

Timing and duration of Palaeoproterozoic events producing ore-bearing layered intrusions of the Baltic Shield: metallogenic, petrological and geodynamic implications

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Abstract

There are two 300-500-km long belts of Palaeoproterozoic layered intrusions: the Northern (Kola) Belt and the Southern (Fenno-Karelian) Belt in the Baltic (Fennoscandian) Shield. New U-Pb (TIMS) ages and radiogenic isotopic (Nd-Sr-He) data have been determined for mafic-ultramafic Cu-Ni-Ti-Cr and PGE-bearing layered intrusions of the Kola Belt. U-Pb ages on zircon and baddeleyite for gabbro-norite and anorthosite from the Fedorovo-Pansky, Monchepluton and Main Ridge (Monchetundra and Chunutundra), Mt. Generalskaya intrusions and gabbro-norite and dykes from the Imandra lopolith of the Kola Belt define a time interval of more than 130 million years, from ca. 2.52 Ga to 2.39 Ga. At least four intrusive phases have been distinguished: three PGE-bearing, and one barren. This spread of ages is wider than that for intrusions of the Fenno-Karelian Belt which clusters at 2.44 Ga. Nd isotopic values for the Northern Belt range from -1.1 to -2.4, implying an enriched mantle “EM-1 type” reservoir for these layered intrusions. Initial Sr isotopic data for the rocks of the intrusions are radiogenic relative to bulk mantle, with I_{Sr} values from 0.703 to 0.704. Geochemical data and $^4\text{He}/^3\text{He}$ isotopic ratios of the minerals reflect a significant contribution from a mantle source rather than the influence of crustal processes during emplacement. The geological and geochronological data indicate that in the eastern part of the Baltic Shield, mafic – ultramafic intrusive magmatism was active over a protracted period and was related to plume magmatism associated with continental breakup that also involved the Superior and Wyoming provinces.

Key words: geochronology, isotopes, Palaeoproterozoic, layered PGE-bearing intrusions, Baltic Shield.

There are about twenty Palaeoproterozoic layered mafic-ultramafic bodies in Finland, most of which occur in a roughly eastwest-trending, 300-km long belt known as the Tornio-Näränkäväära Belt (Alapieti et al., 1990; Vogel et al., 1998; Iljina & Hanski, 2005). The belt (Fig. 1) extends for a few kilometres into Sweden (Tornio intrusion), and for several tens of kilometres into the Russian Karelia (Olanga complex). Together the intrusions make up the Southern, or Fenno-Karelian Belt – FKB (Mitrofanov et al., 1997).

In the northeast of the province, the Northern, or Kola Belt (KB) strikes northwestwards for about 500 km (Fig.1). It includes more than ten isolated layered mafic-ultramafic bodies that are mostly ore-bearing (Mitrofanov et al., 1997). The central part of the Kola Belt has been suggested to be part of a triple junction typical of intraplate rifting (Pirajno, 2007), and is occupied by the Monchegorsk Layered Complex with a fairly complete range of ore types (Cr, Cu, Ni, Co, Ti, V, Pt,

Pd, Rh). The western and eastern arms of the triple junction are composed of large anorthosite-troctolite (Main Ridge, Pyrrshin, Kolvitsa) intrusions (Fig. 1). The most typical PGE-bearing layered pyroxenite-norite-gabbro-anorthosite intrusions of the Kola Belt (Mt. General'skaya, Monchegorsk Layered Complex, Fedorovo-Pansky etc.) are confined to boundaries between Early Proterozoic rifts which were in-filled with volcano-sedimentary rocks overlying the Archaean basement (Schissel et al., 2002; Mitrofanov et al., 2005). In these cases, similar to those in Finland, the intrusive rocks underwent relatively low grade local metamorphism and preserve cumulus and intracumulus minerals.

Convincing arguments in support of the mantle plume hypothesis, either as “shallow plumes” (from ~670 km) or “deep plumes” (from the core-mantle boundary) have been put forward for relatively young well-preserved Palaeozoic and recent large igneous provinces (LIPs) (Coffin & Eldhom, 1994; Heaman, 1997; Ernst & Buchan, 2003; French et al., 2008). Voluminous magmatism is considered to be related to mantle plumes that occurred throughout the Precambrian (Condie, 2001). The best records of a plume source are evident in Neoproterozoic and Palaeoproterozoic magmatic events and often have associated with mineralization (Sharkov et al., 2000; Campbell, 2001; Pirajno, 2007 etc.). Many of these magmas have generated layered mafic-ultramafic complexes with exploitable PGE deposits (Pt, Pd and Rh), and Cr, Ni, Cu, Co, Ti, V mineralization (Li & Naldrett, 1993; Mitrofanov et al., 1997; 2005; Sharkov et al., 2000). Magmatic igneous ore systems are present in the Windimurra (2.8 Ga), Stillwater (2.7 Ga), Bushveld (2.06 Ga), and Duluth (1.1 Ga) complexes (Li & Naldrett, 1993; Carr et al., 1999; Eules & Cawthorn, 1995; Pirajno, 2007). Many of these intrusions include massif-type anorthosites and troctolites enriched in Cr, Ni, Cu, PGE, Ti, V and Fe. All the above mentioned rock types, metals, host rift-associated volcanic rocks, and mafic dykes are found in the relatively recently described East-Scandinavian Palaeoproterozoic Large Igneous Province (ESCLIP) (Iljina & Hanski, 2005) with a total area of more than 200,000 km² (Fig. 1). In Finland, these geological complexes have been widely studied (Alapieti et al., 1990; Huhma et al., 1990; Vogel et al., 1998; Hanski et al., 2001), and the data were summarized by M. Iljina and E. Hanski (2005). Only a few publications on similar Russian complexes of the Baltic Shield have been written or translated into English (Papunen & Gorbunov et al., 1985; Balashov et al., 1993; Bayanova & Balashov, 1995; Amelin et al., 1995; Mitrofanov et al., 1997; Mitrofanov & Bayanova, 1999; Chashchin et al., 2002; Schissel et al., 2002; Mitrofanov et al., 2005).

This paper presents a brief geological description of the Russian mafic-ultramafic intrusions of the Baltic Shield and associated mineralization. It focuses on new U-Pb (TIMS) and Sm-Nd geochronological data which constrain timing of magmatic pulses and the duration of the emplacement of Cr, Cu, Ni, Ti and PGE-bearing layered intrusions of the Kola Belt. Nd, Sr and He isotope data help define geodynamic models for a long-lived Early Precambrian mantle source expressed either in a large mantle diapir or multiple plume processes for one of the earliest clearly identifiable old intraplate LIPs and its metallogeny.

Geological setting of ore-bearing intrusions of the Baltic Shield

Palaeoproterozoic layered pyroxenite-peridotite-gabbro-norite-anorthosite ore-bearing intrusions form two belts in the eastern part of the Baltic (Fennoscandian) Shield (Fig. 1). The Northern Belt includes such intrusions as Mt. General'skaya, Fedorovo-Pansky, Monchegorsk Layered Complex and Imandra lopolith, which are described in the present paper. The Southern Belt contains intrusions of the Olanga group (Kivakka, Tsipringa and Lukkulaivaara) and the Burakovsky intrusion situated in Karelia, and some Finnish intrusions (Koilismaa, Näränkäväära, Koitilainen, Penikat, Akanvaara, Kemi etc.).

The Northern Belt strikes northwestward for over 500 km and is confined to the southwestern edge of the Archaean Kola-Norwegian Block and to the northern and southern borders of the Palaeoproterozoic Pechenga-Imandra-Varzuga rift.

The Southern Belt stretches approximately from east to west over 350 km and occupies the northern margin of the Karelian craton. The intrusions of the Southern Belt formed in a similar pre-rifting geodynamic setting, and are generally confined to the margins of the intracontinental rift structures and are located at contacts with the Archaean basement.

Geology and petrology of the layered intrusions of the Kola Belt

Mt. General'skaya

This intrusion (Fig. 1) has a prominent cross-cutting contact with the Archaean gneiss complex in the west. The basal conglomerates associated with the Palaeoproterozoic Pechenga rift contain abundant gabbro-norite pebbles in the Luostari area. The intrusion is cut by quartz dolerite dykes, which are similar in composition to the lower volcanogenic Sariola Majarvi Suite of the Pechenga structure. The intrusion outcrops over an area of 3.5x1.5 km, but its actual size is assumed to be greater. It gently dips southwestward at an angle of 30°-35° underlying the conglomerates of the Televi Suite and volcanic andesite-basaltic flows of the Majarvi Suite, both of which constitute the lower part of the Pechenga section. The whole rock Rb-Sr age of the volcanic sequence is 2324±28 Ma (Bayanova et al., 1999). The northeast-trending intrusion (10°-20°) has a keel-like shape and a well defined internal structure. The eastern and western contacts dip towards each other at angles of 60-65° and 30-50° respectively. The intrusion is dissected by younger faults of northeastern and northwestern strike; one of the faults separates the intrusion into two large blocks. The total thickness of the intrusive sequence increases from 200 m to 1700 m from northeast to southwest. The section is dominated by gabbro-norite, olivine gabbro-norite, gabbro, anorthosite or leucogabbro, norite, bronzitite, while serpentinized peridotite is minor (Hanski et al., 1990; Bayanova et al., 1999; Skuf'in & Bayanova, 2006 etc.). Low sulphide disseminated PGE mineralization is now being explored by a mining company.

Geological data and two zircon U-Pb ages (2496±10 Ma and 2446±10 Ma, see below) for the rocks of the intrusion show that anorthositic injections took place later than gabbro-noritic ones (Bayanova et al., 1999).

The Fedorovo-Pansky Complex

The Fedorovo-Pansky Layered Complex (Fig.1) outcrops over an area of more than 400 km². It strikes northwestwards for more than 60 km and dips southwestwards at an angle of 30-35⁰. The total rock sequence is about 3-4 km thick. Tectonic dislocations divide the complex into several blocks. The major blocks from west to east (Fig. 2) are known as: the Fedorov, the Lastjavr, the Western Pansky and the Eastern Pansky (Mitrofanov et al., 2005). The Fedorovo-Pansky complex is bordered by the Archaean Keivy terrane and the Palaeoproterozoic Imandra-Varzuga rift. The rocks of the complex outcrop close to the Archaean gneisses only in the northwestern extremities, but their contacts cannot be established due to poor exposure. In the north the complex borders with the alkaline granites of the White Tundra intrusion. The alkaline granites were recently proved to be Archaean with a U-Pb zircon age of 2654±15 Ma (Bayanova, 2004; Zozulya et al., 2005). The contact of the Western Pansky Block with the Imandra-Varzuga volcano-sedimentary sequence is mostly covered by Quaternary deposits. Drilling and excavations in the south of Mt. Kamennik showed that the contact is tectonic since the intrusive and overlying Palaeoproterozoic volcano-sedimentary rocks are sheared and metamorphosed.

The Fedorovo-Pansky Complex comprises predominantly gabbro-norites with varying proportions of mafic minerals and different structural features (Fig. 3). From bottom up, the composite layered sequence is as follows (Fig. 3):

- Marginal Zone (50-100 m) of plagioclase-amphibole schists with relicts of massive fine-grained norite and gabbro-norite, which are referred to as chilled margin rocks;
- Taxitic Zone (30-300 m) that contains ore-bearing gabbro-noritic matrix (2485 Ma, see below) and early xenoliths of plagioclase-bearing pyroxenite and norite (2526-2516 Ma, see below). Syngenetic and magmatic ores are represented by Cu and Ni sulphides with Pt, Pd and Au, and Pt and Pd sulphides, bismutho-tellurides and arsenides;
- Norite Zone (50-200 m) with cumulus interlayers of harzburgite and plagioclase-bearing pyroxenite that includes an intergranular injection Cu-Ni-PGE mineralization in the lower part. The rocks of the zone are enriched in chromium (up to 1000 ppm) and contain chromite that is also typical of the rocks of the Penikat and Kemi intrusions (Finland) derived from the earliest magma portion (Iljina & Hanski, 2005). Basal Cu-Ni-PGE deposits of the Fedorov Block have been explored and prepared for licensing (Schissel et al., 2002; Mitrofanov et al., 2005).
- Main Gabbro-norite Zone (~ 1000 m) that is a thickly layered “stratified” rock series (Fig. 3) with a 40-80 m thinly layered lower horizon (LLH) at the upper part. The LLH consists of contrasting alteration of gabbro-norite, norite, pyroxenite and interlayers of leucocratic gabbro and anorthosite. The LLH contains a reef-type PGE deposit poor in base-metal sulphides. The deposit is now being extensively explored (Mitrofanov et al., 2005). According to the field investigations (Latypov & Chistyakova, 2000) the LLH anorthositic layers have been intruded later as shown by

cutting injection contacts. This is confirmed by a zircon U-Pb age for the anorthosite of 2470 ± 9 Ma (see below).

- Upper Layered Horizon (ULH) between the Lower and Upper Gabbro Zones. The ULH consists of olivine-bearing troctolite, norite, gabbronorite and anorthosite (Fig. 3). It comprises several layers of rich PGE (Pd >> Pt) ore poor in base-metal sulphides (Mitrofanov et al., 2005). The U-Pb age on zircon and baddeleyite of the ULH rocks of 2447 ± 12 Ma (see below) is the youngest among those obtained for the rocks of the Fedorovo-Pansky Complex.

The Monchepluton, intrusions of the Main Ridge (Monchetundra and Chunutundra) and adjacent intrusions - Monchegorsk Layered Complex

The Monchegorsk Layered Complex (Fig. 4) has long been the subject of detailed investigation due to the exploitation of rich Cu-Ni ores of the Monchepluton (Papunen and Gorbunov, 1985; Chashchin and et al., 2001; Smolkin et al., 2004). The Complex is located at a triple junction (Fig. 1) where weakly metamorphosed Early Proterozoic riftogene rocks and deep-seated Archaean rocks metamorphosed at granulite to amphibolite facies become contiguous at the modern erosion level. The Monchepluton is an S-shaped body with an area of ca. 65 km². It consists of two parts which represent independent magma chambers.

The northwestern and central parts of the Monchepluton (NKT: Mts. Nittis, Kumuzh'ya and Travyanaya and Mt. Sopcha) are mainly composed of non-metamorphosed ultramafic rocks, which from bottom up are represented by a 10-100 meter thick basal zone of quartz-bearing norite and gabbronorite, harzburgite (100-200 m), alternating harzburgite and orthopyroxenite (250-400 m), orthopyroxenite (300-700 m) with chromitite lenses (Mt. Kumuzh'ya) and 1-5 meter thick Cu-Ni-bearing dunite-harzburgite layers (Mt. Sopcha, "330 horizon"). The total thickness of the NKT intrusion expands southwards from 200 to 1000 m and culminates at Mt. Sopcha (1600 m).

The southeastern part of the Monchepluton (NPV: Mts. Nyud, Poaz and Vurechuaivench) consists mainly of 100-600 meter thick mafic rocks: basal quartz-bearing gabbronorite and norite (up to 50 m), melanocratic norite with lenses of olivine-bearing harzburgite and norite, ore-bearing "critical horizon" with xenoliths, olivine-free mesocratic and leucocratic norite and gabbronorite, gabbronorite, leucogabbro, anorthosite with PGE mineralization (Mt. Vurechuaivench).

Both parts of the Monchepluton (NKT and NPV chambers) have a trough-like shape with a near-horizontal floor and flanks dipping southwestwards at an angle of 20-40°. The Complex is underlain by the Archaean gneiss and migmatite, and overlain by the Sumi rocks of the Imandra-Varzuga rift (near Mt. Vurechuaivench). The intrusive rocks of the Monchepluton are cut by veins of basic to intermediate pegmatites and diorite, and by dolerite and lamprophyre dykes.

The syngenetic disseminated Cu-Ni ore occurs in layers and is usually spatially confined to the layers of olivine-bearing rocks. The ore location is controlled by the primary structural elements of the intrusions. It also may be found in the upper and basal parts of the intrusions. The mineralization is related to the coarse-grained pegmatoid rocks. Occurrences of syngenetic and nest-disseminated ore with bedded, lens-shaped and stock-like forms are locally confined to the

parts of the intrusions where fine-grained and irregular-grained rocks, pegmatoids and rocks allogenic to the intrusion (“critical horizon”) are widely developed. The distribution of the two last-mentioned rock varieties may in some cases serve to reveal ore-controlling zones. Exploitable Cu-Ni-PGE deposits of veined epigenetic ores in the Monchepluton are confined to the systems of steeply dipping shear fractures trending north-northeast and dipping south-southeast, which trace the primary structural elements of the intrusion (geometry of intrusive blocks, primary jointing etc.). The main ore-controlling elements in the occurrences of epigenetic stringer-disseminated ores are the zones of tectonic dislocations marked by schistose and blastomylonitized rocks. Most favourable for the concentration of injected stringer-disseminated ores are the places where the tectonic zones pass along the bend of the contact between rocks sharply different in physico-mechanical properties, e.g. between ultramafic rocks and Archaean granite-gneiss. The epigenetic sulphide Ni-Cu ores of the complex tend to occur in bodies with a mainly northeast strike and southwest plunge.

The rocks of the Monchepluton were dated earlier by the U-Pb methods on zircons and baddeleyite at the Geological Institute KSC RAS (Bayanova, 2004) and at the Royal Ontario Museum laboratory in Canada (Amelin et al., 1995) with a good convergence of results (see below). These ages fall in the range of 2507-2490 million years and favour the correlation of the Monchepluton mafic-ultramafic layered series with the mafic layered series of the second intrusive phase of the Fedorovo-Pansky massif. In both intrusions, the main phase melts have produced Cu-Ni-PGE economic mineralization where base metals constitute the major part, but the portion of platinum in the PGE disseminated occurrences is at least 20%. The ore bodies within the ultramafic rocks of the Monchepluton (Papunen & Gorbunov et al., 1985) are considerably richer than those of the Fedorov block deposit (Schissel et al., 2002). However, the deposits of the Monchegorsk region have already been mined out, while the Fedorovo-Pansky Complex is now being carefully investigated for future development.

Great areas of the Monchegorsk ore region are occupied by amphibolite-facies high-pressure garnet-bearing gabbro-anorthosite and anorthosite with numerous conformable and cutting veins of leucogabbro and pegmatoid rocks. These are the intrusions of the Main Ridge and Lapland-Kolvitsa granulite belts (Pyrchin etc.) located within strongly metamorphosed country rocks.

The rocks of the intrusions are insufficiently studied by modern geological and petrological methods, but have been investigated by mining companies because of the presence of high PGE and V-Ti concentrations. The Monchetundra intrusion is separated from the Monchepluton by a thick (a few hundreds of meters) blastomylonite zone with a garnet-amphibole mineral association (Smolkin et al., 2004). Regional shear zones cut and transform the primary monolith-like shape of the intrusion composed of roughly layered leucocratic mafic rocks. This results in a lens-like morphology of the intrusions.

Available U-Pb isotope ages of these anorthosites fall in a wide time interval (Bayanova, 2004; Mitrofanov & Nerovich, 2003). The zircons derived from magmatic plagioclase yield an age

varying from 2500 to 2460 Ma for different intrusions. A few generations of metamorphic zircons yield an age of multistage metamorphism that took place 2420, 1940 and 1900 Ma (Mitrofanov & Nerovich, 2003).

The Imandra lopolith

The Imandra lopolith is represented by a chain of six isolated sheet-like bodies with a thickness of up to 3 km that extend for tens of kilometers. These are Mt. Devich'ja, Yagel'naya and Bol'shaya Varaka, Monchepoluostrov, Prikhibin'je and Umbarechka.

The Imandra lopolith cuts the Archaean biotite-amphibole gneiss and amphibolite basement rocks, as well as the volcano-sedimentary rocks of the Palaeoproterozoic Strelna Series of the Imandra-Varzuga palaeorift. The northern part of the lopolith occurs among the Strelna Series rocks, nearly conformably to the horizon of the Seidorechka felsic volcanic rocks. The southern part is sandwiched between the Palaeoarchaean amphibolites and felsic volcanic rocks of the Proterozoic Tominga Series. The intrusive relationships between the lopolith and country rocks are manifested in the recrystallization and formation of hornfels at the exocontact zone, and in the presence of schist and effusive xenoliths in the lopolith rocks. The Imandra lopolith is cut by dykes of varying composition.

The composite section of the lopolith includes the lower, the main, the upper and the roof zones. The lower layered zone has a thickness of 100 to 200 m. At the contact with the country rocks, there is a 5-7-m thick layer of fine-grained mesocratic metamorphosed gabbroids that evidently represent the marginal part of the lopolith. Above this there are alternating plagioclase-bearing orthopyroxenite and poikilitic melanocratic norite and gabbro-norite with a thickness of 45-60 m. The rocks contain up to 1-m thick seams composed of mesocratic gabbro associated with chromite mineralization. The upper part of the lower layered zone (45-55 m) consists of gabbro-norite.

The main zone is dominated by mesocratic gabbro-norite with a thickness of about 2 km. In the lower part of the main zone there are thin (0.1-0.4 m) seams of olivine-bearing gabbro-norite, and in the upper – seams of leucocratic gabbro-norite of varying thickness.

The upper layered zone (ca. 300 m) displays rhythmic layering resulted from the alternation of mesocratic and leucocratic gabbro.

The roof zone ranges from 150 to 500 m in thickness and is composed of mesocratic quartz gabbro and gabbro-diorite. Disseminated Ti-magnetite mineralization is confined to the bottom of the zone. The upper contact of the lopolith has not been established (Bayanova & Balashov, 1995; Bayanova et al., 2001).

In composition (pyroxenite, norite, gabbro, quartz gabbro-diorite, and anorthosite) the V-Ti-Fe mineralization of the Imandra lopolith is similar to that of the Porttivaara block of the Koillismaa layered igneous complex in Finland (Iljina & Hanski, 2005).

Analytical methods

U-Pb (TIMS) method

Following the method proposed by T. Krogh (1973), the samples were dissolved in strong (48%) hydrofluoric acid at a temperature of 205-210°C over one to ten days. In order to dissolve fluorides the samples were reacted with 3.1N HCl at a temperature of 130°C during for 8-10 hours. To determine the isotope composition of lead and concentrations of lead and uranium, the sample was divided into two aliquots in 3.1 N HCl, and a mixed $^{208}\text{Pb} + ^{235}\text{U}$ tracer was added. Pb and U were separated on an AG 1 x 8, 200-400 mesh anion exchanger in Teflon columns. The laboratory blank for the whole analysis was less than 0.1-0.08 ng for Pb and 0.01-0.04 ng for U. All isotopic determinations for zircon and baddeleyite were made on Finnigan MAT-262 and MI 1201-T mass spectrometers and the Pb isotopic composition was analysed on a secondary-ion multiplier on a Finnigan MAT-262 in ion counting mode. The measurements of the Pb isotopic composition are accurate to 0.025% (Finnigan MAT-262) and 0.15% (MI 1201-T) when calibrated against NBS SRM-981 and SRM-982 standards, respectively. The U and Pb concentrations were measured in single-filament mode with the addition of H_3PO_4 and silica gel using the method (Scharer & Gower, 1988; Scharer et al., 1996). Pb and U concentrations were measured within the temperature ranges of 1350-1450 and 1450-1550°C, respectively. All of the isotopic ratios were corrected for mass discrimination during the static processing of replicate analyses of the SRM-981 and SRM-982 standards ($0.12 \pm 0.04\%$ for the Finnigan MAT-262 and $0.17 \pm 0.05\%$ per a.m.u.). The errors in the U-Pb ratios were calculated during the statistical treatment of replicate analyses of the IGF-87 standard and were assumed equal to 0.5% for Finnigan MAT-262 and 0.7% for MI 1201-T. If the actual analytical errors were higher, they are reported in the table of isotopic data. Isochrons and sample points were calculated by the programs (Ludwig, 1991, 1999). The age values were calculated with the conventional decay constants for U (Steiger & Jager, 1977), all errors are reported for a 2 sigma level. Corrections for common Pb were made according to (Stacey & Kramers, 1975). Corrections were also made for the composition of Pb separated from syngenetic plagioclase or microcline if the admixture of common Pb was >10% of the overall Pb concentration and the $^{206}\text{Pb}/^{204}\text{Pb}$ ratios were less than 1000.

^{143}Sm - ^{144}Nd method

In order to define concentrations of samarium and neodymium, the sample was mixed with a compound tracer $^{149}\text{Sm}/^{150}\text{Nd}$ prior to dissolution. It was then diluted with a mixture of HF + HNO_3 (or + HClO_4) in teflon sample-bottles at a temperature of 100° C until complete dissolution. Further extraction of Sm and Nd was carried out using standard procedures with two-stage ion-exchange and extraction-chromatographic separation using ion-exchange tar «Dowex» 50 x 8 in chromatographic columns employing 2.3 N and 4.5 N HCl as an eluent. The separated Sm and Nd fractions were transferred into nitrate form, whereupon the samples (preparations) were ready for mass-spectrometric analysis. Measurements of Nd isotope composition and Sm and Nd concentrations by isotope dilution were performed using a multicollector mass-spectrometer in a Finnigan MAT 262 (RPQ) in a static mode using Re+Re and Ta+Re filament. The measured

reproducibility for ten parallel analysis of Nd isotope composition for the standard La Jolla = 0.511833 ± 6 was less than 0.0024 % (2σ). The same reproducibility was obtained from eleven parallel analyses of the Japanese standard: Ji Nd1 = 0.512078 ± 5 . The error in $^{147}\text{Sm}/^{144}\text{Nd}$ ratios of 0.2% (2σ), the average of seven measures, was accepted for statistic calculations of Sm and Nd concentrations using the BCR standard. The blanks for laboratory contamination for Nd and Sm are 0.3 and 0.06 ng, respectively. Isochron parameters were developed from programs of Ludwig (1991, 1999). The reproducibility of measurements was $\pm 0.2\%$ (2σ) for Sm/Nd ratios and $\pm 0.003\%$ (2σ) for Nd isotope analyses. All $^{147}\text{Sm}/^{144}\text{Nd}$ and $^{143}\text{Nd}/^{144}\text{Nd}$ ratios were normalized to $^{146}\text{Nd}/^{144}\text{Nd} = 0.7219$ and adjusted to $^{143}\text{Nd}/^{144}\text{Nd} = 0.511860$ using the La Jolla Nd standard. The ϵ_{Nd} (T) values and model T_{DM} ages were calculated using the currently accepted parameters of CHUR (Jacobsen & Wasserburg, 1984): $^{143}\text{Nd}/^{144}\text{Nd} = 0.512638$ and $^{147}\text{Sm}/^{144}\text{Nd} = 0.1967$ and DM (Goldstein & Jacobsen, 1988): $^{143}\text{Nd}/^{144}\text{Nd} = 0.513151$ and $^{147}\text{Sm}/^{144}\text{Nd} = 0.2136$.

⁸⁷Rb-⁸⁶Sr method

The samples and minerals were all treated with double distilled acids (HCl, HF, and HNO₃) and H₂O distillate. A sample of 20 to 100 mg (depending on Rb and Sr contents) was dissolved with 4 ml of mixed HF and HNO₃ (5:1) in corked teflon sample-bottles and left at a temperature of about 200° C for a day. The solution was then divided into three aliquots in order to determine Rb and Sr isotope compositions and concentrations. These were measured by isotope dilution using separate ⁸⁵Rb and ⁸⁴Sr tracers. Rb and Sr extraction was performed by eluent chromatography with «Dowex» tar 50x8 (200-400 mesh). 1.5N and 2.3N HCl served as an eluent. Tar volumes in the columns were $\sim 7 \text{ sm}^3$ and $\sim 4 \text{ sm}^3$. The separated Rb and Sr fractions were evaporated until dryness followed by treatment with a few drops of HNO₃. Sr isotope compositions and Rb and Sr contents were measured by a MI-1201-T (Ukraine) mass-spectrometer in the two-ribbon mode using Re filaments. The samples prepared were deposited on the ribbons in the form of nitrate. Sr isotope composition in all the measured samples was normalized to a value of 0.710235 recommended by NBS SRM-987. Errors on Sr isotope analysis (confidence interval of 95%) do not exceed 0.04%, and those of Rb-Sr ratio determination are of 1.5%. Blank laboratory contamination for Rb is 2.5 ng and for Sr 1.2 ng. The adopted Rb decay constant of Steiger and Jager (1977) was used for age calculations.

Results of age determinations for the layered intrusions of the Kola Belt

Mt. General'skaya

Samples for U-Pb dating were taken both from outcrops and from drill cores of the Mt. Generalskaya intrusion. Two samples (SA-416/1 and SA-416/2) of 150 kg were collected at different times from the upper gabbro-norite zone at a road-side outcrop at Pechenga – Luostari which runs across the northern flank of the intrusion. The samples are taken from the lower gabbro-norite zone and consist of medium-grained, massive gabbro-norite, and contain relict ortho- and clinopyroxenes, pseudomorphs and aggregates of hornblende and actinolite replacing pyroxene, large laths of saussuritized plagioclase and accessory Ti-magnetite. Sulphide and apatite are minor.

An 8 mg grain concentrate separated from the gabbro-norite sample (SA 416/1) yielded two morphological types of zircons. In addition, baddeleyite was found for the first time in the layered intrusions of the Kola Peninsula (Bayanova et al., 1999).

Sample S-3464 of 200 kg was collected from a borehole at a depth between 600 and 800 m which penetrated the lower part of the layered unit. The sample is composed of slightly amphibolized meso- and leucocratic gabbro-norite and of less melanocratic gabbro-norite. The rocks are medium-grained, and contain bronzite, augite (rare pigeonite), andesine-labradorite, hornblende, actinolite, chlorite, biotite, apatite and Ti-magnetite.

The U-Pb age obtained on magmatic zircon (Fig. 5) from the gabbro-norite is 2496 ± 10 Ma, MSWD=1.5; the lower intersection is at zero, indicative of modern lead loss. The zircons from the outcrop sample are more discordant than those from the drill core (Fig. 6A and Table 1).

Magmatic zircons from Sample SA-416/1 (type 1) were sent to the Royal Ontario Museum laboratory in Canada to be reanalysed in order to compare the results. The U-Pb age obtained with the use of the abrasion technique was 2505 ± 1.5 Ma (Amelin et al., 1995).

Zircons xenocrysts are stubby columnar and brown-pinkish crystals up to 100 μm in size. They show well defined broad zones. They contain more uranium than magmatic zircons. The $^{207}\text{Pb}/^{206}\text{Pb}$ ages on the zircons xenocrysts ranges from 2660-2606 Ma.

Sample SA-443 of 80 kg is taken from the middle part of the layered unit represented by thin (50 m) micro-rhythms of anorthosite (or leucogabbro). The latest phase has a trough-like shape and consists of rhythmically alternating layers of melanocratic and mesocratic gabbro-norites which are 3-20 cm and 30-80 cm thick respectively. The base of the micro-rhythmic zone contains an up to 4.0 m thick lens-like anorthosite layer. The anorthosite is replaced by slightly amphibolized leucocratic gabbro-norite along strike. The layer has sharp boundaries and includes pseudobrecciated fragments of mesocratic gabbro-norite at its base. Thin anorthosite apophyses penetrate the underlain gabbro-norite and indicate later crystallization of the anorthositic melt.

The anorthosite is dominated by tabular plagioclase, with minor amphibole, relict intercumulus pyroxene, chlorite, biotite and apatite. Co-existing ortho- and clinopyroxenes, pigeonite, and plagioclase of labradorite composition are common.

Two zircon types (magmatic and xenocrysts) were separated from the anorthosite sample. Magmatic zircons are transparent bipyramidal-prismatic grains up to 150 μm in size. The terminations of bipyramids have simple faces. In immersion view, all the grains display narrow zoning towards the crystal edge.

Three zircon populations differing mainly in size were separated for U-Pb dating. The U-Pb age on magmatic zircons from the anorthosite is 2446 ± 10 Ma, MSWD=0.003 (Fig. 6B and Table 1).

Fedorovo-Pansky Complex

Several large samples were selected for the U-Pb dating of the Fedorovo-Pansky Complex.

A 60-kg sample of medium- and coarse-grained gabbronorite was collected from the Lower Layered Horizon in the Eastern Kievev area. The separated zircons are transparent with a vitreous luster. All the grains were divided into three types: Pan-1 – regular bipyramidal-prismatic crystals of up to 120 μm ; Pan-2 – fragments of prismatic crystals; Pan-3 – pyramidal apices of crystals of 80-100 μm . In immersion view, all the zircons display a simple structure with fine zoning and cross jointing (Fig. 5).

The discordia plotted on three points yields the upper intersection with the concordia and the U-Pb age at 2491 ± 1.5 Ma, MSWD=0.05. The lower intersection of the discordia with the concordia is at zero and reflects modern lead losses (Fig. 7A and Table 2). The same zircon sample was analysed in the Royal Ontario Museum laboratory (Canada); the obtained U-Pb zircon age is 2501.5 ± 1.7 Ma (Amelin et al., 1995) that is somewhat older than ours. The age obtained is interpreted as the time of crystallization of the main gabbronorite phase rock (Mitrofanov et al., 1997; Mitrofanov & Bayanova, 1999).

Sm-Nd dating on ortho- and clinopyroxene, plagioclase and whole rock minerals extracted from the same gabbronorite gave an age of 2487 ± 51 Ma, MSWD=1.5 (Balashov et al., 1993).

Three zircon populations of prismatic habit and light-yellow colour were separated from PGE-bearing gabbro-pegmatite (LLH). The zircons from Sample P-8 are stubby prismatic crystals with sharp outlines, about 100 μm in size. The crystals show cross cracks and apparent zoning in immersion view. The zircons from Samples D-15 and D-18 are multi-zoned pinkish fragments of prismatic crystals with adamantine luster and they are 80 and 100 μm in size. The U-Pb zircon age of 2470 ± 9 Ma, MSWD=0.37 (Fig. 7A and Table 2) was obtained from three points: one concordant and two lying in the upper part of the isochron. The lower intersection of the discordia with the concordia (about 300 Ma) indicates lead loss associated with the Palaeozoic tectonic activation of the eastern Baltic Shield and the development of the giant Khibina and Lovozero intrusions of nepheline syenites (Kramm et al., 1993). Zircons from the gabbro-pegmatite are found to have higher U and Pb concentrations than those from the gabbronorite.

Three zircon and two baddeleyite populations were separated from a sample collected from the Upper Layered Horizon in the Southern Suleypahk area. All the zircons from anorthosite are prismatic, light-pink-coloured with vitreous luster. In immersion view, they are zoned and fractured. A population of bipyramidal-prismatic zircons (Pb-1) is made up of elongate (3:1) crystals. Sample Pb-2 contains zircons of round-ellipsoidal habit; and Sample Pb-3 contains transparent flattened crystal fragments of up to 0.75 μm in size.

The separated baddeleyite crystals (first recorded in the anorthosite) were subdivided into two varieties, deep-brown and brown. All the grains are fragments of transparent baddeleyite crystals of 50 μm in size without selvages and inclusions.

A U-Pb isochron plotted from three zircons and two baddeleyites intersects the concordia with an age of 2447 ± 12 Ma, MSWD=2.7 (Fig. 7B and Table 2). The lower intersection of the discordia with the concordia records recent lead loss. The position of baddeleyite points is near-concordant,

while zircon points (sample P6-1) are above the concordia due to uranium loss. This age (2447 ± 12 Ma) is considered to constrain the origin of late-phase anorthosite, because, as shown by Heaman and LeCheminant (1993), baddeleyite is commonly generated in residual melts.

The U-Pb zircon age of the early barren orthopyroxenite from the Fedorov Block, 2526 ± 6 Ma, is believed to be the time of emplacement (Fig. 7C and Table 2). The U-Pb age of 2516 ± 7 Ma (Fig. 7D and Table 2), obtained from zircon from barren olivine gabbro, is interpreted as the time of crystallization. The last Cu-Ni-PGE-bearing taxitic gabbro from the Fedorov Block (Fig. 7E and Table 2) yielded a U-Pb zircon age of 2485 ± 9 Ma (Nitkina, 2006).

Monchegorsk Layered Complex

Ten samples of 50-120 kg were collected for U-Pb dating. Accessory baddeleyite and zircon were better preserved in drill core samples, than in outcrops (Fig. 5).

The oldest rocks studied are pegmatites of gabbro-norite composition, which are associated with the ore-bearing sulphide veins from the basal zone of Mt. Travyanaya and the “critical horizon” (Mt. Hyud, Terassa deposit). Two baddeleyite and three zircon populations were examined from these rocks. All the crystals were unaltered. Baddeleyite grains are up to $80\ \mu\text{m}$ long and light-brown in colour. Zircons are prismatic and isometric, up to $150\ \mu\text{m}$ in size, and feature narrow igneous zoning and various hues of brown. U and Pb concentrations are high, which is typical of pegmatite. A U-Pb age obtained on the five zircon and baddeleyite populations is 2500 ± 5 Ma, $\text{MSWD}=1.7$; the lower intersection of the discordia and the concordia is at 349 ± 81 Ma, indicating Palaeozoic lead losses (Fig. 8A and Table 3). This age is comparable with that of 2493 ± 7 Ma obtained for gabbro-norite of Mt. Nyud, and with a zircon age for the norite of Mt. Travyanaya (Fig. 8B and Table 3). A U-Pb age on baddeleyite and zircon recently obtained for the coarse-grained gabbro-norite of Mt. Vurechuaivench foothills (now considered as a PGE-bearing reef) is 2497 ± 21 Ma, being very similar to that for the Fedorovo-Pansky gabbro-norite (Fig. 8C and Table 3).

To determine the age of the Sopchezero chromite deposit located within the Dunite block of the Monchepluton, cutting dyke rocks were studied. The Dunite Block is composed of rocks poor in accessory minerals. The dykes are assumed to be associated with intrusive mafic rocks of the Monchepluton and are thought to intrude the Dunite block rocks before they had cooled. Thus the age of the dykes would constrain the minimum age limit of the Dunite Block and Sopchezero deposit formation. For U-Pb dating a sample was collected from Borehole 1586 at a depth of 63-125 m, from a coarse-grained gabbro-norite dyke cutting the ultramafic rocks of the Dunite Block. Baddeleyite, two zircon populations and rutile were used for dating. Brown transparent plate-like baddeleyite grains of up to $70\text{-}80\ \mu\text{m}$ in size are well preserved. Light-pink zircons of up to $150\ \mu\text{m}$ in size have good outlines and thin zoning. The U-Pb age on zircon and baddeleyite is 2496 ± 14 Ma, $\text{MSWD}=0.011$; the lower intersection of the discordia with the concordia is at 313 ± 271 Ma (Fig. 8D and Table 3). The point for the rutile has a near concordant value of ~ 1.84 Ga that reflects the time of its formation. A similar U-Pb age (2506 ± 10 Ma) has also been obtained on zircon from a coarse-grained gabbro-norite dyke from Borehole 1518 (Fig. 8E and Table 3). The gabbro-norite

dyke cuts the ultramafic rocks of the Dunite Block, therefore the Dunite Block must be older than the Monchepluton.

Small intrusions and dykes of the Monchegorsk Layered Complex were considered by most geologists to have the same age as the Monchepluton. In order to verify these relationships, diorite of the Yarva-Varaka intrusion was studied. Three zircon types and baddeleyite were selected from a sample of quartz diorite and granophyric hypersthene diorite collected in the upper part of the Yarva-Varaka section. Stubby prismatic, pink-brown zircons of up to 150 μm in size were divided by their colour hues into three populations. In immersion view, they are multi-zoned. Baddeleyite grains and fragments are prismatic in habit, light-brown-coloured and up to 80 μm in size. A U-Pb age obtained on four points is 2496 ± 9 Ma, MSWD=0.93, the lower intersection is at zero, indicating recent lead losses (Fig. 8F and Table 3).

The Ostrovsky intrusion also belongs to the series of small mafic-ultramafic intrusions of the Monchegorsk Layered Complex. It was considered to correlate in age with the Monchepluton and was interesting as a target for Cu-Ni prospecting. A sample for U-Pb dating was taken from mafic pegmatite veins in the middle part of the upper gabbro-norite zone (Mt. Ostrovskaya). The pegmatite body is more than 1 m thick, up to 2 m long and has a complex morphology, with sinuous contacts with the coarse-grained slightly-amphibolized host pigeonite gabbro-norite. The sample is dominated by coarse-grained to pegmatoid gabbro-norites with a poikilitic texture, made up mostly of calcic plagioclase and amphibolized clinopyroxene. The 60 kg sample produced two types of baddeleyite and two types of zircon. Baddeleyite grains of type 1 are up to 80 μm in size, with a deep-brown colour and flattened and tabular structure. Larger, up to 120 μm baddeleyite grains of type 2 were found within a fringe of metamict zircon and were exposed to aeroabrasion for 15 minutes in order to remove the metamict fringe. Zircons are prismatic, up to 125 μm in size, and are subdivided into light-brown and brown varieties. Zircons show well developed joints and thin zoning in immersion view. The U-Pb isochron age on two baddeleyite and two zircon points is 2445 ± 11 Ma, MSWD=0.12, the lower intersection of the discordia with the concordia is at 500 ± 99 Ma (Fig. 8G and Table 3).

To establish age correlations between the gabbro-norite of the Monchepluton and the anorthosite of the Main Ridge intrusion, rock samples of the Monchetundra and Chunutundra intrusions were studied.

The Monchetundra intrusion has a complex structure and an overview of geological and geochronological investigations is given by Smolkin et al. (2004). It includes the upper zone comprised mainly of amphibolized gabbro-norite and gabbro-anorthosite, and the lower zone, which consists of gabbro-norite, norite and plagiopyroxenite (drilled by the deep borehole M-1).

The middle part of the upper zone, which contains a prominent horizon of slightly-altered medium-coarse-grained gabbro-norite with trachytoid texture, was sampled for U-Pb dating. The sample yielded three zircon types. Prismatic acicular crystals up to 200 μm in size and their brown fragments were divided into three types by colour. In immersion view, multi-zoning, mineral

inclusions, strong jointing, corrosion of the surface and spotted uneven grain colour are observed. The U-Pb ages (Fig. 8H-I and Table 3) on zircon from trachytoid gabbro-norite are 2505 ± 6 Ma, MSWD = 0.31 and 2501 ± 8 Ma, MSWD = 3 (Bayanova & Mitrofanov, 2005).

A sample was also taken from the rocks of the differentiated series of the Chunutundra intrusion. Zircons from medium-grained leucogabbro with trachytoid texture were divided into five types. Four types are up to 150 μm isometric fragments of brown and pink colour, while the last fraction is represented by up to 120 μm twinned pinkish-brown zircons with adamantine luster. The U-Pb isochron plotted on five points has the upper intersection with the concordia at 2467 ± 7 Ma, MSWD = 1.4 and the lower intersection is at zero (Fig. 8J and Table 3). This age is close to the age obtained on magmatic zircon from anorthosite of the Pyrrshin intrusion (Mitrofanov & Nerovich, 2003) and on zircons from later anorthositic injections of the LLH (Fedorovo-Pansky Complex).

Imandra lopolith

Several large samples were taken from the different parts of the lopolith for dating. Baddeleyite was found in a chromite horizon and underlying norite of the Bol'shaya Varaka area. A sample of up to 350 kg collected in order to study chromite mineralization, produced about 10 mg of zircon-baddeleyite concentrate. Angular and prismatic baddeleyite fragments were up to 80 μm in size; they are black and nearly opaque and have gouges on their faces (Fig. 5). The U-Pb isochron plotted on three points has the upper intersection with the concordia at 2446 ± 39 Ma, MSWD = 5.1. The lower intersection is at zero reflecting recent lead loss (Fig 9A and Table 4). The best-preserved and undisturbed U-Pb system was displayed by a zircon grain with a coordinate lying on the concordia within the determination error. The coordinate of one of the baddeleyite points is strongly discordant, while the other baddeleyite point lies above the concordia due to lead removal. The baddeleyite fraction was exposed to acid treatment in order to remove a thin white coating, following a similar technique to that applied to baddeleyite from carbonatites of the Phalaborwa intrusion in South Africa (Reischmann, 1995). The U-Pb age on zircon and baddeleyite of 2446 ± 39 Ma is the oldest for the Imandra lopolith (Table 5).

A gabbro-norite sample was taken from Borehole No 6 within the depth interval of 57.8-75.5 m in the Umbarechka Block of the intrusion. The mineral Sm-Nd isochron corresponds to an age of 2444 ± 77 Ma at $\epsilon_{\text{Nd}} = -2.0 \pm 0.6$, MSWD = 0.31. Zircons separated from the sample were analysed at the Royal Ontario Museum laboratory in Canada (Amelin et al., 1995), where a U-Pb age of 2441 ± 1.6 Ma was obtained (Table 5). The two ages determined by different systems give a good correlation and reflect crystallization time for the early rocks of the lower part of the lopolith – ca. 2.46 Ga, which we refer to the main phase intrusion.

A close U-Pb age on zircon and baddeleyite was obtained for norite of the main phase of the Umbarechka intrusion. Three zircon points and baddeleyite lie on the isochron and give an age of 2437 ± 7 Ma (Fig. 9B and Table 4). Thus, the analysed rocks of the main phase of the Imandra lopolith formed within the interval of 10 million years, i.e., from 2446 to 2437 Ma.

The second phase of the lopolith represented by leucogabbro-anorthosite and ferrogabbro with Ti-magnetite mineralization was studied in the Prikhibin'je and Bol'shaya Varaka Blocks at the upper contact of the lopolith, where one may observe a 400-m thick ferrogabbro zone with 10-m anorthosite and Ti-magnetite gabbro, and a 400-m granophyric zone. The ferrogranophyre contains up to 64% SiO₂ and up to 15% of total iron oxide, and has high contents of P₂O₅, Co and Cu. The granophyre is close to the roof rhyodacite in silica content, total iron oxide, Zr and REE (Bayanova et al., 2001).

Baddeleyite was found in thin-sections of granophyre for the first time. Baddeleyite grains are brown, up to 100 µm in size and form single crystals and growths with ilmenite in plagioclase, pyroxene and later quartz and biotite interstices. Single crystals are confined to aggregates of biotite, which is genetically related to an early magmatic phase in the granophyric rocks. In all cases the baddeleyite and ilmenite were earlier minerals than quartz and micropegmatite. This suggests that they originated during the early stages of crystallization of the intercumulus melt under conditions of low silica activity. Zircon forms individual crystals and thin fringes (≤ 10 µm) around baddeleyite, which together with titanite, develop after ilmenite and are affected by subsolidus alteration. Multi-zoned light-coloured stubby prismatic zircon crystals up to 100-200 µm in size form aggregates in the micropegmatite matrix (Bayanova, 2006).

Three baddeleyite populations of flattened transparent brown about 100 µm large crystals were separated from a 20 kg anorthosite sample. The U-Pb age obtained on these baddeleyites is 2437±11 Ma, MSWD =1.7 (Fig. 9C and Table 4).

We consider that the granophyric rocks resulted from remelting of host volcano-sedimentary rocks along with a contamination by a felsic material. A 35-meter thick layer of newly-formed hybrid ferrodiorite melt gave rise to ferrogranophyric rocks. This conclusion is supported by similar U-Pb crystallization ages obtained for the rocks from the apex of the Imandra lopolith (2437±11 Ma) and for the granophyres from its endocontact zone (2434±15 Ma) (Table 5) (Bayanova & Balashov, 1995). The roof granophyres of the Bushveld complex that originated from reworking of the host felsic rocks show similar relationships (Eules & Cawthorn, 1995).

The rocks of the lower layered part of the Imandra lopolith are cut by monzodiorite dykes, which are part of a third phase of emplacement (Bayanova et al., 2001). A 50 kg sample from a monzodiorite dyke cutting the Umbarechka Block yielded well-preserved accessory zircons. Light, pale pink and pinkish transparent zircon grains have prismatic habit are about 100 µm in size, and display a thin faint zoning; they were subdivided into four types by colour. The U and Pb contents in the accessory minerals from the dyke are higher than those in zircon and baddeleyite from the Imandra lopolith.

The U-Pb isochron plotted on four varieties of zircon from the monzodiorite dyke gives an age of 2398±21 Ma, MSWD =2.4; the lower intersection is at zero, suggesting modern lead losses (Fig. 9D and Table 4). A close U-Pb age was obtained for zircon and baddeleyite from a pegmatoid

olivine gabbro-norite dyke. Three zircon types and baddeleyite yield a U-Pb age of 2395 ± 5 Ma, MSWD = 0.63 (Fig. 9E and Table 4).

Hence, dating of different parts of the Imandra lopolith makes it possible to determine the duration of crystallization processes and define three pulses of the Imandra lopolith evolution lasting for about 50 million years. The interval agrees with the available data for the older Fedorovo-Pansky, Mt. General'skaya and Monchegorsk layered complexes.

Discussion

Specific features of the isotope investigation of the intrusions

U-Pb (TIMS), Sm-Nd, and Rb-Sr methods have been applied in this work for different purposes.

The U-Pb concordia and isochron method has been used to define the age of the rocks. The values obtained on zircons and baddeleyites from the same sample usually lie at the isochron (Fig. 7, 8 and 9), indicating a similar age of magma crystallization and subsequent transformations. Coordinates of baddeleyites are near the concordia line. The method however encounters the obstacle that mafic rocks contain very few zircon and baddeleyite grains. Samples of tens of kilograms yield only a few milligrams of these minerals.

The samples of gabbro-norite and anorthosite taken from Mt. General'skaya and from the underlying granite-gneiss contain both magmatic and zircon xenocrysts. Refractory zircon xenocrysts reflect local processes of crustal contamination and lie outside the isochron (Fig. 6) showing a $^{207}\text{Pb}/^{206}\text{Pb}$ age of more than 2.6 Ga.

In order to compare our U-Pb results (23 isochron points), some samples were sent to the Royal Ontario Museum laboratory (Canada). The data obtained there (Amelin et al., 1995) agree with ours within error (Table 5).

The Sm-Nd system is not an accurate geochronometer (~2-5%). However, the Sm-Nd isochron method can allow the establishment of crystallization times for mafic rocks on major rock-forming minerals (olivine, orthopyroxene, clinopyroxene, and plagioclase). It is especially important for dating rocks with syngenetic ore minerals. For example, this method has been used to determine for the first time the age (2482 ± 36 Ma) of the early ore body in the Fedorovo-Pansky Cu-Ni-PGE deposit of the Kola Peninsula that has economic importance (Serov et al., 2007).

For the mafic-ultramafic intrusions of the Kola belt, the Sm-Nd ages (16 points, Tables 5 and 6) overlap because of high errors, but are commonly close to the U-Pb (TIMS) data on zircon and baddeleyite. They are especially valid for the marginal fast-crystallizing rocks of the Taxitic Zone of the Fedorovo-Pansky Complex, where the early barren orthopyroxenite and gabbro have the following ages: 2521 ± 42 Ma and 2516 ± 35 Ma (Sm-Nd method) and 2526 ± 6 Ma and 2516 ± 7 Ma (U-Pb method) respectively. The ore-bearing norite of the Fedorov block yielded an age of 2482 ± 36 Ma (Sm-Nd method) and 2485 ± 9 Ma (U-Pb method) according to Nitkina (2006) and Serov et al. (2007) (Fig. 7 and 10).

It is also important to stress that the Sm-Nd method provides valuable petrological and geochemical markers: $\epsilon_{\text{Nd}}(\text{T})$ and T_{DM} . The ϵ_{Nd} shows the degree of mantle magma source depletion, while T_{DM} indicates an approximate age of the mantle protolith (Faure, 1986).

The paper presents 37 ϵ_{Nd} values (Tables 5, 6, 8, and 9) and 24 T_{DM} measurements (Tables 6, 8, and 9) calculated from new data collected, and from previously published materials (Tolstikhin et al., 1992; Amelin & Semenov, 1990; Huhma et al., 1990; Hanski et al., 2001).

The Rb-Sr whole rock and mineral isochron method is mostly valuable for dating unaltered felsic igneous rocks and metamorphic amphibolite-facies associations (Faure, 1986). In our work, 10 Rb-Sr isotope values for the rocks (Table 9) are considered to have only a petrological implication. Together with specific trace elements (Cu, Ni, Ti, V, and LREE), ϵ_{Nd} (2.5 Ga), REE (Fig. 11, Table 7), and $^4\text{He}/^3\text{He}$ (Table 10) data, the values of initial $^{87}\text{Sr}/^{86}\text{Sr}$ (I_{Sr} [2.5 Ga]) indicate an enriched mantle reservoir 2.5 billion years ago which is comparable with the modern EM-I.

The timing, pulsation, and total duration of magmatic activity

The largest and richest ore deposits of the Monchepluton and Fedorovo-Pansky Complexes have been carefully studied by geochronological methods.

The layered or differentiated series of mafic-ultramafic rocks, from troctolite to leucogabbro-anorthosite, and syngenetic Cu-Ni-PGE ores of the Monchepluton formed within the time interval of 2516 (max) to 2476 (min) Ma. Without analytical errors, the time interval is from 2507 Ma to 2493 Ma (Table 5). Some researchers (Smolkin et al., 2004) suggest that the Vurechuaivench part of the pluton, composed of gabbroids and anorthosites containing PGE deposits, is an independent magma chamber and that the age of rock and syngenetic PGE ore emplacement is 2497 ± 21 Ma.

The Fedorov Block of the Fedorovo-Pansky Complex represents an independent magma chamber, the rocks and ores of which differ significantly from those of the Western Pansky Block (Schissel et al., 2002). The 2-km thick rock sequence, from the Marginal Zone to the Lower Gabbro Zone (Figs. 2 and 3), is a layered or differentiated syngenetic series of relatively melanocratic pyroxenite-norite-gabbro-norite-gabbro dated at 2526 ± 6 and 2516 ± 7 Ma. The Taxitic Zone is penetrated by concordant and cutting Cu-Ni-PGE-bearing gabbro-norite (Fedorovo deposit) of the second pulse of magmatic injection, which is slightly younger (2485 ± 9 Ma; Table 5, Fig. 7).

The Western Pansky Block from the Main Gabbro-norite Zone, without the Lower Layered Horizon and probably without the upper part (above 3000 m, Fig. 3), can also be considered a single syngenetic series of relatively leucocratic, mainly olivine-free gabbro-norite-gabbro crystallized within the interval of $2503\text{-}2498\text{-}2491 \pm 5$ Ma (Fig. 3). In the lower part of the Block there are Norite and Marginal Zones (Fig. 2). The Marginal Zone contains poor disseminated Cu-Ni-PGE mineralization. This rock series can be correlated with certain parts of the Monchepluton and the Fedorov Block. The 40-80 m thick Lower Layered Horizon (LLH) is prominent because of its contrasting structure with predominant leucocratic anorthositic rocks. The exposed part of the horizon strikes for almost 15 km (Fig. 2) and can be traced in boreholes down to a depth of 500 m (Mitrofanov et al., 2005). By its morphology, the horizon seems to be part of a single layered series.

Nevertheless, there are anorthositic bodies that in outcrops show cutting contacts and apophyses (Latypov & Chistyakova, 2000); the cumulus plagioclase compositions in the rocks of the horizon are different from those in the surrounding rocks (Fig. 3); and the age of the PGE-bearing leucogabbro-pegmatite, which is rather precisely defined by concordant and near-concordant U-Pb data on zircon as 2470 ± 9 Ma (Fig. 7A, Table 2), is slightly younger than the ages of the surrounding rocks (e.g., 2491 ± 1.5 Ma). The LLH rocks, especially the anorthosite and the PGE mineralization, probably represent an independent magmatic pulse.

The upper part and olivine-bearing rocks of the Western Pansky Block and the anorthosite of the Upper Layered Horizon (ULH) with the Southern PGE Reef (Figs. 2, 3) have been poorly explored. They differ from the main layered units of the Block in rock, mineral and PGE mineralization composition (Mitrofanov et al., 2005). Until now, there has been obtained only one reliable U-Pb age (2447 ± 12 Ma) for the PGE-bearing anorthosite of the block, which may represent another PGE-bearing magmatic pulse.

The early magmatic activity about 2.5 Ga manifested itself in the gabbronorite of the Monchetundra (2505 ± 6 Ma and 2501 ± 8 Ma; Fig. 8H, J) and Mt. Generalskaya (2496 ± 10 Ma; Fig. 6A). The magmatic activity that resulted in the formation of anorthosite took place about 2470 and 2450 Ma. It also contributed to the layered series of the Chunutundra (2467 ± 7 Ma; Fig. 8J) and Mt. Generalskaya (2446 ± 10 Ma; Fig. 6B), Monchetundra gabbro (2453 ± 4 Ma; Table 5, Mitrofanov et al., 1993), and pegmatoid gabbronorite of the Ostrovsky intrusion (2445 ± 11 Ma; Fig. 8G).

The Imandra lopolith is the youngest large layered intrusion within the Kola Belt. It varies from the other intrusions of the Kola Belt both in its emplacement age and its metallogeny. There are five U-Pb zircon and baddeleyite ages for the rocks of the main magmatic pulse represented by norite, gabbronorite, leucogabbro-anorthosite, gabbrodiorite, and granophyre; all formed within the interval from 2445 to 2434 Ma (Table 5).

Thus, several eruptive pulses of magmatic activity have been established in the complex intrusions of the Kola Belt, including at least four pulses (or phases) in the Fedorovo-Pansky Complex: a 2526-2516 Ma barren pulse, and three ore-bearing of 2505-2485 Ma, 2470 Ma, and 2450 Ma. For similar intrusions of the Fenno-Karelian Belt, for example, Penikat intrusion in Finland, five magmatic pulses varying only in geochemistry have been distinguished from the same deep chamber (Iljina & Hanski, 2005).

A total duration for magmatic processes of over 80 million years in the Kola Belt intrusions is unexpected for many researchers.

The multi-phase magmatic duration of the Fenno-Karelian Belt intrusions was short-term and took place about 2.44 Ga years ago. However, there are only a few U-Pb precise age estimations for the Fenno-Karelian Belt intrusions (Iljina & Hanski, 2005). A joint Russian-Finnish research collaboration intended for dating the intrusions of the both belts has recently been initiated. It is expected that the research will result in updating the knowledge about the timing and duration of the Paleoproterozoic ore-forming intrusions on the Baltic Shield.

The Kola results underline that the layering of the intrusions with thinly-differentiated horizons and PGE reefs was not contemporaneous (or syngenetic) with each intrusion defining its own metallogenic trends in time and space.

Metallogenic implications

The Palaeoproterozoic magmatic activity in the eastern Baltic Shield is associated with the formation of widespread ore deposits: Cu-Ni (\pm PGE), Pt-Pd (+Rh, \pm Cu, Ni, Au), Cr, Ti-V (Richardson & Shirey, 2008; Mitrofanov & Golubev, 2008).

On the Kola Peninsula, economic Cu-Ni (+PGE) deposits are known in the Monchegorsk (~2500 Ma) and Pechenga (~1980 Ma) type intrusions. In the Monchepluton (the Monchegorsk type), syngenetic disseminated Cu-Ni (+PGE) ore bodies of magmatic origin are confined to basal parts of magmatic chambers (Papunen & Gorbunov, 1985), while massive rich redeposited ores in the veined bodies of the Monchepluton bottom as well as beyond it (offset bodies) also contain a relatively high portion of platinum among platinum-group elements. They are associated mainly with ca. 2500 Ma magnesia-rich mafic-ultramafic rocks with ϵ_{Nd} (2.5 Ga) values varying from -1 to -2. In comparison, Cu-Ni (\pm PGE) ores of the Pechenga type intrusions that are not discussed are related to the 1980 Ma gabbro-wehrlite rocks with ϵ_{Nd} (1.98 Ga) values varying from +1 to +3 (Hanski et al., 1990; Mitrofanov & Golubev, 2008). The basal ores of the Fedorovo deposit are first of all valuable for platinum-group elements (Pt, Pd, Rh), but nickel, copper and gold are also of economic importance here (Schissel et al., 2002). The ore-forming magmatic and post-magmatic processes are closely related to the Taxitic Zone gabbro-norite of 2485 \pm 9 Ma magmatic pulse.

Pt-Pd (\pm Cu, Ni, Rh, Au) reef-type deposits and ore occurrences of the Vurechuaivench Foothills (Monchepluton) and Western Pansky Block (Fedorovo-Pansky Complex) seem, in terms of genesis, to be associated with pegmatoid leucogabbro and anorthosite rocks enriched in late-stage fluids. Portions of this magma produce additional injections of ca. 2500 Ma (Vurechuaivench), ca. 2470 Ma (the Lower, Northern PGE reef), and ca. 2450 Ma (the Upper, Southern PGE reef of the Western Pansky Block and PGE-bearing mineralization of the Mt. General'skaya intrusion). These nonsimultaneous injections are quite close in terms of composition, prevalence of Pd over Pt, ore mineral composition (Mitrofanov et al., 2005), and isotope geochemistry of Sm-Nd and Rb-Sr systems. The ϵ_{Nd} values for the rocks under consideration vary from -1 to -3, which probably indicates a single long-lived magmatic hearth.

Chromium concentration (>1000 ppm) is typical geochemical feature of the lower mafic-ultramafic rocks of the layered intrusions of the Baltic Shield (Alapieti, 1982; Iljina & Hanski, 2005). The chromite mineralization is known in the basal series of the Monchepluton, Fedorovo-Pansky Complex, Imandra lopolith (Russia), Penikat and Narkaus intrusions (Finland) and in chromite deposits of the Kemi intrusion (Finland) and Dunite Block (Monchepluton, Russia). On the contrary, Fe-Ti-V mineralization of the Mustavaara intrusion (Finland) tends to most leucocratic

parts of the layered series, and to leucogabbro-anorthosite and gabbro-diorite of the Imandra lopolith (Russia) and Koillismaa Complex (Finland).

Thus, PGE-bearing deposits of the region are represented by two types: the basal and the reef-like ones. According to modern economic estimations, the basal type of deposits is nowadays more preferable for mining, even if the PGE concentration (1-3 ppm) is lower compared to the reef-type deposits (> 5ppm). Basal deposits are thicker and contain more platinum, copper and, especially, nickel. These deposits are accessible to open pit mining.

Petrological and geodynamic implications

Magmatic processes since the Palaeoproterozoic (2.53 Ga) have affected almost the whole region of the East-Scandinavian (Kola-Lapland-Karelian) province and a mature continental crust formed (2.55 Ga) in the Neoproterozoic (Gorbatshev & Bogdanova, 1993). Thick (up to 3 km) basaltic volcanites of the Sumian age (2.53-2.40 Ga) in Karelia, Kola and northeast Finland cover an area of greater than 200.000 km². In the north, magmatic analogues of these volcanic rocks are represented by two belts of layered intrusions and numerous dyke swarms (Vuollo et al., 2002, Vuollo & Huhma, 2005). This together composes a single time- and space-related megacyclic association, the East-Scandinavian Large Igneous Province (ESCLIP). All the magmatic units of the province covering a huge area show similar geological, compositional and metallogenic features.

Regional geological settings indicate anorogenic rift-like intraplate arrangements involving volcano-plutonic belts connecting different domains of the Paleoproterozoic Kola-Lapland-Karelia protocontinent. This resembles early advection extensional geodynamics of passive rifting that is typical of intraplate plume processes (Pirajno, 2007).

Geochemical and isotope-geochemical data shed light on features of deep magma source for the ESCLIP rocks. T_{DM} values (Faure, 1986) are approximately the age of the depleted mantle reservoir (DM) with slightly enriched Sm-Nd ratios. The T_{DM} values (Tables 6, 8 and 9) lie within the interval of 3.1-2.8 Ga. The ϵ_{Nd} values (Tables 5, 6, 8 and 9) vary from -1.1 to -2.4 and similar I_{Sr} values (0.703-0.704) obtained for discrete layered intrusions form a narrow range of enriched compositions. It is difficult to argue for a local crustal contamination and we suggest that the magmas producing different rocks of the ESCLIP layered intrusions were derived from a single homogenous mantle source enriched both with typically magmatic ore elements (Ni, Ti, V, and Pt) and lithophile elements including light REE (Fig. 11). To some extent, this reservoir is comparable with the modern EM-1 (Fig. 13) source (Hofmann, 1997).

$^4\text{He}/^3\text{He}$ ratio is also a reliable isotope tracer of mantle plume processes (Tolstikhin & Marty, 1998; Bayanova et al., 2006; Pirajno, 2007). Their use in studying Precambrian rocks and requires special case. Table 10 shows recent helium isotope data for the rocks and minerals of the Kola Belt intrusions. The data indicate that the $^4\text{He}/^3\text{He}$ isotope ratios of $n \times 10^{6-5}$ correspond to those of the upper mantle and differ from those of the crust ($n \times 10^8$) and lower mantle ($n \times 10^4$) (Tolstikhin & Marty, 1998). The helium isotope data tend to favour a source dominated by mantle derived magmas with only local crustal contamination.

According to the available data (Campbell, 2001; Condie, 2001; Vuollo et al., 2002; Bleeker, 2003; Ernst & Buchan, 2003; present study), the peak of the mafic-ultramafic magmatic activity of the Kola-Karelian, Superior and Wyoming provinces has been estimated at ~2.45 Ga. Figure 13 presents an attempt to demonstrate some reconstruction of the Archaean supercontinent embodying these three provinces of Europe and North America (Heaman, 1997). Insert (A) shows trends of the Kola and Fenno-Karelian Belts of 2.52-2.44 Ga layered intrusions with the intraplate nature interpreted from the results of the present study.

The ESCLIP layered intrusions are directly related to the Baltic Shield metallogeny (Mitrofanov & Golubev, 2008). The >80 Ma duration and multiphase history of the Kola Belt layered mafic intrusions (i.e., 2.53- to 2.45 Ga) has been shown here. It has also been underlined that the younger intrusions of the Fenno-Karelian Belt (Fig. 1) cluster at 2.44 Ga (Iljina & Hanski, 2005). The partially asynchronous evolution of these two belts, that are thought to be arms of a mantle plume, is now being examined in more detail as a follow-up to this study within the framework of Russian-Finnish research collaboration.

Summary and conclusions

A number of new U-Pb and Sm-Nd isotope data were obtained for various rocks of the mafic layered intrusions of the Kola Belt (Baltic Shield), including those which bear PGE, Ni-Cu and Ti-V mineralization. A surprisingly long period of multiphase magmatic activity, from 2530 to 2450 Ma (about 80 million years), resulted in the intrusion of large-scale ore-bearing intrusions of the Kola Belt. Magmatism continued until about 2400 Ma and generated wide-spread dykes and small-scale intrusions. These results contrast with the published data indicating short-term evolution interval (~2440 Ma) for similar intrusions of the Fenno-Karelian Belt (Iljina & Hanski, 2005).

The two belts of mafic layered intrusions of the Baltic Shield (the Kola and Fenno-Karelian belts), together with the surrounding volcanic rocks and dyke swarms, compose the Palaeoproterozoic East-Scandinavian Large Igneous Province (ESCLIP) with an area of more than 200.000 km². The petrological-geodynamic interpretation proposed by the present paper of the ESCLIP is a product of a vast long-lived plume is based on the enriched isotope characteristics of the magmas and also the large volume and widespread distribution of the magmas. We acknowledge that alternatives involving super-long duration of the homogenous deep-seated magma sources are possible.

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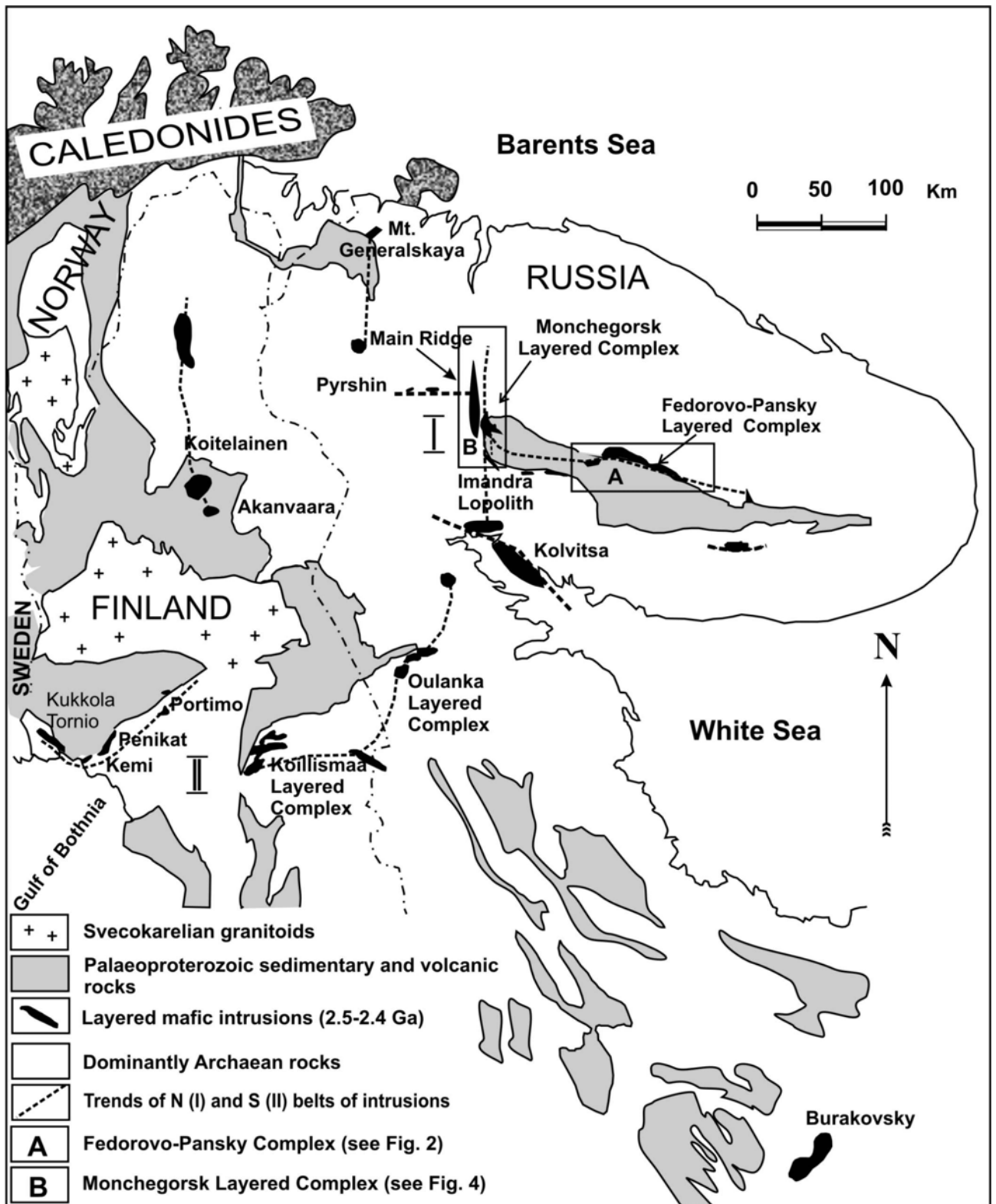


Fig.1. Generalized geological map of the northeastern part of the Baltic Shield and the location of Early Proterozoic mafic layered intrusions.

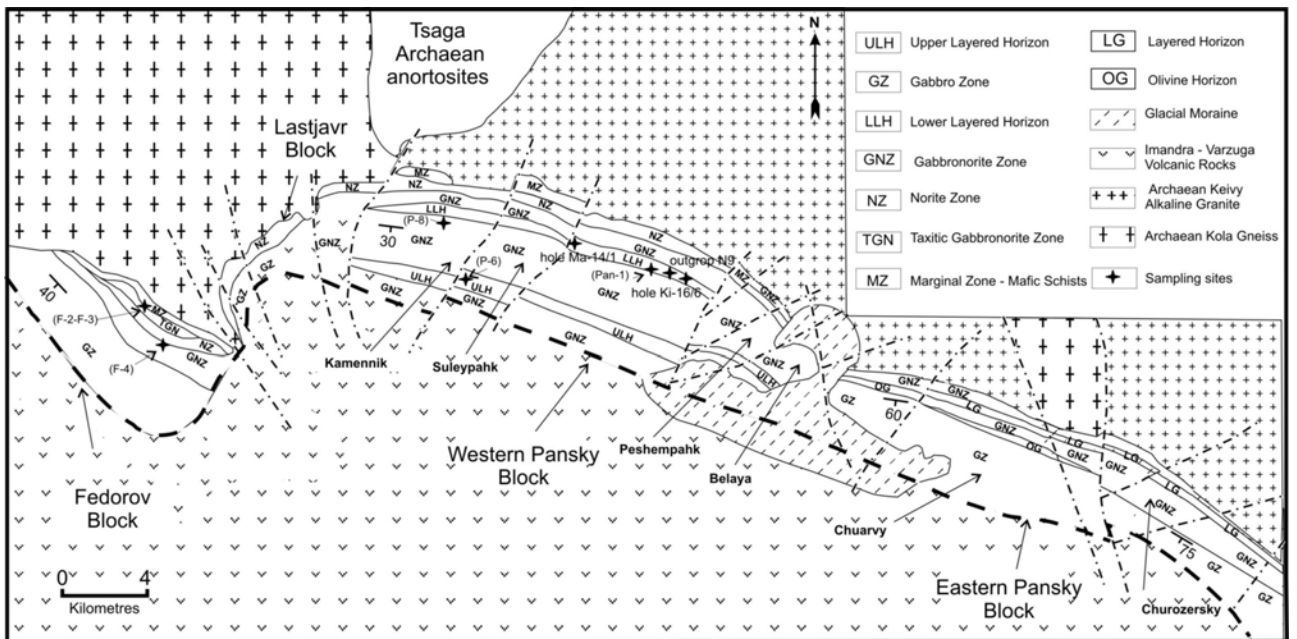


Fig. 2. General geological map of the Fedorovo-Pansky Layered Complex (from Mitrofanov et al., 2005).

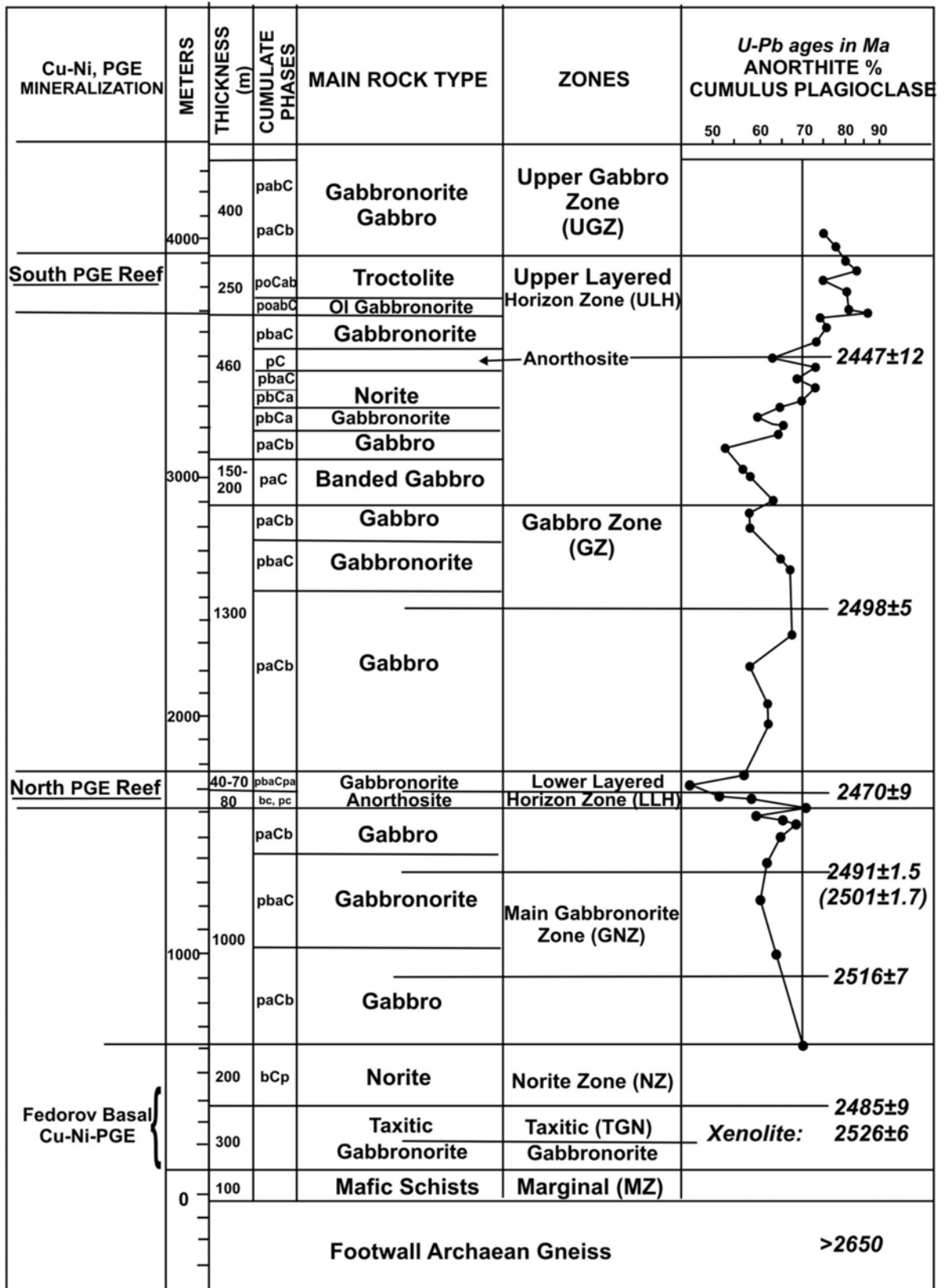


Fig. 3. Composite "stratigraphic" section of the Fedorovo-Pansky Complex with Cu-Ni and PGE mineralization (modified after Schissel et al., 2002). The cumulate mineral terminology used in this paper is that of cumulate phase minerals in small letters, in order of volume percent, preceding the capital C for cumulate, with postcumulate mineral phases following. Major mineral abbreviations are: a = augite, b = bronzite, c = chromite, o = olivine, p = plagioclase (see Table 5 for references). Modified after Schissel et al., 2002.

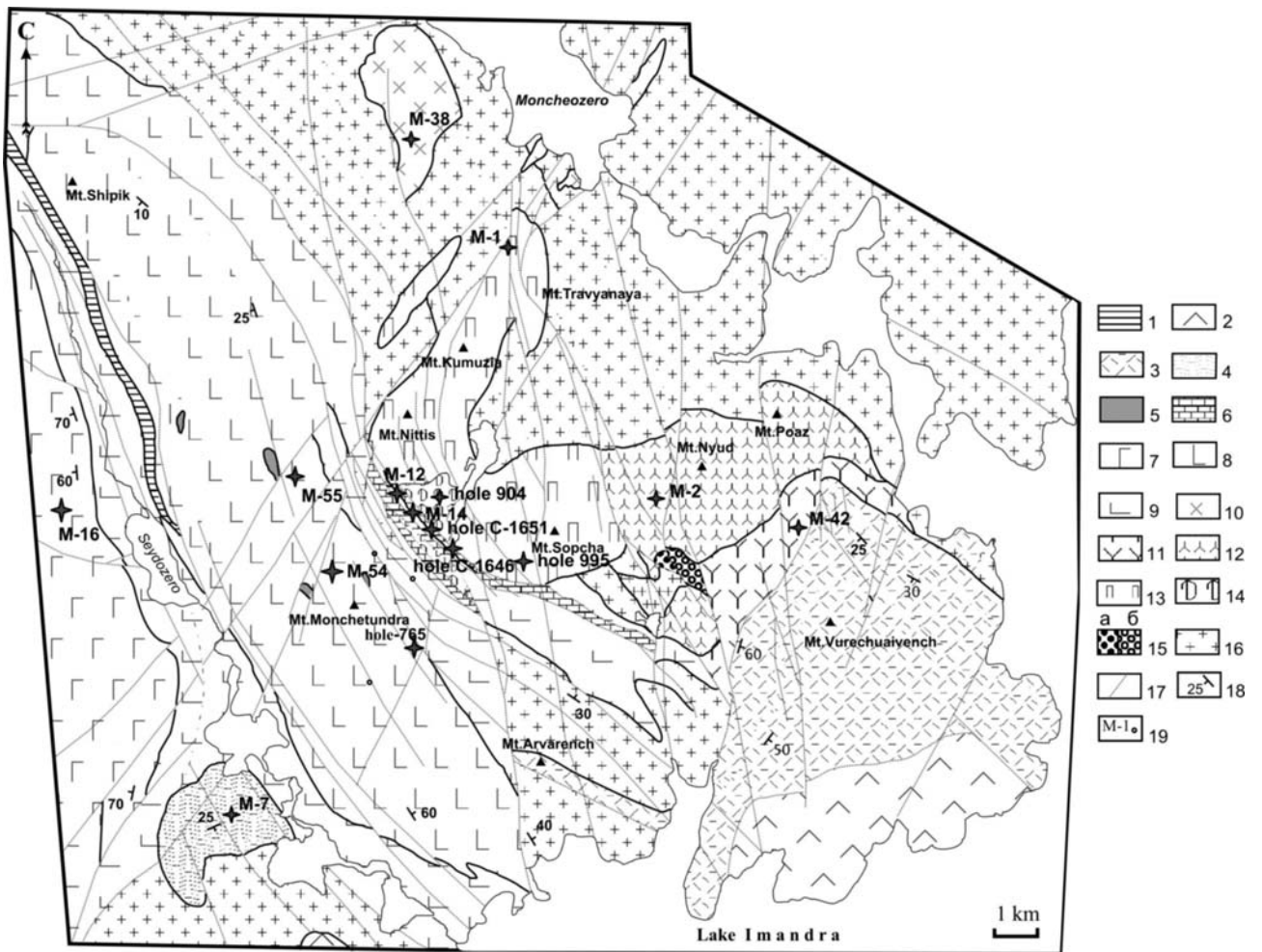
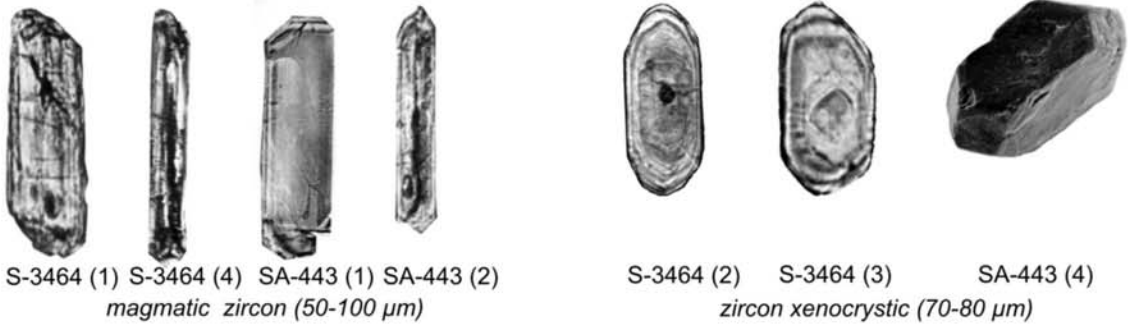


Fig. 4. Geological map of the Monchegorsk layered complex (Smolkin et al., 2004).

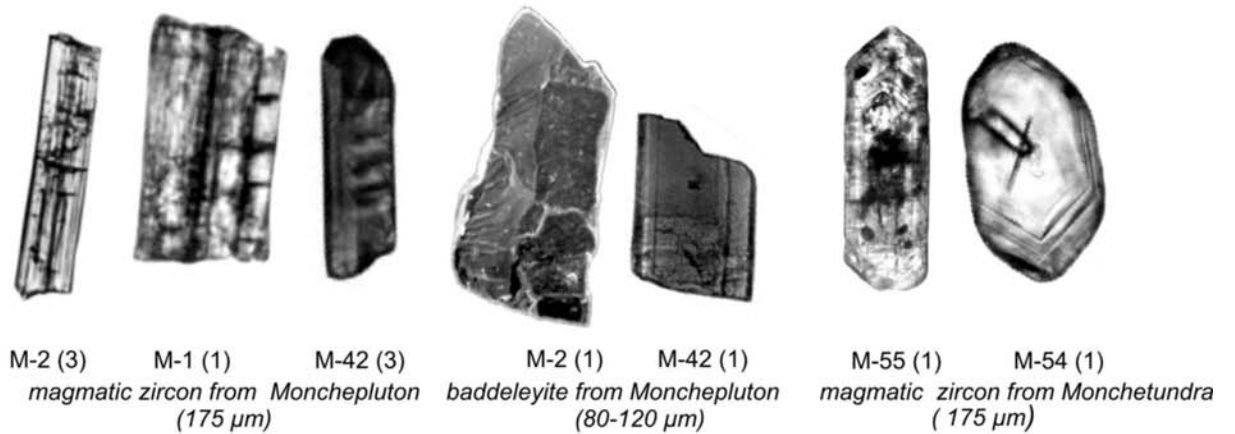
Mt. Generalskaya intrusion



Fedorovo-Pansky Complex



Monchepluton and Monchetundra



Imandra lopolith

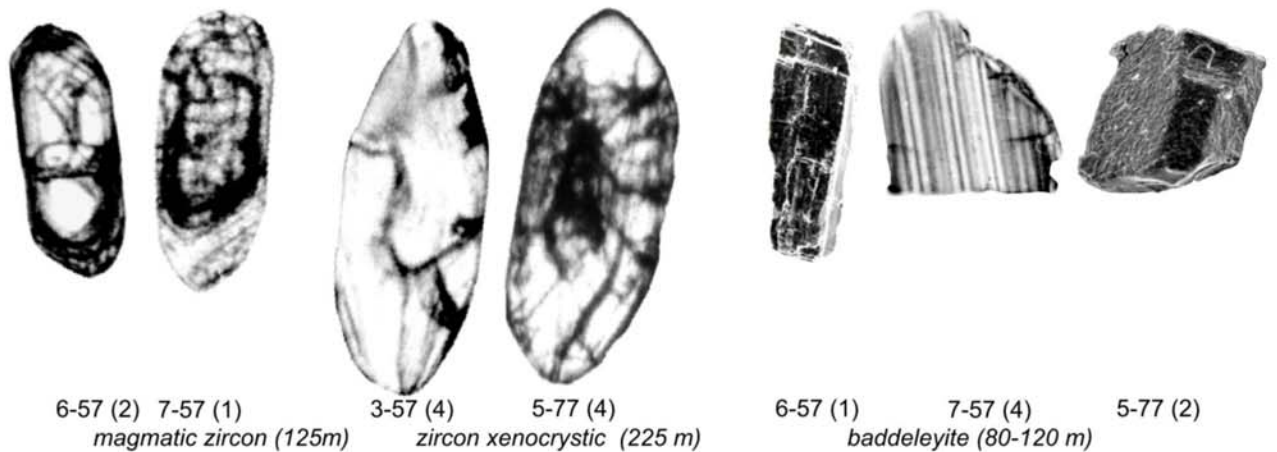


Fig. 5. Photomicrographs of analyzed zircon and baddeleyite.

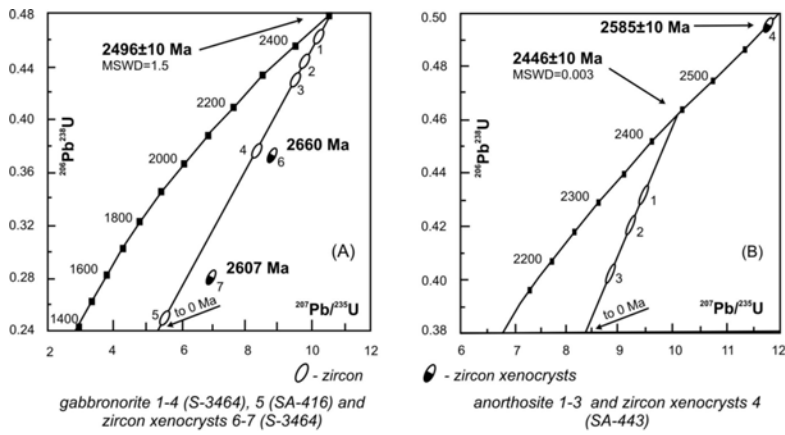


Fig. 6. U-Pb concordia diagrams for zircon from gabbronorite (A) and from anorthosite (B) of the Mt. Generalskaya intrusion (see Table 1).

Legend:

1- large gabbro-dolerite dykes, 2 –metagabbroid rock of the Umbarechka-Imandra complex, 3 – metasedimentary and metavolcanic rocks of the Imandra-Varzuga zone, 4 – lherzolite, websterite, orthopyroxenite and gabbronorite of the Ostrovsky intrusion, 5 - troktolite, 6 – large norite, orthopyroxenite and gabbro dykes, 7 – gabbro-anorthosite of the Chunutundra massif, 8 - metagabbro, gabbronorite and alternating orthopyroxenite and norite of the Monchetundra intrusion, 9 – cataclasis and recrystallization of gabbroid rocks of the Chuna and Monchetundra massifs, and that of Archean amphibolite, gneiss and diorite, 10 – norite, diorite and granophyric quartz diorite of the Yarva-Varaka massif, 11-14 Monchepluton: 11 - metagabbro, gabbronorite and anorthosite of the Vurechuaivench Foothills, 12 – olivine norite, norite, gabbronorite of the Nyud-Poaz, 13 - peridotite and pyroxenite of the NKT 14 – dunite of the Sopcheozero (Dunite) Block, 15 – diorite (a) and metagabbro (б) of the Xth anomaly, 16- Archean basement rocks; 17 – faults, 18 – dip, 19 – numbered boreholes.

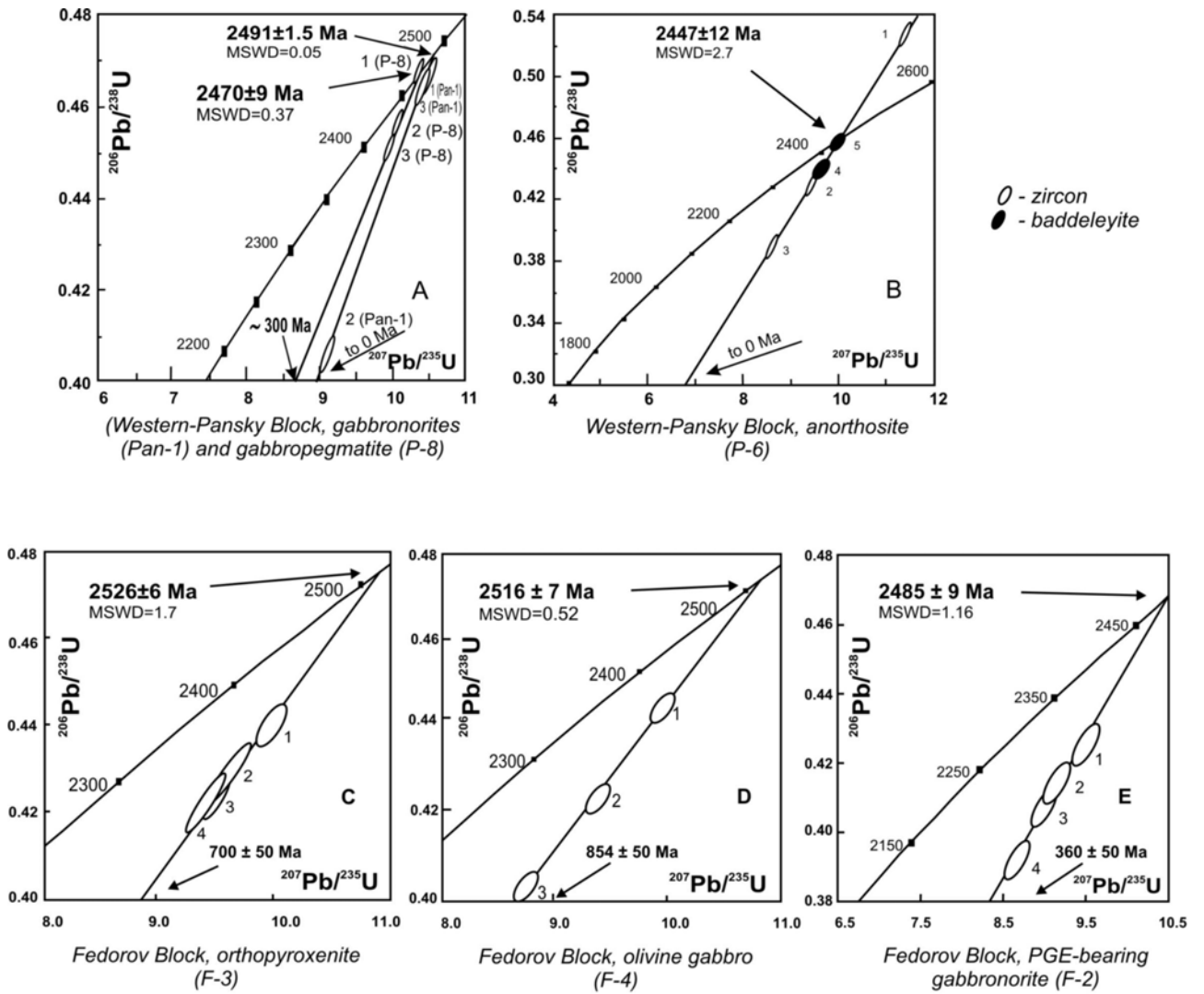
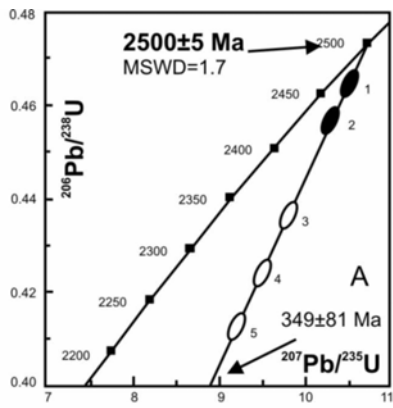
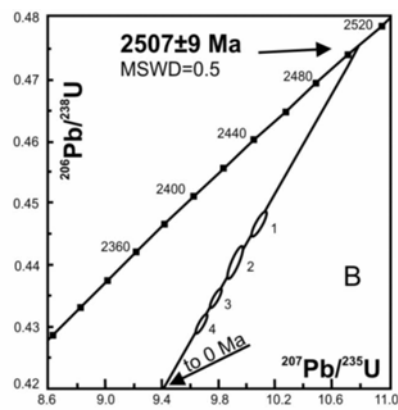


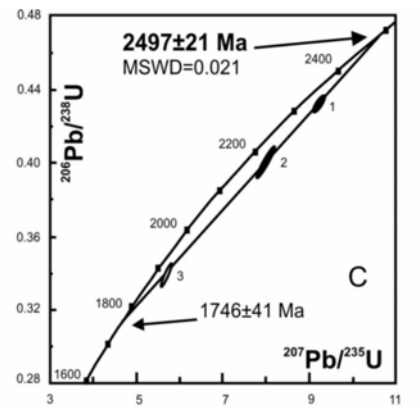
Fig. 7. U-Pb concordia diagrams for the Western-Pansky (A, B) and Fedorov (C, D, E) Blocks of the Fedorovo-Pansky Complex (see Table 2 for U-Pb data and references).



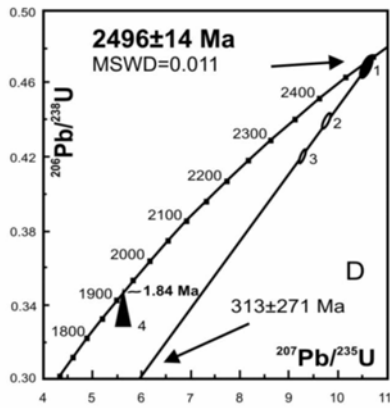
Mt. Nyud, Monchepluton, gabbronite of "critical horizon" (M-2)



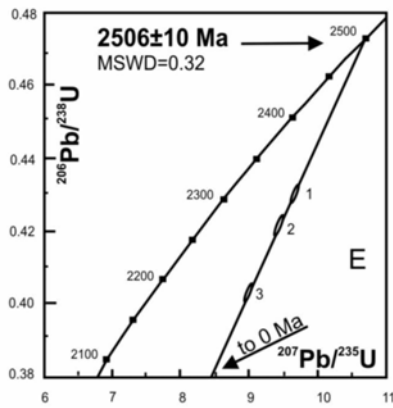
Mt. Travyanaya, marginal zone, Monchepluton, norite (M-1)



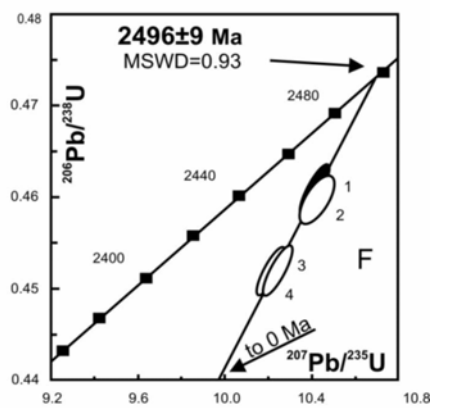
Vurechuaivench Foothills, Monchepluton, coarse-grained metagabbronite (M-42)



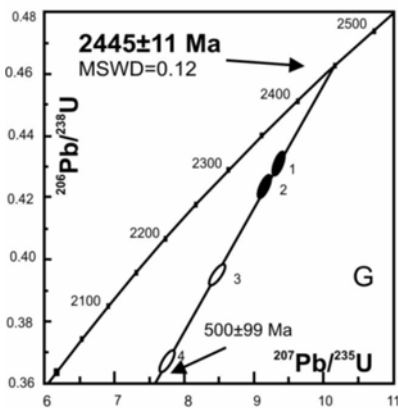
Dunite block, coarse-grained gabbronite dyke cutting ultramafic rocks, Borehole 1586 (M-14)



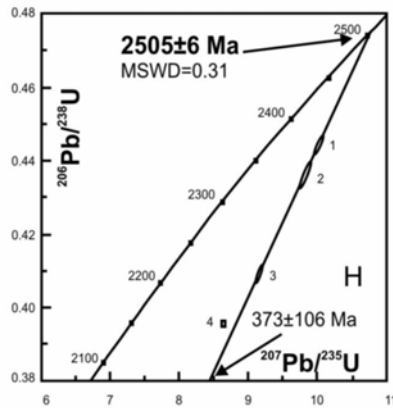
Dunite block, Monchepluton, coarse-grained gabbronite dyke cutting ultramafic rocks, Borehole 1518 (M-12)



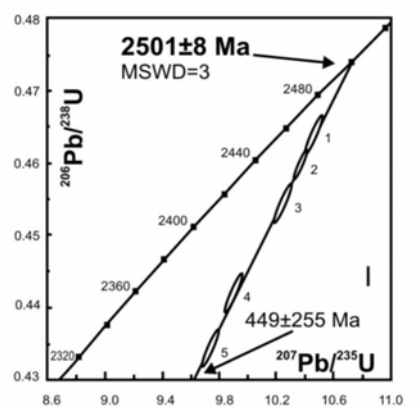
Mt. Yarva-Varaka, diorite (M-38)



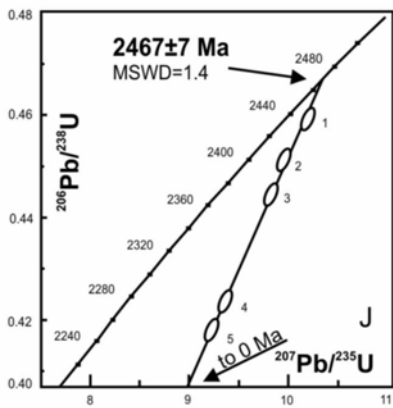
Ostrovsky intrusion, gabbronite-pegmatite (M-7)



Monchetundra, trachtyoid gabbronite (M-55)



Monchetundra, trachtyoid gabbronite (M54)



Chunutundra, trachtyoid anorthosite (M-16)

- - zircon
- - baddeleyite
- ▲ - secondary rutile

Fig. 8. U-Pb concordia diagrams for zircon, baddeleyite and rutile from different rocks of the Monchegorsk Layered Complex (see Table 3).

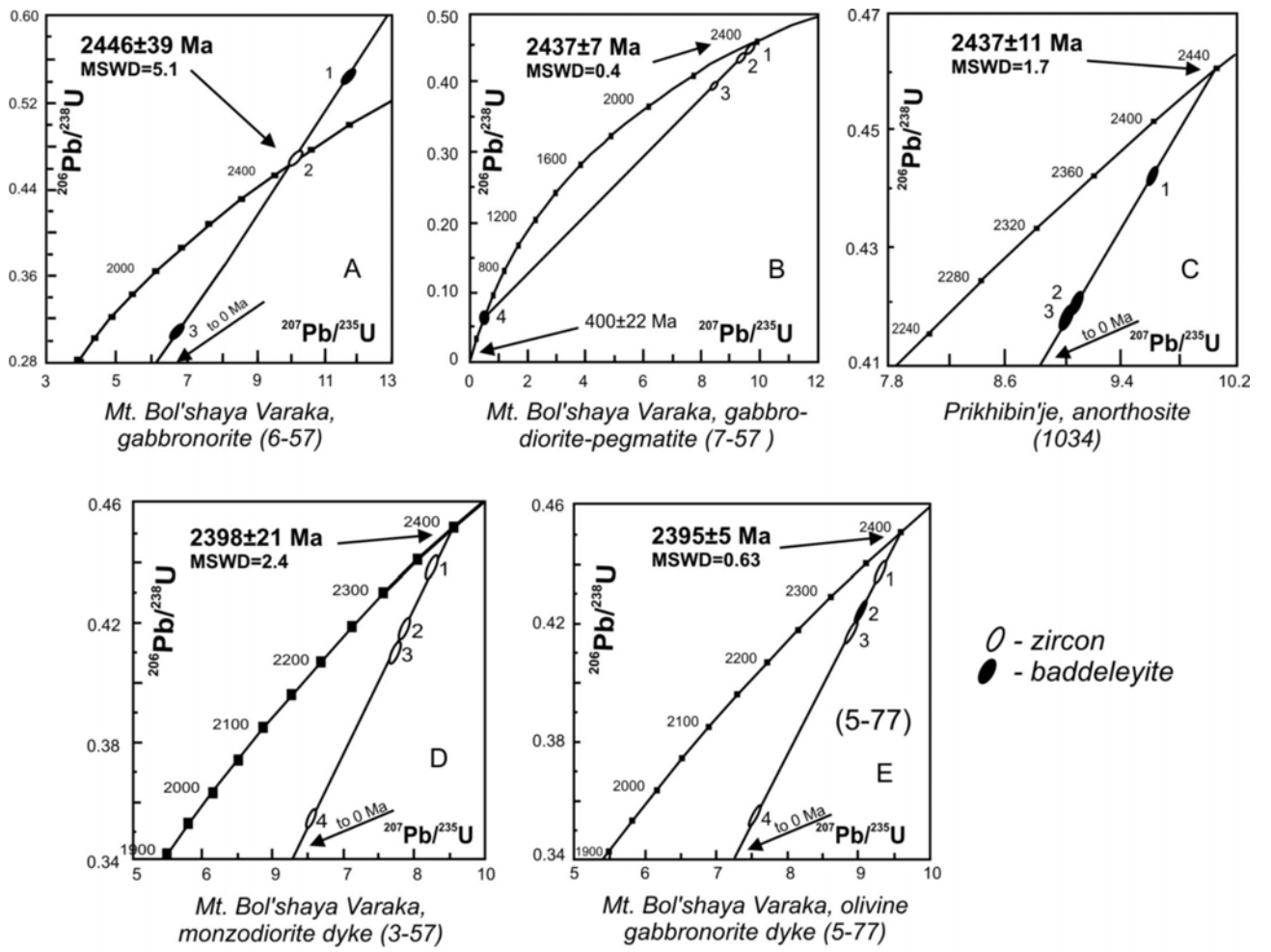


Fig. 9. U-Pb concordia diagrams for zircon and baddeleyite from rocks of the Imandra lopolith (see Table 4).

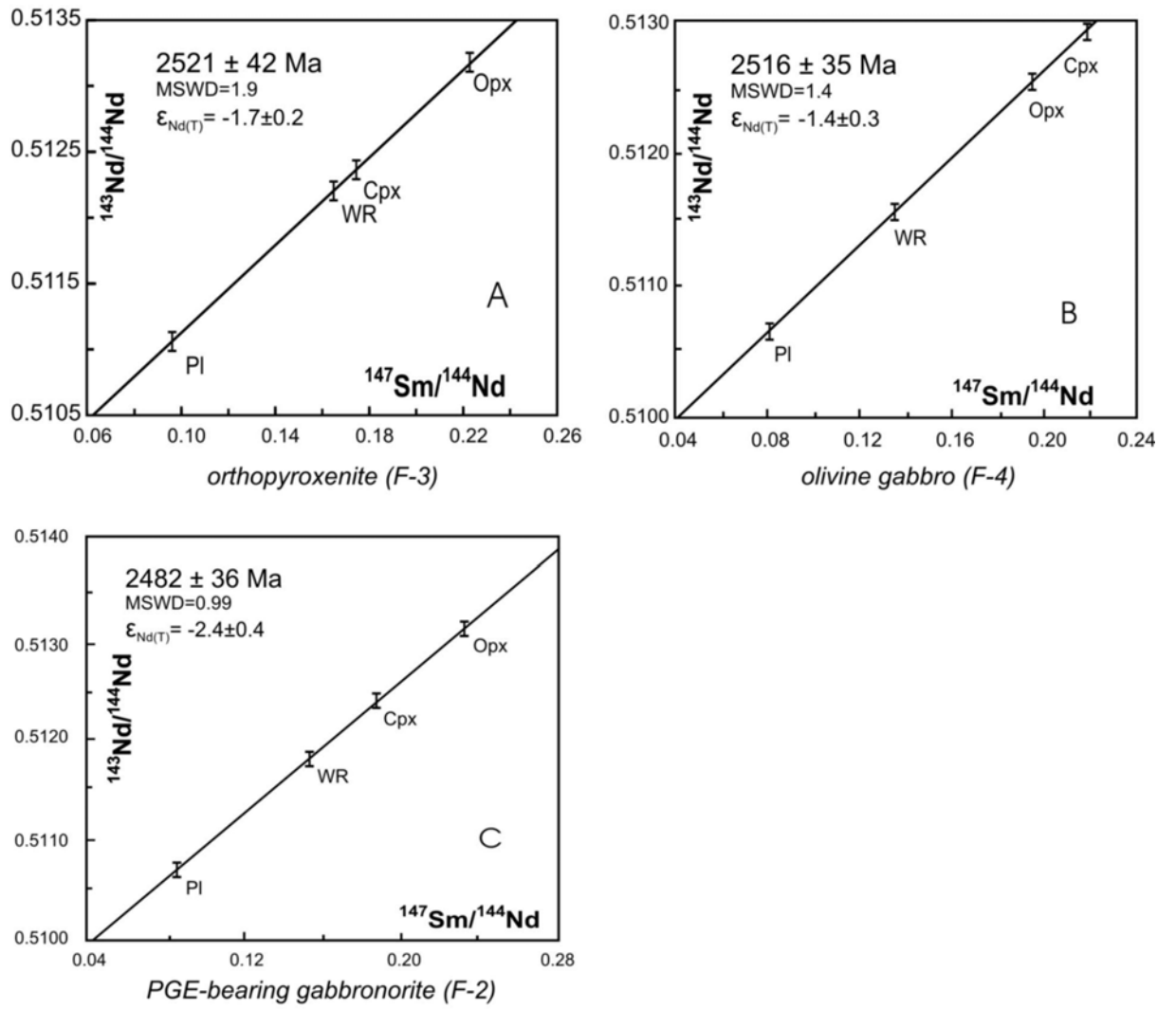


Fig. 10. Mineral Sm-Nd isochrons for rocks and rock-forming minerals of the Fedorov Block of the Fedorovo-Pansky Complex (Serov et al., 2007; see Table 6).

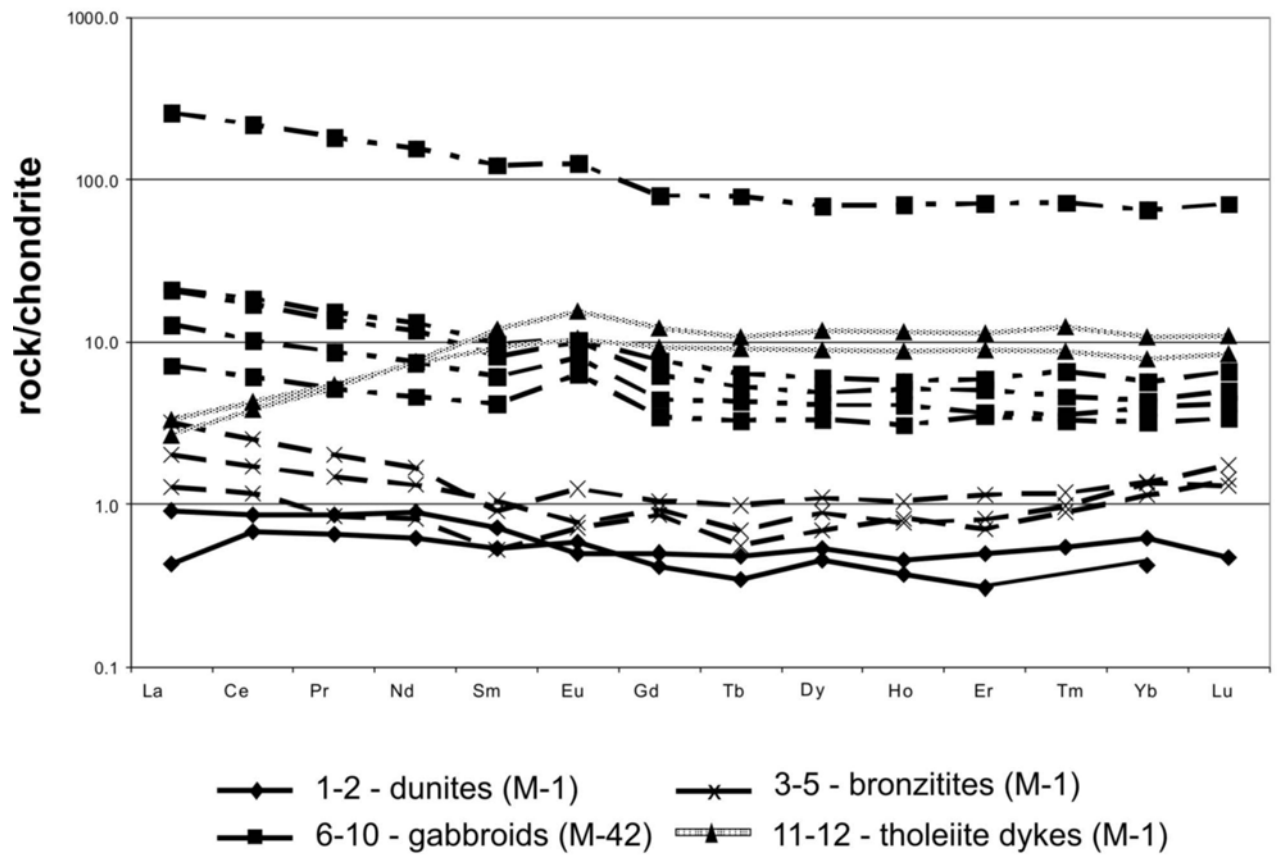


Fig. 11. REE data for rocks of the Monchepluton (plotted according to the data given in Table 7).

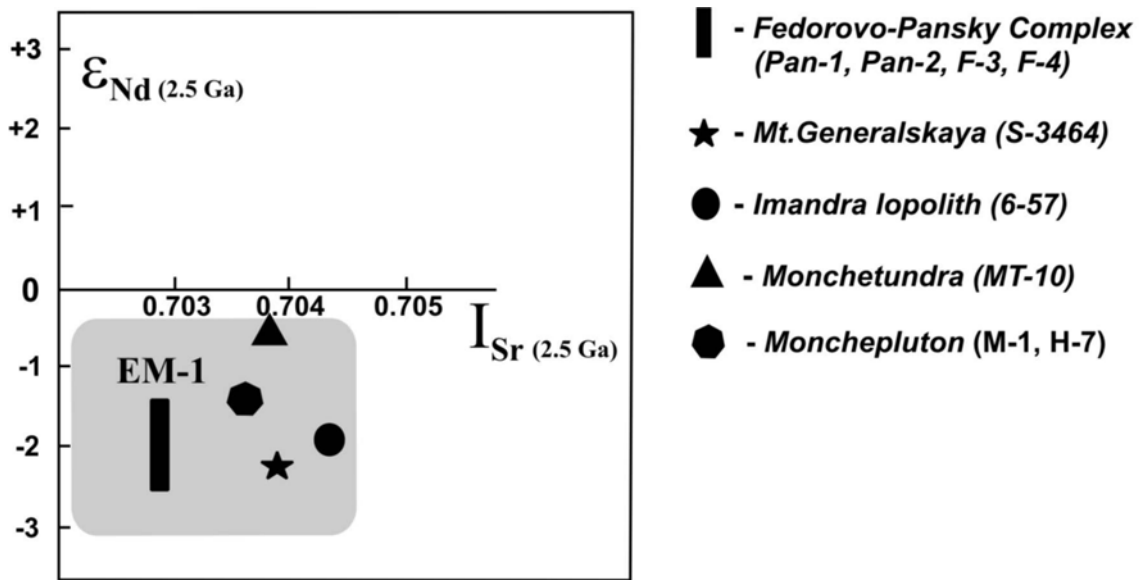


Fig. 12. ϵ_{Nd} - I_{Sr} plot of rocks from the Northern (Kola) Belt layered intrusions. Grey colour in the diagram shows EM-1 reservoir plotted for the layered intrusions of the Kola Peninsula based on the Sm-Nd and Rb-Sr isotope data given in Table 9.

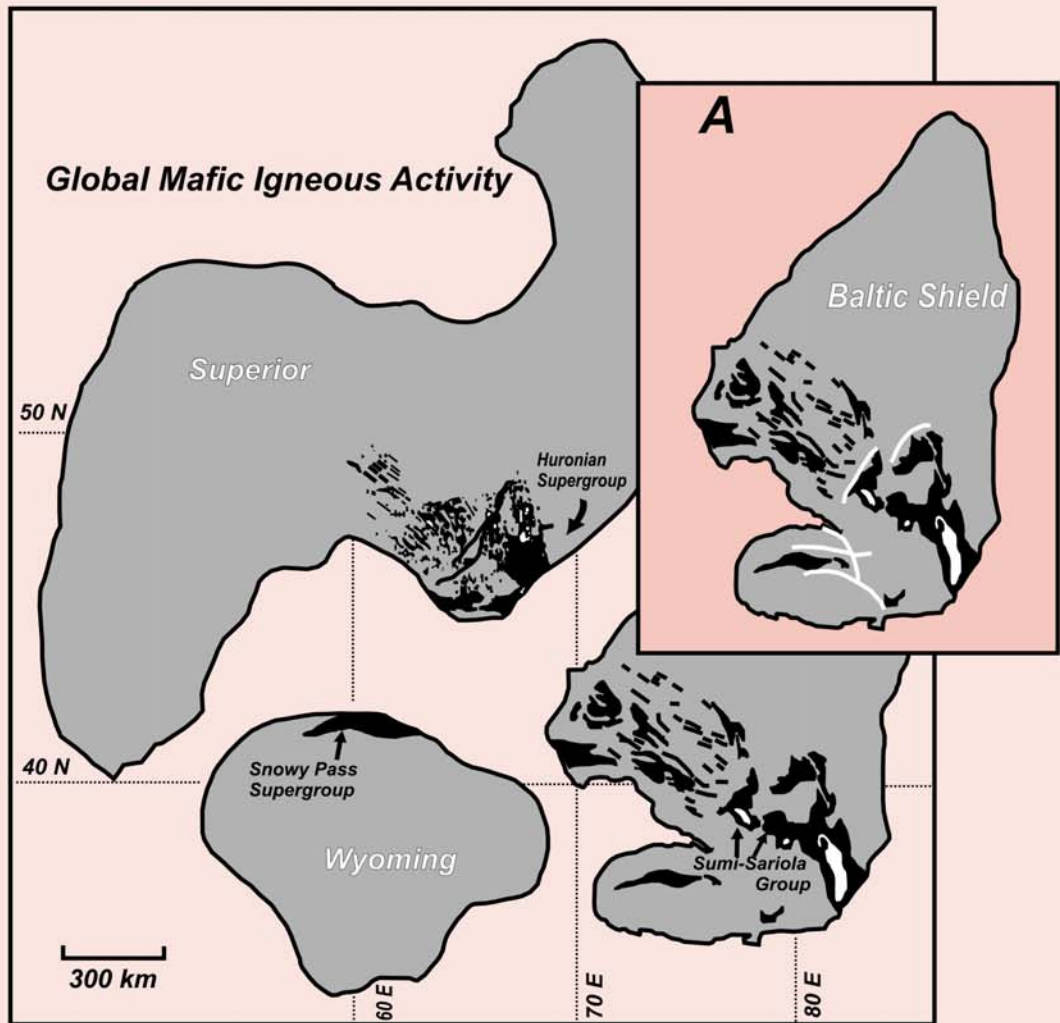


Fig. 13. Continental reconstruction at 2.45 Ga involving the Superior, Wyoming, and Karelian cratons (Heaman, 1997). Patterned regions correspond to 2.45 Ga supracrustal rocks, including flood basalts, such as Huronian Supergroup (Superior), Snowy Pass Supergroup (Wyoming), and Sumi-Sarlola-Strelna Supergroups (Karelia). Orientation of Karelia is based on alignment of Hearst and Karelian dike swarms and interpretation that they represent rift-parallel dykes. Estimated paleolatitudes are determined from paleomagnetic studies; paleolongitudes are arbitrary.

Black colour depicts Paleoproterozoic mafic magmatism (layered intrusions and mafic dykes). **A** - shows trends (/ /) of 2.52-2.44 Ga layered intrusions located on of the N-E Baltic Shield.

Table 1. *U-Pb zircon isotope data for the rocks of the Mt. Generalskaya intrusion (from Bayanova et al., 1999)*

Sample No	Weight (mg)	Concentration (ppm)		Pb isotopic composition ¹			Isotopic ratios ²		Age ² (Ma)
		Pb	U	$\frac{^{206}\text{Pb}}{^{204}\text{Pb}}$	$\frac{^{206}\text{Pb}}{^{207}\text{Pb}}$	$\frac{^{206}\text{Pb}}{^{208}\text{Pb}}$	$\frac{^{207}\text{Pb}}{^{235}\text{U}}$	$\frac{^{206}\text{Pb}}{^{238}\text{U}}$	$\frac{^{207}\text{Pb}}{^{206}\text{Pb}}$
<i>(gabbro norite (S-3464))</i>									
1	0.7	68.0	112.1	2700	5.931	2.783	10.4131	0.45964	2501
2	0.9	157.1	275.5	5200	6.046	2.972	9.8889	0.43927	2490
3	0.8	107.5	194.5	4200	6.020	2.970	9.5879	0.42551	2491
4	0.8	121.3	246.3	1585	5.835	2.881	8.4272	0.37318	2495
<i>(gabbro norite (SA-416))</i>									
5	1.3	76.4	227.5	2880	6.010	2.512	5.5480	0.24787	2480
<i>(zircon xenocrysts (S-3464))</i>									
6	0.9	133.9	296.1	1000	5.311	5.050	8.9624	0.37118	2607
7	0.6	113.9	343.5	1400	5.281	5.621	6.9341	0.27819	2660
<i>(anorthosite (SA-443))</i>									
1	0.8	109.4	199.8	4600	6.170	3.212	9.4737	0.43046	2452
2	0.8	115.9	208.6	1100	5.851	2.870	9.2448	0.41956	2454
3	0.5	52.6	101.9	1750	5.991	3.131	8.8624	0.40153	2456
<i>(zircon xenocrysts (SA-443))</i>									
4	0.5	369.9	400.1	2600	5.641	1.023	11.7529	0.49322	2585

¹All ratios are corrected for blanks of 0.1 ng for Pb and 0.04 ng for U and for mass discrimination of $0.17 \pm 0.05\%$

²Correction for common Pb was determined for the age according to Stacey and Kramers (1975).

Table 2. U-Pb baddeleyite (bd) and zircon isotope data from the Western-Pansky and Fedorov Blocks of the Fedorovo-Pansky Complex

Sample No	Weight (mg)	Concentration (ppm)		Pb isotopic composition ¹			Isotopic ratios ²		Age ² (Ma)
		Pb	U	$\frac{^{206}\text{Pb}}{^{204}\text{Pb}}$	$\frac{^{206}\text{Pb}}{^{207}\text{Pb}}$	$\frac{^{206}\text{Pb}}{^{208}\text{Pb}}$	$\frac{^{207}\text{Pb}}{^{235}\text{U}}$	$\frac{^{206}\text{Pb}}{^{238}\text{U}}$	$\frac{^{207}\text{Pb}}{^{206}\text{Pb}}$
<i>(Western-Pansky Block, gabbro (Pan-1); from Bayanova, 2004)</i>									
1	3.30	95.0	144	11740	6.091	3.551	10.510	0.4666	2491
2	1.90	70.0	142	10300	6.100	4.220	9.135	0.4061	2489
3	1.60	84.0	144	6720	6.062	3.552	10.473	0.4650	2491
<i>(Western-Pansky Block, gabbro (P-8); from Balashov et al., 1993)</i>									
1	5.90	95.0	158	3240	5.991	3.081	10.435	0.4681	2471
2	7.30	181.0	287	8870	6.161	2.260	10.092	0.4554	2465
3	1.25	125.0	200	3400	6.012	2.312	10.082	0.4532	2468
<i>(Western-Pansky Block, anorthosite (P-6); from Bayanova, 2004)</i>									
1	0.75	218.0	322	5740	6.230	3.263	11.682	0.5352	2438
2	0.10	743.0	1331	3960	6.191	3.151	9.588	0.4393	2438
3	0.20	286.0	577	2980	6.021	3.192	8.643	0.3874	2474
4 (bd)	1.00	176.0	396	14780	6.290	63.610	9.548	0.4380	2435
5 (bd)	0.26	259.0	560	3360	6.132	54.950	9.956	0.4533	2443
<i>(Fedorov Block, orthopyroxenite (F-3); from Nitkina, 2006)</i>									
1	0.75	48.0	60.9	825	4.9191	1.3039	10.0461	0.44249	2504
2	0.80	374.0	598.6	4588	6.0459	1.9650	9.6782	0.43153	2484
3	0.85	410.2	630.2	4521	6.0281	1.6592	9.5667	0.42539	2488
4	1.00	271.0	373.1	2552	5.9916	1.2393	9.4700	0.42406	2476
<i>(Fedorov Block, olivine gabbro (F-4); from Nitkina, 2006)</i>									
1	1.80	725.3	1322.8	14649	6.1121	3.8177	10.0132	0.44622	2484
2	2.00	731.3	1382.8	8781	6.1522	3.5517	9.4306	0.42454	2467
3	1.95	680.9	1374.0	7155	6.2645	3.6939	8.7401	0.40155	2433
<i>(Fedorov block, PGE-bearing gabbro (F-2); from Nitkina, 2006)</i>									
1	0.30	498.0	833.4	2081	5.9502	2.2111	9.49201	0.42493	2477
2	0.65	513.8	932.2	5274	6.1519	2.6371	9.1373	0.41378	2458
3	0.55	583.2	999.3	3194	6.1132	2.0528	8.9869	0.40832	2452
4	0.80	622.5	1134.5	4114	6.1161	2.1914	8.6638	0.39165	2460

¹All ratios are corrected for blanks of 0.1 ng for Pb and 0.04 ng for U and for mass discrimination of $0.17 \pm 0.05\%$.

²Correction for common Pb was determined for the age according to Stacey and Kramers (1975).

Table 3. U-Pb baddeleyite (bd), zircon and rutile (ru) isotope data from the Monchegorsk Layered Complex

Sample No	Weight (mg)	Concentration (ppm)		Pb isotopic composition ¹			Isotopic ratios ²		Age ² (Ma)
		Pb	U	$\frac{^{206}\text{Pb}}{^{204}\text{Pb}}$	$\frac{^{206}\text{Pb}}{^{207}\text{Pb}}$	$\frac{^{206}\text{Pb}}{^{208}\text{Pb}}$	$\frac{^{207}\text{Pb}}{^{235}\text{U}}$	$\frac{^{206}\text{Pb}}{^{238}\text{U}}$	$\frac{^{207}\text{Pb}}{^{206}\text{Pb}}$
<i>("critical horizon", Mt. Nyud Monchepluton, gabbro-norite (M-2); from Bayanova, 2004)</i>									
1 (bd)	0.70	93.1	198.8	9432	6.0586	103.8300	10.4643	0.4636	2495
2 (bd)	0.40	170.8	364.4	3589	5.9833	50.7070	10.3199	0.4574	2494
3	0.40	117.4	183.1	4590	6.0153	1.8703	9.8274	0.4359	2492
4	0.80	187.9	308.4	13664	6.1169	1.9750	9.4740	0.4227	2483
5	0.50	152.2	252.5	5300	6.0842	1.8994	9.2129	0.4125	2477
<i>(marginal zone, Mt. Travyanaya Monchepluton, norite (M-1); from Smolkin et al., 2004)</i>									
1	0.30	308.3	504.8	5778	6.0202	2.3805	10.0760	0.4458	2497
2	0.35	185.4	319.8	8358	6.0582	2.7721	9.9277	0.4402	2493
3	0.40	264.5	441.6	23762	6.1006	2.2929	9.7814	0.4342	2491
4	0.40	434.8	793.1	6273	6.0541	3.2613	9.7060	0.4314	2489

<i>(Vurechuaivench Foothills Monchepluton, coarse-grained metagabbro (M-42); present study)</i>									
1 (bd)	0.80	150.1	271.4	2982	6.3099	3.2054	9.23762	0.43446	2393
2 (bd)	0.65	65.1	122.6	2080	6.5863	2.7920	8.13574	0.40516	2295
3	0.75	137.4	288.5	911	6.4805	2.3018	5.75090	0.34208	2228
<i>(Dunite block, Monchepluton, coarse-grained gabbro dyke cutting ultramafic rocks, hole 1586 (M-14); from Bayanova, 2004)</i>									
1 (bd)	0.50	5.3	10.3	1307	5.7748	12.4320	10.5720	0.4684	2494
2	0.80	358.7	309.2	13360	6.1029	0.5312	9.8622	0.4391	2486
3	0.60	321.8	362.1	3791	6.0407	0.7838	9.3919	0.4199	2479
4(ru) ³	1.30	7.5	4.5	28	1.7085	0.8077	5.7139	0.3328	2022
<i>(Dunite block, Monchepluton, coarse-grained gabbro dyke cutting ultramafic rocks, hole 518 (M-12); from Smolkin et al., 2004)</i>									
1	0.45	221.4	409.6	2152	5.9237	3.5776	9.6682	0.4303	2487
2	0.30	321.7	542.6	11260	6.1264	2.2049	9.4976	0.4249	2478
3	0.50	164.9	302.4	1952	5.9508	2.5602	8.9806	0.4031	2472
<i>(Mt. Yarva-Varaka, diorite (M-38); from Bayanova, 2004)</i>									
1 (bd)	0.50	32.9	70.0	5615	6.0158	53.309	10.419	0.4608	2497
2	0.80	310.5	515.4	2587	5.9089	2.9118	10.420	0.4597	2501
3	1.40	151.8	262.7	4840	5.9895	3.1873	10.242	0.4519	2501
4	0.70	273.2	472.5	4590	5.9970	3.1828	10.217	0.4518	2497
<i>(Ostrovsky intrusion, gabbro-pegmatite (M-7); from Bayanova, 2004)</i>									
1 (bd)	0.45	28.6	63.9	1820	6.162	29.610	9.350	0.4311	2405
2 (bd)	0.55	36.3	83.7	3380	6.346	42.038	9.210	0.4248	2389
3	0.45	336.4	694.8	9420	6.378	3.797	8.471	0.3953	2407
4	0.35	89.1	187.6	3700	6.375	2.942	7.756	0.3667	2384
<i>(Monchetundra, trachytoid gabbro (M-55); from Smolkin et al., 2004)</i>									
1	0.50	110.9	172.9	8690	6.0591	1.9737	10.0171	0.444092	2493
2	0.35	37.7	61.0	1122	5.7260	2.2170	9.82794	0.436123	2492
3	0.25	168.3	277.7	6350	6.0914	1.8221	9.15540	0.409417	2479
4	0.30	122.6	213.5	6159	6.2294	1.9243	8.64155	0.395489	2439
<i>(Monchetundra, trachytoid gabbro (M-54); from Smolkin et al., 2004)</i>									
1	0.50	308.9	494.5	9172	6.0283	2.4881	10.47020	0.46359	2503
2	0.35	374.3	587.5	18868	6.0791	2.2742	10.40220	0.46050	2496
3	0.40	72.8	118.6	6833	6.0271	2.5023	10.25210	0.45405	2498
4	0.25	206.3	333.1	7831	6.0324	2.1148	9.90668	0.44177	2499
5	0.45	196.6	311.9	14844	6.1123	1.9269	9.74196	0.43412	2484
<i>(Chunatundra, trachytoid anorthosite (M-16); from Bayanova, 2004)</i>									
1	0.20	80.3	138.4	2710	6.020	3.404	10.216	0.4589	2471
2	2.05	122.83	214.1	4420	6.144	3.293	9.999	0.4527	2455
3	0.30	141.3	251.0	5140	6.137	3.291	9.831	0.4443	2461
4	0.60	92.7	169.4	7860	6.148	2.970	9.388	0.4228	2467
5	0.20	46.5	86.1	1090	5.805	3.242	9.262	0.4180	2463

¹All ratios are corrected for blanks of 0.08 ng for Pb и 0.04 ng for U and for mass discrimination of 0.12±0.04%.

²Correction for common Pb was determined for the age according to Stacey and Kramers (1975).

³Corrected for isotope composition of light cogenetic plagioclase: $^{206}\text{Pb}/^{204}\text{Pb}=14.041\pm 0.005$, $^{207}\text{Pb}/^{204}\text{Pb}=14.581\pm 0.007$, $^{208}\text{Pb}/^{204}\text{Pb}=35.58\pm 0.02$.

Table 4. U-Pb baddeleyite (bd) and zircon isotope data from the rocks of the Imandra lopolith (from Bayanova, 2004)

Sample No	Weight (mg)	Concentration (ppm)		Pb isotopic composition ¹			Isotopic ratios ²		Age (Ma)
		Pb	U	$\frac{^{206}\text{Pb}}{^{204}\text{Pb}}$	$\frac{^{206}\text{Pb}}{^{207}\text{Pb}}$	$\frac{^{206}\text{Pb}}{^{208}\text{Pb}}$	$\frac{^{207}\text{Pb}}{^{235}\text{U}}$	$\frac{^{206}\text{Pb}}{^{238}\text{U}}$	$\frac{^{207}\text{Pb}}{^{206}\text{Pb}}$
<i>(Mt. Bol'shaya Varaka, gabbro-norite (6-57))</i>									
1 (bd)	0.60	257	463	5500	6.141	46.2300	12.0200	0.5435	2461
2	1.10	123	183	8040	6.231	2.0930	10.3600	0.4730	2445
3 (bd)	0.70	170	534	2570	6.162	31.0400	6.7081	0.3088	2430
<i>(Mt. Bol'shaya Varaka, gabbro-diorite-pegmatite (7-57))</i>									
1	0.90	274	474	8100	6.248	3.1196	9.9036	0.4532	2440
2	0.50	71	129	2310	6.108	3.4370	9.5718	0.4387	2437
3	0.60	125	232	2770	6.132	3.4848	9.4215	0.4311	2440
4 (bd)	0.30	60	128	1220	6.034	24.2930	9.4746	0.4425	2405
<i>(Prihibin'je, anorthosite (1034))</i>									
1 (bd)	0.90	457.9	1032.6	28560	6.334	117.5	9.6090	0.4419	2431
2 (bd)	1.00	174.6	410.7	6280	6.316	87.0	9.1080	0.4219	2419
3 (bd)	0.80	330.6	790.5	14210	6.351	93.5	9.0000	0.4162	2422
<i>(Mt. Bol'shaya Varaka, monzodiorite dyke (3-57))</i>									
1	0.50	119	102	17330	5.821	0.5701	9.6611	0.4530	2397
2	0.71	114	104	1910	5.933	0.5801	9.2121	0.4322	2398
3	1.30	103	96	1530	5.612	0.5601	8.6314	0.4060	2393
4	0.50	152	184	1960	5.971	0.5812	6.9612	0.3260	2396
<i>(Mt. Bol'shaya Varaka, olivine gabbro-norite dyke (5-77))</i>									
1	0.82	127	241	1630	6.173	4.5011	9.2951	0.4372	2393
2 (bd)	0.60	104	212	1490	5.547	8.2507	9.0262	0.4237	2397
3	0.63	332	607	1280	6.072	2.9501	8.9004	0.4171	2399
4	0.68	286	626	1590	6.144	3.1284	7.5613	0.3543	2399

¹The ratios are corrected for blanks of 0.08 ng for Pb and 0.04 ng for U and for mass discrimination $0.17 \pm 0.05\%$.

²Correction for common Pb was determined for the age according to Stacey and Kramers (1975).

Table 5. Summary of U-Pb and Sm-Nd geochronology for layered intrusions located in the eastern Baltic Shield

Layered intrusions	Age (Ma)		$\epsilon_{\text{Nd(T)}} @$ U/Pb age
	U-Pb	Sm-Nd	
<u>Northern belt</u>			
<u>Mt. Generalskaya</u>			
gabbro	2496±10 ¹ (2505±1.6) ²	2453±42 ¹	-2.3
anorthosite	2446±10 ¹		
<u>Monchepluton</u>			
Mt. Travyanaya, norite	2507±9 ¹⁵ ;		
Dunit block, gabbro dyke	2506±10 ¹⁵ ; 2496±14 ¹⁵		
Nyud Terrace, gabbro	2500±5 ¹⁴		
Nyud Terrace, gabbro	2493±7 ¹ (2504±1.5) ²	2492±31 ³	-1.4
Vurechuaivench Foothills, metagabbro	2497±21 ¹⁵		
<u>Main Ridge</u>			
Monchetundra, gabbro	2463±25 ⁴ ; 2453±4 ⁵		
Monchetundra, gabbro	2505±6 ¹⁴ ; 2501±8 ¹⁴		
Chunatundra, anorthosite	2467±7 ¹⁵		
Ostrovsky intrusion			
gabbro-pegmatite	2445±11 ¹⁵		
<u>Fedorov-Pansky Complex</u>			
orthopyroxene	2526±6 ¹²	2521 ± 42 ¹³	-1.7
olivine gabbro	2516±7 ¹²	2516 ± 35 ¹³	-1.4
magnetite gabbro	2498±5 ⁶		
gabbro	2491±1.5 ⁷ (2501±1.7) ²	2487±51 ⁷	-2.1
Cu-Ni PGE-bearing gabbro	2485±9 ¹²	2482 ± 36 ¹³	-2.4
PGE-gabbro-pegmatite	2470±9 ⁷		
PGE-anorthosite	2447±12 ⁷		
<u>Imandra lopolith</u>			
gabbro	2446±39 ⁷ (2441±1.6) ²	2444±77 ⁷	-2.0

gabbro-diorite-pegmatite	2440±4 ⁶		
norite	2437±7 ⁶		
leucogabbro-anorthosite	2437±11 ⁶		
granophyre	2434±15 ⁶		
olivine gabbronorite (dyke)	2395±5 ⁶		
monzodiorite dyke	2398±21 ⁶		
<u>Southern belt</u>			
Kivakka			
olivine gabbronorite	2445±2 ⁷	2439±29 ⁸	-1.2
Lukkulaisvaara			
pyroxenite	2439±11 ⁷ (2442±1.9) ²	2388±59 ⁸	-2.4
Tsipringa			
gabbro	2441±1.2 ²	2430±26 ⁸	-1.1
Burakovskaya intrusion			
gabbronorite	2449±1.1 ²	2365±90 ⁸	-2.0
Kovdozero intrusion			
pegmatoid gabbronorite	2436±9 ⁶		
Finnish group			
Koitelainen	2433±8 ⁹	2437±49 ¹¹	-2.0
Koilismaa	2436±5 ¹⁰		
Nyaryankavaara	2440±16 ¹⁰		
Penikat		2410±64 ⁹	-1.6
Akanvaara	2437±7 ¹¹	2423±49 ¹¹	-2.1

- | | | |
|------------------------------|---------------------------|---------------------------------|
| 1. Bayanova et al., 1999 | 6. Bayanova, 2004 | 11. Hanski et al., 2001 |
| 2. Amelin et al., 1995 | 7. Balashov et al., 1993 | 12. Nitkina, 2006 |
| 3. Tolstikhin et al., 1992 | 8. Amelin, Semenov, 1990 | 13. Serov et al., 2007 |
| 4. Vrevsky, Levchenkov, 1992 | 9. Huhma et al., 1990 | 14. Bayanova & Mitrofanov, 2005 |
| 5. Mitrofanov et al., 1993 | 10. Alapieti et al., 1990 | 15. present study |

Table 6. *Sm-Nd isotope data on whole rock and mineral separates of the Fedorov Block of the Fedorovo-Pansky Complex (from Serov et al., 2007)*

Sample No	Concentration (ppm)		Isotopic ratios		T _{DM} (Ga)	Sm-Nd (Ma)	ε _{Nd} (2.5Ga)
	Sm	Nd	¹⁴⁷ Sm/ ¹⁴⁴ Nd	¹⁴³ Nd/ ¹⁴⁴ Nd			
<i>(orthopyroxenite (F-3))</i>							
WR	0.32	1.17	0.1648	0.512196±12	3.05	2521±42	-1.73
Opx	0.12	0.38	0.2228	0.513182±16			
Cpx	2.21	7.67	0.1745	0.512349±17			
Pl	0.26	1.62	0.0960	0.511071±29			
<i>(olivine gabbro (F-4))</i>							
WR	0.63	2.80	0.1357	0.511548±8	2.94	2516±35	-1.53
Opx	0.23	0.72	0.1951	0.512555±15			
Cpx	0.83	2.28	0.2187	0.512947±16			
Pl	0.24	1.77	0.0815	0.510677±14			
<i>(PGE-bearing gabbronorite (F-2))</i>							
WR	0.42	1.66	0.1537	0.511807±20	3.18	2482±36	-2.50
Pl	0.41	2.88	0.0865	0.510709±14			
Cpx	1.78	5.73	0.1876	0.512387±8			
Opx	0.13	0.33	0.2323	0.513088±40			

Table 7. *REE data for rocks of the Monchepluton*

Sample No	1	2	3	4	5	6	7	8	9	10	11	12
	<i>dunites (M-1)</i>		<i>bronzitites (M-1)</i>			<i>gabbroids (M-42)</i>				<i>tholeiite dykes (M-1)</i>		
La	0.216	0.102	0.754	0.479	0.302	5.020	3.030	61.30	4.910	1.700	0.635	0.780
Ce	0.532	0.419	1.540	1.050	0.716	11.400	6.290	134.00	10.500	3.740	2.340	2.614
Pr	0.082	0.063	0.192	0.140	0.081	1.450	0.825	17.40	1.310	0.489	0.495	0.520
Nd	0.420	0.294	0.784	0.614	0.383	6.160	3.550	73.10	5.510	2.150	3.540	3.437
Sm	0.110	0.083	0.140	0.161	0.081	1.500	0.938	18.80	1.250	0.637	1.820	1.415
Eu	0.029	0.034	0.072	0.045	0.042	0.596	0.464	7.32	0.585	0.368	0.896	0.605
Gd	0.104	0.086	0.216	0.192	0.178	1.590	0.911	16.40	1.290	0.713	2.500	1.892
Tb	0.018	0.013	0.037	0.026	0.021	0.238	0.161	2.97	0.198	0.123	0.399	0.339
Dy	0.137	0.117	0.278	0.227	0.178	1.530	1.040	17.50	1.240	0.847	2.970	2.266
Ho	0.026	0.021	0.059	0.044	0.047	0.325	0.231	3.98	0.289	0.175	0.657	0.497
Er	0.083	0.051	0.190	0.135	0.118	0.980	0.607	11.80	0.842	0.578	1.880	1.477
Tm	0.014	<1.d.	0.030	0.025	0.023	0.169	0.091	1.84	0.118	0.084	0.315	0.224
Yb	0.107	0.072	0.232	0.230	0.193	0.968	0.670	11.10	0.742	0.548	1.810	1.325
Lu	0.012	<1.d.	0.044	0.033	0.036	0.168	0.107	1.81	0.127	0.086	0.279	0.213

REE determinations were measured by Ion probe Cameca 4F analyses by E. Deloule at the Centre de Recherches Petrographiques et Chimiques – CRPG-CNRS, Nancy. The data are provided by the Russian-French project RFFI-CNRS 01-05-22001 that was headed by the authors of the present study.

* - <1.d.- below limit of determination

Table 8. *Sm-Nd isotope analyses and model ages for the rocks of the Monchegorsk Layered Complex*

Average standard values: N=11 (La Jolla: = 0.511833±6); N=100 (JNdi1: =0.512098±15).

Sample	Rock type	Concentration (ppm)		Isotope ratios		T _{DM} (Ga)	ε _{Nd} (‰)
		Sm	Nd	¹⁴⁷ Sm/ ¹⁴⁴ Nd	¹⁴³ Nd/ ¹⁴⁴ Nd±2σ		
<i>(Monchepluton)</i>							
M-1	Mineralized norite	1.750	8.04	0.131957	0.511493 ±3	2.91	-1
M-2	Gabbro-norite-pegmatite	0.920	4.15	0.134055	0.511537 ±4	2.90	-1
M-42	Metagabbro-norite	1.480	6.67	0.134404	0.511462 ±3	2.85	-2
C-1	Orthopyroxenite	0.564	2.56	0.133100	0.511477 ±6	3.09	-2
H-4	Olivine pyroxenite	0.584	2.60	0.136000	0.511577 ±8	3.01	-1
H-7	Gabbro-norite	0.703	3.06	0.138900	0.511680 ±5	3.07	-1
<i>(Monchetundra)</i>							
M-54	Trachytoid gabbro-norite	0.940	3.54	0.159479	0.511963 ±5	3.08	-1
<i>(Chunatundra)</i>							
M-16	Trachytoid anorthosite	0.730	3.08	0.143365	0.511740 ±5	2.84	-0
<i>(Mt. Yarva-Varaka)</i>							
M-38	Granophyric quartz diorite	7.380	38.48	0.115923	0.511184 ±2	2.91	-2
<i>(Ostrovsky intrusion)</i>							
M-7	Gabbro-norite-pegmatite	3.820	18.69	0.123446	0.511333 ±2	2.90	-2
<i>(dyke rocks of the Monchepluton)</i>							
M-9	Coarse-grained melanonorite	0.469	2.24	0.126386	0.511354 ±5	2.96	-2
M-12	Gabbro-norite dyke	0.581	2.35	0.149344	0.511795 ±11	2.98	-1
M-14	Gabbro-norite dyke	1.770	8.86	0.120709	0.511573 ±4	2.40	3

Table 9. Sm-Nd and Rb-Sr isotope data for rocks of the Mt. Generalskaya, Fedorovo-Pansky, Imandra, Monchepluton and Monchetundra intrusions

Sample No	Concentration (ppm)		Isotopic ratios		$\epsilon_{\text{Nd}}(2.5 \text{ Ga})$	$T_{\text{DM}}(\text{Ga})$	$^{87}\text{Rb}/^{86}\text{Sr}$	$^{87}\text{Sr}/^{86}\text{Sr}(\pm 2\sigma)$ @2.5 Ga
	Sm	Nd	$^{147}\text{Sm}/^{144}\text{Nd}$	$^{143}\text{Nd}/^{144}\text{Nd}(\pm 2\sigma)$				
<i>(Mt. Generalskaya)</i>								
<i>S-3464, gabbro</i>	1.147	5.362	0.129320	0.511449±14	-2.30	2.91	0.00534	0.70421±22
<i>(Fedorov-Pansky intrusion)</i>								
<i>Pan-1, gabbro</i>	0.762	3.293	0.139980	0.511669±7	-2.00	2.98	0.00135	0.70315±10
<i>Pan-2, gabbro</i>	0.423	1.662	0.153714	0.511807±20	-2.50	3,18	0.00174	0.70295±17
<i>F-4, olivine gabbro</i>	0.629	2.801	0.135695	0.511548±8	-1.53	2.94	0.00144	0.70288±22
<i>F-3, orthopyroxenite</i>	0.318	1.166	0.164803	0.512196±12	-1.73	3.05	0.00205	0.70333±15
<i>(Imandra lopolith)</i>								
<i>6-57, gabbro</i>	2.156	10.910	0.119130	0.511380±3	-2.00	2.88	0.00339	0.70455±27
<i>(Monchepluton)</i>								
<i>M-1, quartz norite</i>	1.750	8.040	0.131957	0.511493 ±3	-1.51	2.91	0.01053	0.70341±9
<i>H-7, gabbro</i>	0.920	4.150	0.134055	0.511537 ±4	-1.37	2.90	0.00227	0.70368±24
<i>(Monchetundra)</i>								
<i>MT-10, medium-grained pyroxenite</i>	0.483	1.913	0.152689	0.511925±33	-0.36	2.81	0.00495	0.70399±17

Table 10. Isotopic $^4\text{He}/^3\text{He}$ ratios of PGE layered intrusions of the Baltic Shield (data are provided by Kamensky and Novikov, Geological Institute KSC RAS, unpublished)

Hole No/sampling depth (m)	Rock, mineral	$^4\text{He} \times 10^{-6}$ $\mu\text{cm}^3/\text{g}$	$^4\text{He}/^3\text{He}$ $\times 10^6$	Low Mantle contribution**%
<i>(Fedorovo-Pansky intrusion)</i>				
hole, Ki-16/6	<i>Amphibole</i>	81.00	9.10	0.11
hole, Ma-14/1	<i>Orthopyroxene</i>	9.90	12.80	0.08
outcrop, No 9	<i>Ilmenite</i>	43.90	16.50	0.06
<i>(Monchepluton, Mt. Sopcha)</i>				
hole, 995/315	<i>Olivinite, rock</i>	17.00	6.25	0.16
hole, 995/315	<i>Olivine</i>	25.00	5.88	0.17
hole, 995/315	<i>Orthopyroxene</i>	31.00	6.25	0.16
hole, 995/315	<i>Plagioclase</i>	47.00	5.56	0.18
hole, 995/315	<i>Magnetite</i>	132.00	4.35	0.23
<i>(Main Ridge, Monchetundra)</i>				
hole, 765/905,9	<i>Clinopyroxene</i>	163.00	4.76	0.21
hole, 765/905,9	<i>Orthopyroxene</i>	21.00	4.76	0.21
hole, 765/985,3	<i>Amphibole</i>	97.00	4.76	0.21
hole, 765/985,3	<i>Clinopyroxene</i>	115.00	5.00	0.20
outcrop, MT-5	<i>Gabbro</i>	1.30	2.00	0.41
<i>(Dunite block)</i>				
hole, 904/102	<i>Dunite, rock</i>	218.00	1.47	0.68
hole, 904/102	<i>Olivine</i>	115.00	1.35	0.74
hole, 1651/244,9	<i>Chromitite, ore</i>	56.00	1.43	0.70
hole, C-1651/373,5*	<i>Dunite-Bronzitite</i>	28.00	0.83	1.20
hole, C-1622/7*	<i>Chromitite, ore</i>	2.80	0.69	1.44
hole, C-1646/450*	<i>Dunite</i>	2.20	1.29	0.77
hole, C-1651/373.5*	<i>Dunite-Bronzitite-contact</i>	0.13	0.60	1.68

Note:

errors are according to the calculation method (Tolstikhin and Marty, 1998)

* Step wise heating experiment (fraction under the temperature 1300°C).

** Mantle components are given from value $^4\text{He}/^3\text{He}$ 0.55×10^4 (solar helium from lower mantle reservoir), Tolstikhin and Marty (1998).