd/Sm})_n$ ratios of 2.0 to 2.4, and $(\text{Dy/Yb})_n$ ratios of 1.3 to 1.7. It is proposed here, on the basis of these REE characteristics, that the lower subzone of the gabbro-norite zone be further subdivided into an upper and lower part.

As mentioned above, the REE abundances of samples formed from a single magma should become more enriched with increasing fractionation. Consider Group I and Group II magmas separately (Figures 14a and 14b, respectively). It is apparent from Figure 14 that Group II is considerably more enriched in the LREE than Group I (i.e., La is about 2 to 10X as abundant). Chondrite-normalized REE patterns for Group II liquids are all subparallel to parallel, and the most enriched liquid occurs at the highest level, stratigraphically. These characteristics suggest that the samples of Group II are fractionally crystallized derivatives of one parent magma. Group I liquids, with the exception of 354/279, also have subparallel to parallel chondrite-normalized REE patterns. The liquid for sample 354/279 has a flatter LREE pattern, which suggests it was influenced by some additional process. This sample will be considered in more detail below. The calculated liquid for 353/397 is more enriched in the REE than the

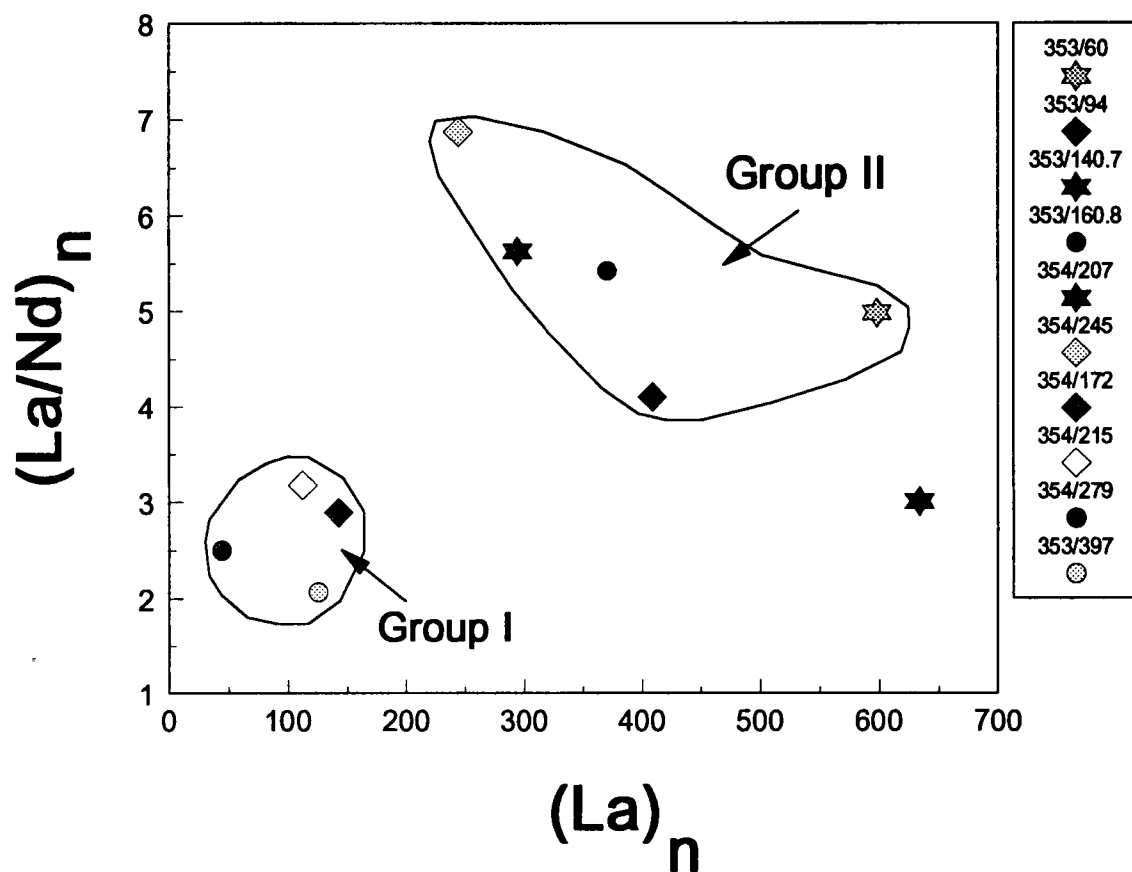


Figure 13. $(La/Nd)_n$ versus $(La)_n$ for calculated parent liquids of the gabbro-norite zone. Note that two distinctive groups are delineated indicating that there were at least two parent magmas involved in the formation of the gabbro-norite zone.

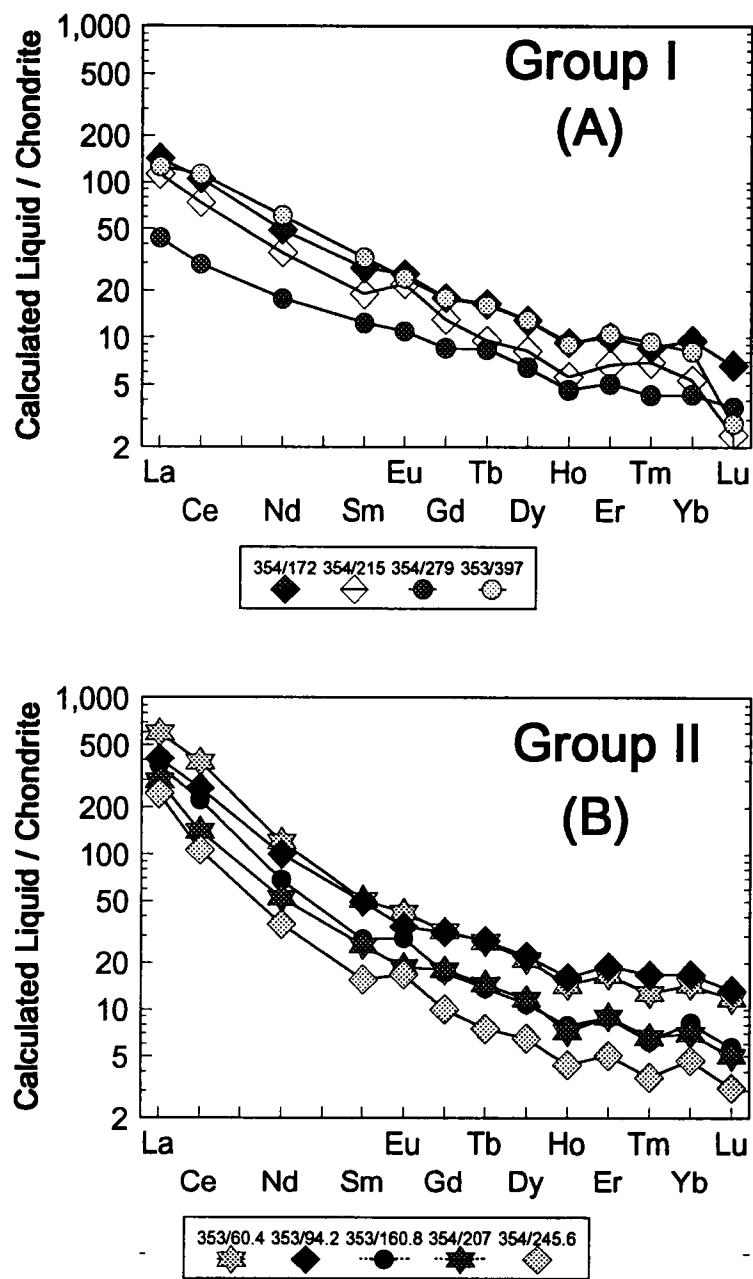


Figure 14. Chondrite-normalized REE plots showing the two groups of parental magmas for the gabbronorite zone. (A) Group I. (B) Group II. Note the difference in REE abundances between the two groups.

liquids for samples from higher up in the sequence. It is possible that there may have been a pulse of new material following crystallization of sample 353/397. This is also supported by the major element chemistry in which the X_{Mg} changes from 0.77 to 0.81 from sample 353/397 to sample 354/279, respectively (Higgins et al., 1996).

Sample 354/279 is strongly depleted in the LREE (see Figure 7), which suggests that it may have formed from a relatively ultramafic liquid that could have come from the ultramafic zone. Indeed, the calculated parent magma for 354/279 has a similar chondrite-normalized REE pattern, albeit more enriched by a factor of 1.2 to 1.5, to sample 28/223 from the border group (Figure 15). If 28/223 is assumed to represent a liquid composition for the ultramafic zone, the calculated magma for 354/279 can be reproduced through 30% fractional crystallization of this liquid (see Figure 15). This suggests that some of the liquid composition, represented by 28/223, prior to solidification, was actually injected into the gabbro-norite zone. This magma crystallized to form sample 354/279. A similar model could explain the origin of all of the olivine-pyroxenite-harzburgite layers observed in the lower portions of the gabbro-norite zone (Sharkov et al., 1995; Higgins et al., 1995). In addition, major-element data indicate that these layers correspond to reversals in magma chemistry to more Mg-rich compositions (e.g., X_{Mg} for orthopyroxene varies from 0.74 to 0.85 at harzburgite horizon I in the lower part of the gabbro-norite zone; Sharkov et al., 1996). As such, this provides further evidence to support pulses of material derived from the ultramafic zone. In order to better constrain these ideas, however, additional analyses are needed of these olivine-pyroxenite/harzburgite layers, including both major- and trace-elements, beyond the scope of the present study.

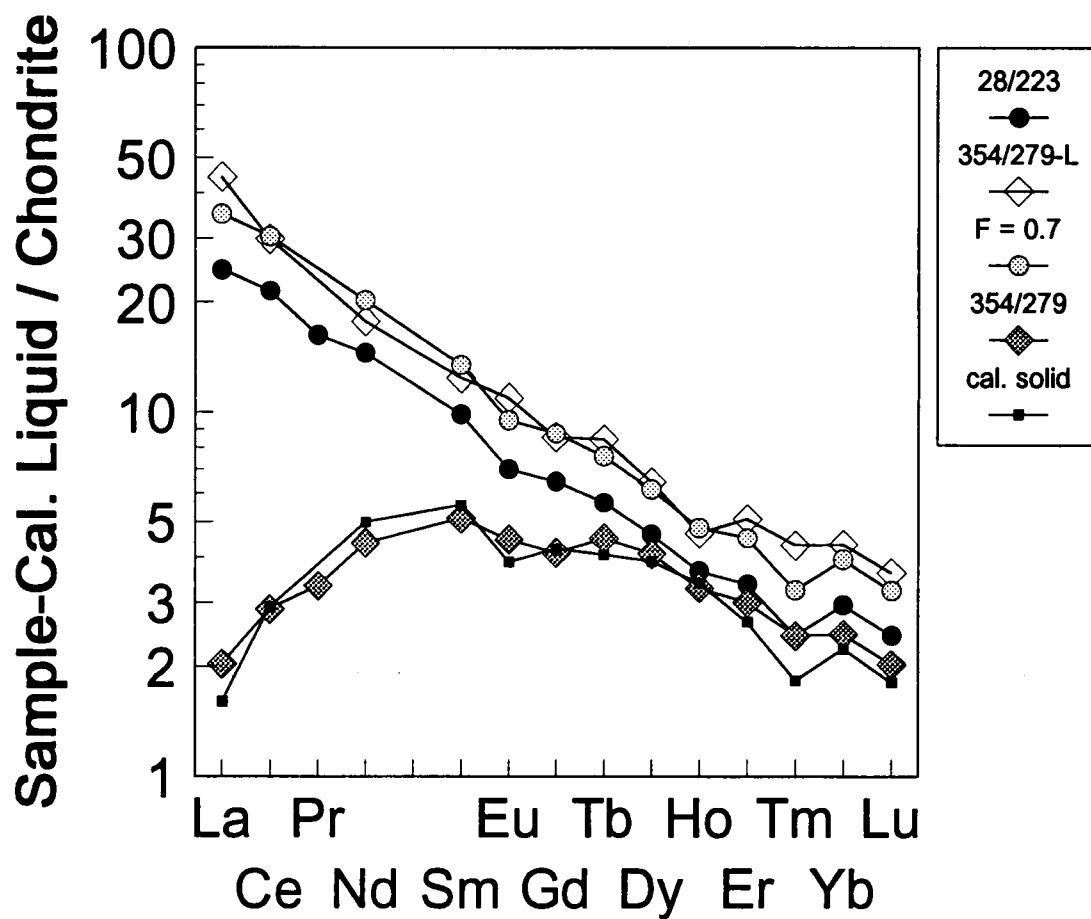


Figure 15. A chondrite-normalized REE plot showing the relationship of the border group sample, 28/223, to the parent liquid for sample 354/279 from the gabbro-norite zone. Note that with 30% crystallization of a liquid with the composition of sample 28/223, the parent liquid for 354/279 can be reproduced.

Sample 353/140 has a REE pattern that is distinctly different from the other samples of the gabbro-norite zone (see Figure 7). Indeed, the parental liquid calculated from sample 353/140 has a $(\text{Ce/Yb})_n$ ratio of 40.4, a $(\text{Nd/Sm})_n$ ratio of 2.8, and a (Dy/Yb) ratio of 1.7, all of which are higher than values determined for liquids in equilibrium with the other samples. Moreover, this liquid also has much higher Zr and Hf values than the other gabbro-norite zone liquids. As noted earlier, this sample is anomalous and does not fit with the proposed groups for this zone (see Figure 13). One possible reason for the low HREE values could be the source region for this liquid. Perhaps, sample 353/140 crystallized from a liquid that was derived from a garnet-rich source, which would account for the HREE-depleted signature. However, there is no evidence that suggests why the source region for this sample would be so much different than that for the other samples of the gabbro-norite zone. Another possible explanation is that the parental liquid for sample 353/140 may have been contaminated by Archean crust. In fact, using a simple mass-balance calculation, the observed REE patterns and the high Zr and Hf values in this melt can be produced through contamination with 5 to 10% of a typical felsic Archean crustal rock (Rudnick and Fountain, 1995).

It was demonstrated above that the early Proterozoic Karelian volcanic rocks were not the extrusive equivalents of the ultramafic zone. Similarly, a plot of $(\text{La/Sm})_n$ versus $(\text{Dy/Yb})_n$ reveals that they form a separate and distinct field from the gabbro-norite zone liquids (Figure 16). Thus, at the present time, there is no extrusive analog that has been identified. As suggested by Amelin et al. (1995), the BLI might

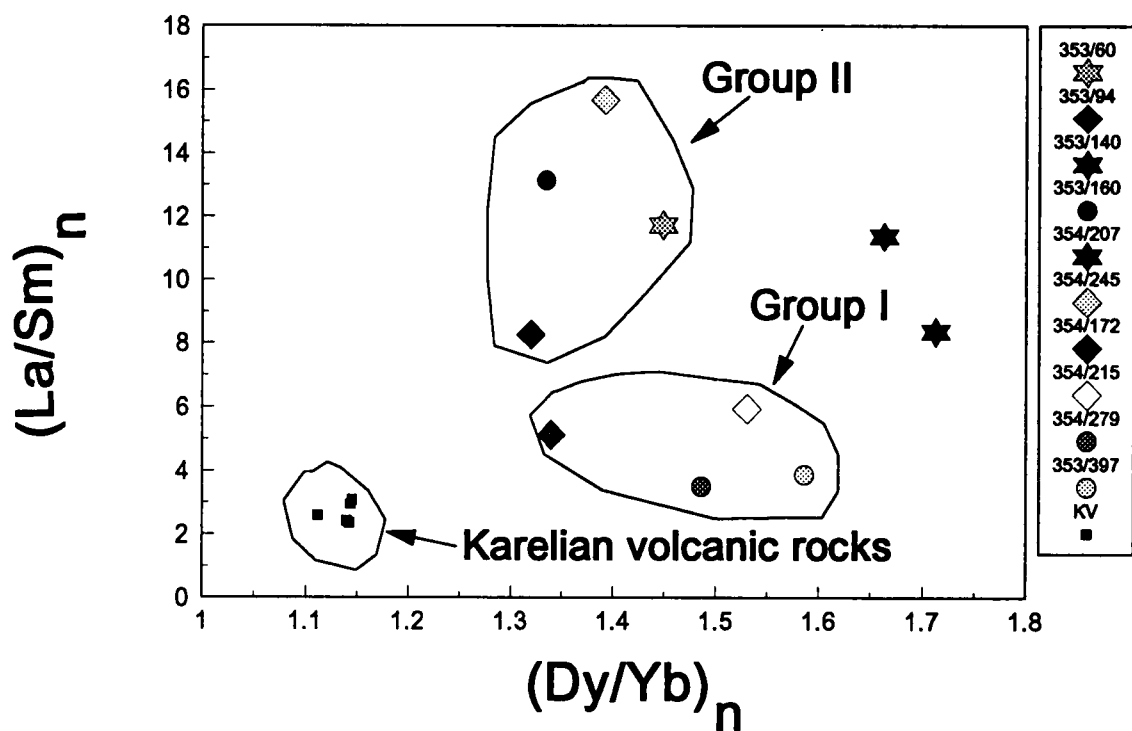


Figure 16. A plot of $(La/Sm)_n$ versus $(Dy/Yb)_n$ showing three different fields, one for Karelian volcanic rocks (data are from Bogatkov, 1988), and one for each of the two sections of the gabbonorite zone. Note that, as with the ultramafic zone, the early Proterozoic Karelian volcanic rocks could not have formed from the same magma as the gabbonorite zone.

represent an early magmatic episode for which any related volcanic rocks are either completely eroded or not yet exposed.

Composition of the mantle during the Archean-Proterozoic transition: REE evidence

Despite the fact that the volcanic rocks which are coeval with the BLI do not have similar REE abundances to the calculated liquids of either the ultramafic or gabbro-norite zones, the LREE-enrichment of both sets of rocks can be used to infer something about the mantle beneath the Baltic Shield during the early Proterozoic-Archean transition. Indeed, it is likely that the mantle under the Baltic Shield during this time was LREE-enriched. This is evidenced not only by the occurrence of volcanic rocks in Karelia with these patterns, but also by the presence of similarly LREE-enriched volcanic rocks in the northeastern Baltic Shield (Melezhik and Sturt, 1994). Moreover, layered intrusions of the northeastern Baltic Shield (e.g., the Olanga Complex, the Monchegorsk, the Pana tundra), like the BLI, tend to be characterized by LREE-enriched patterns (Melezhik and Sturt, 1994), and as already mentioned, their ϵ_{Nd} values are negative as expected if they were derived from an enriched mantle. Finally, these LREE-enriched characteristics are not confined to early Proterozoic layered intrusions of the Baltic Shield. Indeed, similar REE patterns have been observed for Archean-Proterozoic intrusions in other parts of the world (e.g., the Stillwater and Bushveld complexes; Lambert and Simmons, 1987 and Sharpe and Hulbert, 1985; respectively), which suggests that the LREE-enrichment which affected the mantle beneath the Baltic Shield may not have been a localized event, but perhaps occurred on a world-wide scale.

8. SUMMARY AND CONCLUSIONS

Trace-element abundances in whole-rocks from the ultramafic and gabbro-norite zones of the BLI have defined certain geochemical characteristics that must be explained in any interpretation of the petrogenesis of this large pluton. Samples from the ultramafic zone are LREE-enriched with $(\text{Ce/Yb})_n$ ratios of 2.2 to 2.5, $(\text{Nd/Sm})_n$ ratios of 1.1 to 1.2, and $(\text{Dy/Yb})_n$ ratios of 1.3 to 1.6. These REE characteristics are consistent with the ultramafic zone having formed from a LREE-enriched magma. It is suggested that the chilled margin of the BLI, when recalculated to account for excess olivine, represents the parental magma of the ultramafic zone. This magma has the following geochemical characteristics: a $(\text{Ce/Yb})_n$ of 2.6, a $(\text{Nd/Sm})_n$ ratio of 1.1, and a $(\text{Dy/Yb})_n$ ratio of 1.6. The LREE-enrichment in the calculated parent magma suggests that the mantle source region was either LREE-enriched or that the liquid was contaminated by LREE-enriched crustal rocks in route to the BLI magma chamber. Negative ϵ_{Nd} values for rocks from the BLI are most consistent with a mantle source region that was LREE-enriched for a significant period of geologic time prior to the partial melting episode which produced the parental magma(s). This enrichment of the mantle source region likely resulted from metasomatism.

Samples from the gabbro-norite zone are LREE-enriched and are characterized by $(\text{Ce/Yb})_n$ ratios of 2.1 to 6.8, $(\text{Nd/Sm})_n$ ratios of 1.1 to 1.5, and $(\text{Dy/Yb})_n$ ratios of 1.2 to 1.7. Liquids in equilibrium with these samples have $(\text{Ce/Yb})_n$ ratios of 11 to 27, $(\text{Nd/Sm})_n$ ratios of 1.8 to 2.4, and $(\text{Dy/Yb})_n$ ratios of 1.3 to 1.7. At least two different parental magmas were involved in the formation of the lower subzone of the

gabbro-norite zone as evidenced by the variation in REE for calculated parental liquids for samples of different stratigraphic height. Group I samples, from the lower portion of the gabbro-norite zone, are characterized by parental liquids with (La/Nd) ratios of 2.1 to 3.2, (Ce/Yb)_n ratios of 6.9 to 13.9, and (Dy/Yb)_n ratios of 1.3 to 1.6. In contrast, Group II samples, from the upper part of the gabbro-norite zone, are characterized by (La/Nd)_n ratios of 4.1 to 6.9, (Ce/Yb)_n ratios of 15.8 to 27.3, and (Dy/Yb)_n ratios of 1.3 to 1.7.

The similarities in REE characteristics between the parent liquid of sample 354/279 (an olivine pyroxenite from the lower part of the gabbro-norite zone) and sample 28/223 (thought to represent a liquid composition for the ultramafic zone) suggest that the olivine-pyroxenite/harzburgite layers that occur in the lower portions of the gabbro-norite zone likely crystallized from liquids derived from the ultramafic zone. This provides evidence that the source of magma for the ultramafic zone was not exhausted after its formation, but continued to periodically intrude pulses of material to the gabbro-norite zone.

The REE characteristics of the ultramafic and gabbro-norite zones are consistent with derivation from a LREE-enriched source. Moreover, volcanic rocks of similar age in Karelia and the northeastern Baltic Shield are similarly LREE-enriched. At the present time, however, there are no volcanic rocks that have been analyzed that appear to be related to the magmatism that formed the BLI. This implies that the magma which crystallized to form the BLI either never made it to the surface or that it may have been eroded away. Despite this, the fact that the majority of early Proterozoic aged rocks from the Baltic Shield have LREE-enriched signatures indicates that the

mantle during this time was likely LREE-enriched. Moreover, the narrow range in ϵ_{Nd} values for other early Proterozoic layered intrusions of this region (e.g., ϵ_{Nd} values ranging from -2.3 to -2.0) suggests that crustal contamination was not a major factor in producing the observed LREE-enrichments. Finally, intrusions of similar ages occurring on other cratons (e.g., the Stillwater Complex, the Bushveld Complex) display these same types of LREE-enriched patterns. This may indicate that the LREE-enrichment was not a localized event, but instead occurred on a worldwide basis.

Future work will entail a more detailed petrologic study of the ultramafic zone and also of the volcanic rocks which surround the BLI. Moreover, samples of the other layers within the BLI (i.e., the pyroxenite zone, the pigeonite-gabbro-norite zone, and the magnetite-gabbro-norite zone), that have yet to be studied in detail, will be analyzed in order to further constrain the origin of this important Russian intrusion. Finally, additional research dealing with other Precambrian intrusions of the Baltic Shield (e.g., the Monchegorsk) is currently underway by our group. These studies, when combined with data from the BLI, should provide a more accurate picture of the magmatic events occurring on the Baltic Shield during the Archean-Proterozoic transition.

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SUMMARY AND CONCLUSIONS

1. SUMMARY AND CONCLUSIONS

The Burakovsky layered intrusion of southern Karelia, Russia is the largest layered mafic intrusion in the Baltic Shield with a surface area of over 700 km². The intrusion has been divided into a layered series and a border group. The layered series has been further subdivided into: (1) ultramafic zone (~ 3000 meters); (2) pyroxenite zone (~ 200 meters); (3) gabbro-norite zone (~ 1000 meters); (4) pigeonite gabbro-norite zone (~ 1200 meters); and a (5) magnetite gabbro-diorite zone (~ 800 meters). Major-element variations in rocks from the BLI indicate a progression towards more evolved (e.g., Fe-rich) compositions upward in the intrusion, consistent with an origin by fractional crystallization. These variations are evidenced by variations in the Fo content of olivine from the ultramafic zone of 89 to 83. Moreover, orthopyroxene cumulates from the gabbro-norite zone reveal similar trends, with X_{Mg} decreasing from 0.84 to 0.58 upward. The variations observed in the ultramafic zone are much smaller and occur over a much wider interval (i.e., 1600 meters compared with 500 meters, respectively) than those in the gabbro-norite zone. This suggests either that the rate of replenishment was much higher or that the volume of magma was much greater in the ultramafic zone relative to the gabbro-norite zone.

Major-element modeling with the program MAGFOX and a starting composition represented by a sample from the border group reveals > 40% crystallization of olivine prior to the addition of a new phase. This is broadly consistent with the stratigraphy of the ultramafic zone of BLI in which the lower 2/3 is composed of olivine cumulates. Furthermore, the major-element data suggest that the gabbro-norite zone may not have

formed from the same magma as the ultramafic zone, indicating that there may have been at least two parent liquids for the BLI. REE data for these two zones support this idea. Indeed, parent liquids for samples from the gabbro-norite zone are much more LREE-enriched than the proposed parent magma for the ultramafic zone (e.g., $(\text{Ce/Yb})_n$ of 6.9 to 27.3 and 2.6, respectively).

REE data for samples from the gabbro-norite zone reveal that there were at least two parental magmas involved in the formation of the lower subzone. These magmas are delineated as Group I, from the lower portion, and Group II from the upper portion. Calculated liquids for samples of Group II are more LREE-enriched than liquids for Group I samples (e.g., $(\text{Ce/Yb})_n$ of 15.8 compared with 6.9, respectively). The occurrence of several olivine pyroxenite/harzburgerite layers within the portion of the gabbro-norite zone delineated by Group I samples, indicate that the source of the magma for the ultramafic zone was not exhausted after its formation, but continued to periodically intrude pulses of material. Evidence to support this idea is provided by similar REE characteristics between sample 354/279, an olivine pyroxenite from the gabbro-norite zone, and sample 28/223, believed to represent a liquid composition for the ultramafic zone.

The REE characteristics of parental liquids for the ultramafic and gabbro-norite zones are consistent with derivation from a LREE-enriched source. Moreover, the BLI and other layered intrusions on the Baltic Shield of similar age (e.g., the Mt. General'skaya and Imandra plutons) have ϵ_{Nd} values ranging from -2.3 to -2.0, providing further evidence to support this idea. Finally, Archean-Proterozoic layered intrusions on other cratons (e.g., the Stillwater and Bushveld complexes) also have these same types of

REE patterns, indicating that the LREE-enrichment was not confined to the mantle beneath the Baltic Shield, but instead may have occurred on a worldwide basis.

VITA

Stefan John Higgins was born in Carrollton, Georgia on February 11, 1970. He attended Central Primary and Elementary Schools, both members of the Carroll County School System. His high school education was received at Carrollton High School, a part of the Carrollton City School System. Following high school graduation in June, 1988, Stefan began college at the University of Arizona in Tuscon, Arizona in July, 1988. In June of 1990, he transferred to West Georgia College, Carrollton, Georgia where he completed a Bachelor of Science degree in Geology in June, 1993. In August, 1993, Stefan started graduate studies in the Department of Geological Sciences at the University of Tennessee, Knoxville. He submitted the first part of this thesis for publication in *The Canadian Journal of Earth Sciences* in January, 1996. Stefan also published a paper entitled "X-ray digital-imaging petrography: Technique development for lunar mare soils" in the May, 1996 issue of *Meteoritics & Planetary Science*. He received the degree of Master of Science with a major in Geology in May, 1996.