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Evolution of Paleoproterozoic Magmatism: Geology, Geochemistry, and Isotopic Constraints

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Abstract—Evolution of tectono-magmatic processes in the Paleoproterozoic is divisible into four stages each controlled by its own type of endogenic activity. The critical moment in geological history was at the third stage onset 2.05 Ga ago, when the Archean style of evolution gave way to subsequent one. Magmas of siliceous high-Mg (boninite-like) series (SHMS) dominant during the first stage (2.5–2.3 Ga) had the mixed mantle–crust genesis. They originated in highly depleted ultramafic reservoir of asthenospheric mantle being contaminated by the Archean crustal material during ascent to the surface. Typical of the SHMS magmas were the high content of silica, Al, Mg, Cr and elevated concentrations of Ni, Co, Cu, V, PGE and incompatible elements, LREE included, while concentrations of Fe, Ti, Nb and alkalies, especially K, were at a relatively low level. Magmatism of K-granitoid type was of a limited significance. The second stage, especially its second substage (2.3–2.05 Ga), was distinct because of mass eruptions of picritic and basaltic magmas enriched in Fe, Ti, Mn, P and incompatible elements, LREE in particular, which also had elevated Cr, Ni, Co, Cu and Ba concentration being relatively depleted in Mg and Al. This change in geochemistry of magmatism was independent of tectonic processes, which retained the former style. The third stage (2.05–1.8 Ga) was marked by opening of first oceans (Jormua, Purtuniq and other ophiolites) and by formation of orogens comparable with Phanerozoic orogens that was associated with development of subduction zones and back-arc basins with relevant magmatism. High Mg, Fe, Ti, Cr, Ni, Cu, V but low Th concentrations were typical of intense picrite-basaltic volcanism marking onset of the third stage. Magmas of suprasubduction genesis had considerably higher Si, Al, P, Zn, Th concentrations and very low CaO/Al_2O_3 ratios. Development of orogens came to the end 1.82–1.80 Ga ago. Appearance of giant belts of intraplate, usually silicic high-K volcanism juxtaposed on stabilized Paleoproterozoic orogens with abnormally thick crust was confined to the fourth stage spanning besides the initial Mesoproterozoic (1.8–1.5 Ga). Anorthosite–rapakivi granitoid batholiths, which originated above mantle plume heads, corresponded to intermediate magma chambers of relevant magmatic systems. As is suggested, juvenile basaltic melts of this stage retained within the crust provoking a large-scale melting of sialic material. The high K, Ti, Zn, Pb, Zr and elevated Be, Sn, Y, Nb, Rb, F, W, Mo, Li and U contents characterized geochemistry of magmas, which originated at the fourth stage

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INTRODUCTION

The Paleoproterozoic (2.5–1.6 Ga) is one of very important though inadequately studied eras in the Earth history. It was a time of cardinal changes in biosphere with evolutionary innovations in the organic world (Knoll, 1996; Semikhatov et al., 1999, and references therein). We also share the opinion of other researchers that tectono-magmatic processes changed radically about 2 Ga ago (Bogatikov et al., 2000). One more distinctive feature of Paleoproterozoic Earth is evident from available though fragmentary Sr-isotope data. Almost throughout that era, i.e., from 2.5 to1.85 Ga, the $^{87}Sr/^{86}Sr$ ratio in seawater remained at a low level slightly exceeding values calculated for the concurrent upper mantle. An initial weak tendency of that ratio rising turned into the obvious one only 1.85–1.65 Ga ago (Mirota and Veizer, 1994; Veizer et al., 1976, 1990, 1992a, 1992b; Gorokhov et al., 1998; Whittaker et al., 1998; Kuznetsov et al., 2003). Consequently, the Earth mantle and products of its magmatic differentiation, i.e., the juvenile volcanic and intrusive complexes with associated hydrothermal systems, represented main sources of chemical elements and their compounds, which are important for metabolic and fermentative activity of Paleoproterozoic biota (Fedonkin, 2003).

The insight into Paleoproterozoic magmatism evolution is therefore important for understanding processes in the early biosphere. In this work based on analysis of worldwide data, we discriminate four successive stages of Paleoproterozoic magmatism, each

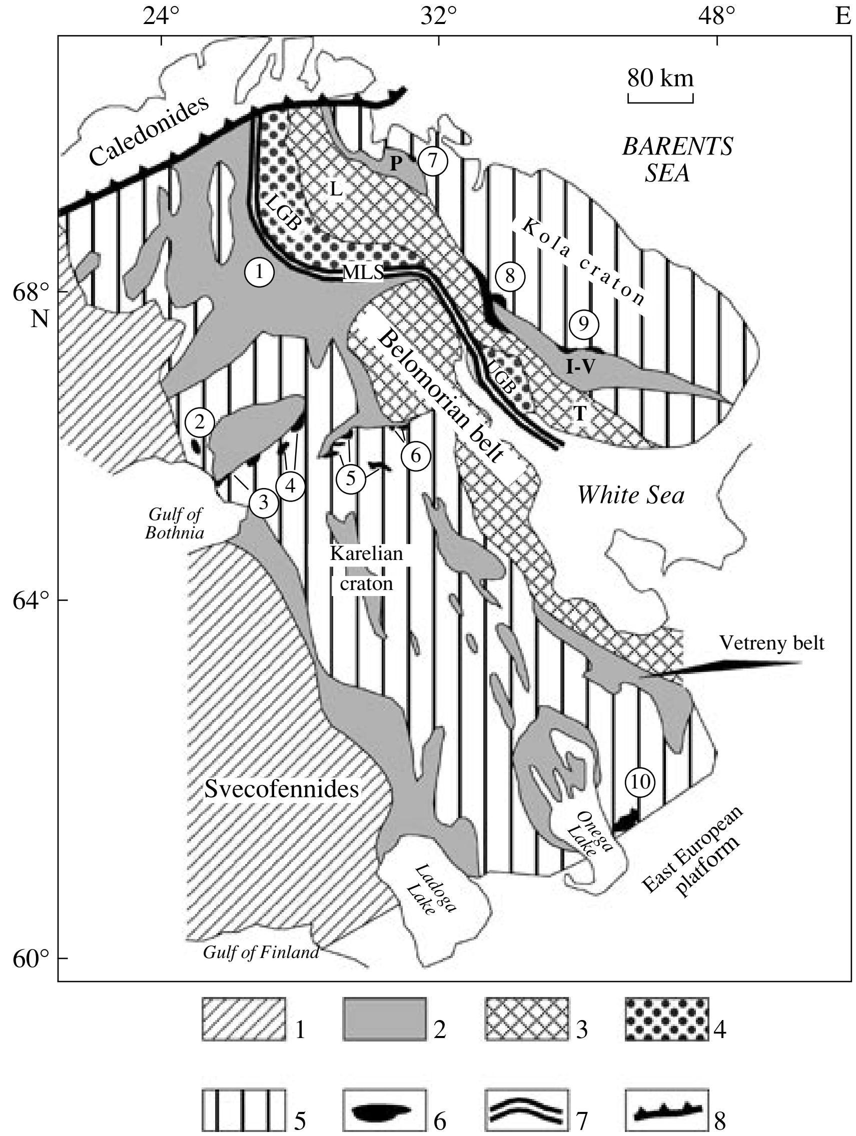

Fig. 1. Geological scheme of the Baltic Shield (after Sharkov et al., 1997): (1) Svecofennides; (2) Pechenga (P) and Imandra–Varzuga (I-V) Paleoproterozoic complexes of volcanogenic-sedimentary rocks; (3) Lotta (L), Tersk (T) segments of the Tersk–Lotta mobile belt; (4) Lapland (LGB) and Umba (UGB) granulite belts; (5) Archean basement; (6) layered intrusions, encircled numbers denote Koitilainen (*1*), Tornio (*2*), Kemi (*3*), Penikat (*4*) Koilismaa (*5*) Olanga (*6*), Mt. General'skaya (*7*), Monchetundra (*8*), Fedorova-Pana (*9*) and Burakovo (*10*) plutons; (7) Main Lapland Suture (MLS); (8) northern boundary of the Baltic Shield.

representing its own type of magmatic activity with peculiar geochemical characteristics. The analysis is based on data on the Baltic and Canadian shields, which are most completely studied in geological and petrological aspects, and where composition of rocks is established by modern analytical methods. Data on the other shields are used for comparison to show global manifestation of the processes under consideration.

MAGMATIC ROCKS OF THE FIRST STAGE (2.5–2.3 Ga)

By the onset of Paleoproterozoic time, the Earth's crust was stabilized to a considerable extent. This is evident from worldwide development at that time of graben-like rift structures filled in with volcanogenic-sedimentary rocks and from emplacement of giant dike swarms and large layered plutons.

Three main structural provinces of the Early Paleoproterozoic are distinguishable in the eastern part of the *Baltic Shield* (Fig. 1). These are (1) the Kola and Karelian cratons, the large areas of crustal upwarping and extension with widespread intraplate mafic–ultramafic magmatism; (2) the Lapland–Umba granulite belt (LUGB) between two cratons that corresponds to a zone of compression and downwarping with intense enderbite-charnockite-granite magmatism of crustal origin at the site of large sedimentation basin; (3) the

Belomorian and Tersk–Lotta belts or intervening mobile zones of slightly sloping tectonic flows of crustal material from the cratons toward the LUGB.

Magmatic rocks of the first stage belong predominantly to the peculiar siliceous high-Mg (boninite-like) series (SHMS). Being widespread in the shield eastern part, they form here the Baltic igneous province (Sharkov et al., 1997) that originated in the course of the Sumian (2.5–2.4 Ga) and Sariolian (2.4–2.3 Ga) tectono-magmatic cycles of comparable tectonic and magmatic events. Volcanic belts of the province are confined to linear graben-shaped structures, being concurrent to dike swarms of gabbronorites, large layered mafic–ultramafic intrusions of the Kola and Karelian cratons (Monchegorsk, Burakovo and other plutons) and to small synkinematic intrusions of the same composition in the Belomorian and Tersk–Lotta mobile belts (Fig. 1). As is known, there are the Late Archean boninite-like rocks in some greenstone belts of the Baltic Shield (Shchipansky et al., 1999, 2001), but magmas of this kind became dominant only in the Early Paleoproterozoic.

The SHMS magmas show (1) deficiency of Fe, Ti, Nb and alkalies, especially of potassium; (2) elevated concentrations of Ni, Co, Cu, V, PGE and incompatible elements, LREE inclusive, along with low REE fractionation in general: $(La/Yb)_N = 3.1–5.7$; and (3) a high content of Al, Mg, and Cr. In petrochemical diagrams, data points characterizing these magmas plot in the field of boninites, the high-Mg rocks of island arcs. Being close to the latter in bulk chemical composition and concentrations of trace and rare-earth elements, the SHMS rocks reveal distinct isotopic parameters however: in boninites of present-day island arcs, εNd(T) ranges from +6 to +8 (Pearce et al., 1992), whereas this parameter in the SHMS rocks varies from +1.0 to –6.0 in general and commonly from –1.0 to –2.0 (Sharkov and Smolkin, 1997; Hanski et al., 2001a). The $^{87}Sr/^{86}Sr$ initial ratio in the rocks under consideration is usually within the range of 0.7032–0.7044 (Amelin and Semenov, 1996a). The geochemical characteristics of the SHMS rocks used to be considered as the result of contamination of mantle-derived magmas by the Archean crustal material during their ascent to the surface (Puchtel et al., 1996; Sharkov et al., 1997).

The Sumian–Sariolian volcanogenic-sedimentary deposits, which are 1.6 to 5.5 km thick in the Kola and Karelian cratons, are confined to lower parts of successions in rift structures, e.g., in the Pechenga–Varzuga, East Karelian, Vetreny, Central Karelian, Salla, Perapohja, Lapland, and other belts. They discordantly overlie rocks of consolidated Archean basement. Prevailing among volcanics are lava flows, lava and tuff breccias widely ranging in composition from low-Ti picrites and prevailing basalt to andesites, dacites, and rhyolites, while siliciclastic sediments are of subordinate abundance. Volcanic rocks highly altered in general are preserved surprisingly well in places. For instance, fresh mafic–ultramafic lavas with original microspinifex texture and unaltered magmatic minerals (olivine, pyroxene, chromite) and volcanic glass of intermediate to silicic composition are known from the Vetreny belt (Kulikov, 1988; Sharkov et al., 2004a). The Sm–Nd isochron age of Sumian volcanics from this belt is 2.45–2.41 Ga (Puchtel et al., 1996, 1997). Characteristic of these rocks are MgO contents ranging from 7 wt % in fine-grained varieties to 26 wt % in cumulates and very high concentrations of siderophile elements: Cr (450–3000 ppm), V (130–270 ppm), Co (34–107 ppm), Ni (60–110 ppm). The primitive bulk composition of rocks is contrasting with their enrichment in LILE and LREE, negative Nb and Ti anomalies detectable in spidergrams, and with negative εNd(T) values from –1.3 to –0.71 (Puchtel et al., 1996). Age of felsic volcanics from a southern area of the Paana–Kuolajarvi structure corresponds to 2434 ± 24 Ma (Turchenko et al., 1991).

The Paleoproterozoic Pechenga–Varzuga belt, a largest one in the Kola craton, comprises the Pechenga and Imandra–Varzuga rift structures separated by the Archean basement ledge (Fig. 1). Like in Karelia, volcanogenic-sedimentary complexes of both structures are composed in the basal interval of alternating conglomerates and sandstones resting on weathering crust of the Archean basement, and amphibolites after basalts and basaltic andesites are main rock types in the higher interval (Smolkin et al., 1995). Dacites, rhyolites and low-Ti picrites are subordinate rocks of the complex. The U–Pb zircon age (simply the age further on, if the other dating method is not quoted) of the rocks is 2442 ± 1.7 Ma (Amelin and Heaman, 1995).

Close in age norite and gabbronorite dike swarms and large layered mafic–ultramafic intrusions of the SHMS are emplaced along periphery of the Sumian–Sariolian volcanic fields (Alapieti et al., 1990; Sharkov and Smolkin, 1998). The differentiated intrusive series is represented by dunite, harzburgite, pyroxenite, norite, gabbronorite and anorthosite; late differentiates correspond in composition to V-bearing magnetite gabbronorite, gabbrodiorite, and granophyre. Judging from ore mineralization in plutons, the relevant magmas carried Cr, Ni, Cu, V, and PGE to the surface. Large anorthosite massifs that originated at the same time in the Kola craton are close in geochemical and isotopic parameters to gabbroids of differentiated intrusions.

In the Belomorian and Tersk–Lotta intervening mobile zones, igneous mafic–ultramafic rocks of the first stage occur in dispersed numerous small bodies of synkinematic intrusions occupying areas not greater than 0.5 to 2 km^2. These intrusive bodies attributed to the coronite complex of the Belomorian belt were emplaced 2.46–2.35 Ga ago (Sharkov et al., 2004b). In the bulk composition and geochemical characteristics, rocks of this complex are close to rocks of layered intrusions in nearby cratons, being interrelated with the SHMS magmas. Their extrusive equivalents are represented by metavolcanics of the Kandalaksha Formation

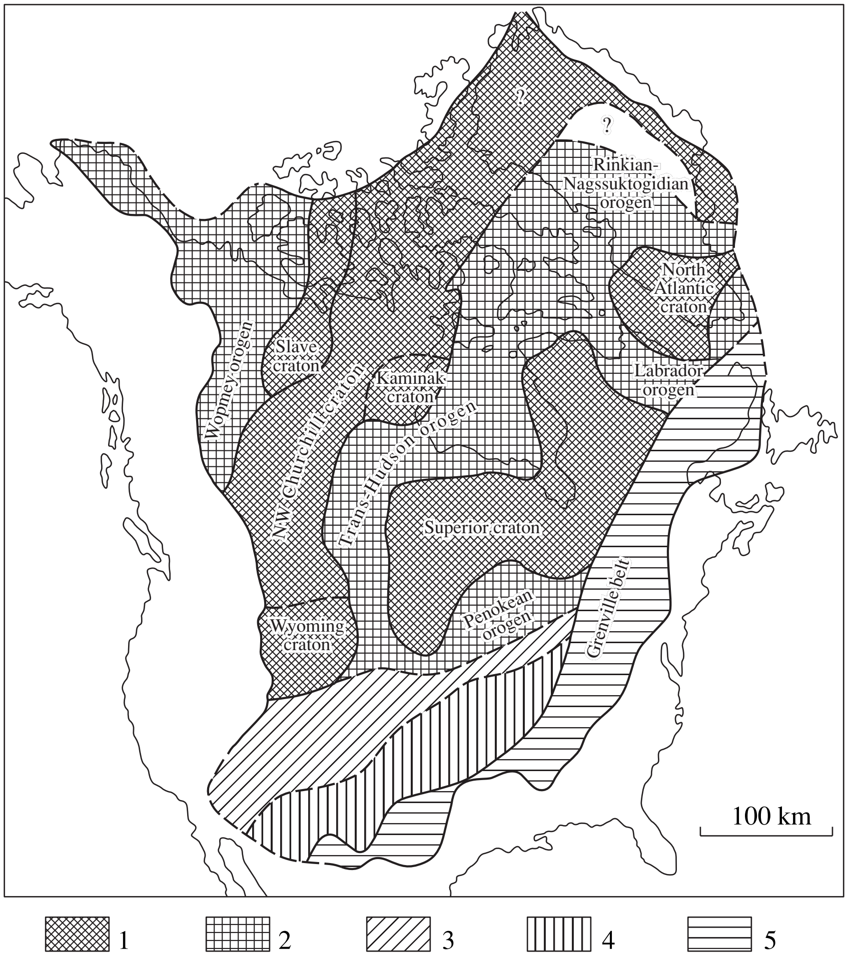

Fig. 2. Principal domains of Canadian Shield (after P. Hoffman from Khain and Bozhko, 1988) and stabilization time of crust: (1) > 2.5 Ga; (2) 1.9–1.8 Ga; (3) 1.8–1.7 Ga; (4) 1.7–1.6 Ga; (5) 1.2–1.0 Ga.

in the study region and of the Tana Formation in Finland, which are "clamed" between the Late Paleoproterozoic Main Lapland Suture and the Belomorian belt.

Manifestations of the Early Paleoproterozoic (Sumian) K-granitoid magmatism are of a very limited distribution and range in age from 2.45 to 2.40 Ga (Bogdanova and Bibikova, 1993; Levchenkov et al., 1994; Lauri, 2002). Massifs of these granitoids situated along the contact of the East Karelian zone with the Belomorian belt are studied better than in other places. Their rocks extremely depleted in MgO, CaO and TiO_2 show moderate to high concentration of Ba, Zr, Zn, Ga and Nb, low concentrations of Sc, V, Ni and Co, negative ("crustal") εNd(T) values ranging from –2.0 to –4.8 (Lauri et al., 2002). Granitoid intrusions of this kind and eruptions of comagmatic rhyolitic lavas were of a very limited significance in the Early Paleoproterozoic magmatism of the Baltic Shield.

In the *Canadian Shield*, areas of the Archean consolidation correspond to the Superior, Churchill, Slave, Wyoming, North Atlantic or Nain and other cratons separated by younger foldbelts (Fig. 2) by the end of Paleoproterozoic at the time of Hudsonian orogeny (Hoffman, 1988). The SHMS volcanics are confined to lower intervals of volcanic-sedimentary successions in graben-like structures of cratons, and concurrent layered mafic–ultramafic intrusions are emplaced into neighboring basement blocks. For instance, the Huronian Supergroup in the south of the Superior craton contains the SHMS volcanics in its lower Thessalon Formation (Jolly, 1987), and concurrent layered mafic-ultramafic intrusions of the East Bull Lake Suite, which are comparable with layered plutons of Fennoscandia, are confined to the same area (Vogel et al., 1998). Age of the Huronian volcanics is accepted to be 2450 ± 25 Ma based on correlation with the dated Copper Cliff rhyolite of the Sudbury district, an analogue of the Thessalon rhyolites. The upper age limit of the supergroup corresponds to 2219.4 ± 4 Ma, as it is inferred from dating the Nipissing sills that intruded the Huronian rocks (Corfu and Andrew, 1986).

Basalts of the supergroup with moderate Ti, Nb and Mg concentrations are relatively enriched in Cr, Cu, Ni, V, Zn, LREE and reveal negative Nb and Ti anomalies, whereas rhyolites having high Zr concentration are enriched in LREE to a sensible or greater extent. In the western Hearne domain of the Churchill craton, volcanic rocks of the stage under consideration are represented by basalts from an upper part of the Hurwitz Group that discordantly overlies the 2.45 Ga Kaminak dikes (Heaman, 1994). These basalts also enriched in LREE show negative Nb, Ti, Ta anomalies and εNd(T) values from 0 to +0.8 (Sandeman, 2003). In geochemical characteristics (relatively high Mg, Cr, Co concentrations and negative Nb, Ta, P, Ti anomalies), the Kaminak dike swarm is close to the SHMS. Parameter εNd(T) in the dike rocks varies from 0 to –1.27, suggesting a weak contamination of parental magmas.

The giant Hearst–Matachewan dike swarm 2.47 to 2.45 Ga old (Heaman, 1997) occupies in the Superior craton an area 800 × 450 km. As is assumed, the dikes represent feeders of volcanic succession of a large igneous province eliminated by erosion, and their rocks originated from magmas, which experienced differentiation in intermediate magma chambers to become enriched in Fe and Ti (Phinney and Halls, 2001). The ~2.45 Ga dike swarms resembling boninite in composition are known also in West Greenland (Hall and Hughes, 1990; Heaman, 1996), from where they are traceable to Scotland (Scourie dikes of ultramafic and gabbronorite composition) and westward to the Wyoming craton (Heaman, 1996). Like in the Baltic Shield, magmas of boninitic affinity appeared in the Canadian Shield about 2.7 Ga ago in the Late Archean. The Stillwater massif, a large pluton of layered mafic–ultramafic rocks comparable in structure and composition with the Monchegorsk complex, exemplifies this magmatism in the Wyoming craton (McCallum, 1996); the ~2.7 Ga Kidd Munro and Tisdale volcanics in the Abitibi belt (Kerrich, 1998).

Rocks of the Early Paleoproterozoic SHMS are known in all other Precambrian shields, but in the Southern Hemisphere, they occur frequently as subordinate members of volcanic series dominated by rhyolites. For example, the Woongara Supergroup of the Hammersley basin in the south of the Archean Pilbara craton, West Australia, comprises succession of dominant rhyolitic lavas and sills enclosing insignificant amount of doleritic and basaltic bodies of the SHMS intercalated with banded iron formation. Rhyolites are 2449 ± 3 Ma old (Barley et al., 1997). Rocks of basic composition reasonably enriched in incompatible elements and lacking distinct Ti anomalies show negative εNd(T) values and have almost undifferentiated REE spectra. Rhyolites relatively enriched in HFSE and LREE are characterized by negative Nb, Ta, Ti and Eu anomalies (Barley et al., 1997). In the Yilgarn craton of West Australia, there are known the Widgiemooltha norite and gabbro dike swarm of the SHMS and the Jimberlana intrusion of mafic–ultramafic rocks, which are 2.42 Ga old (Fletcher et al., 1987). The Westfold Hill dikes of East Antarctic are composed of the ~2.4 Ga boninite-like rocks (Kuehner, 1989). In South Africa (Zimbabwe), the rock series under consideration is represented by the famous Great Dike approximately 500 km long; this huge differentiated intrusion of mafic–ultramafic rocks with platinum mineralization is 2586 ± 16 Ma old according to recent dating results (Mukasa et al., 1998). Volcanics of the SHMS are also widespread in the Indian Shield, where they occur in the lower part of the Paleoproterozoic Dongargarh Supergroup (Sensarma et al., 2002), being represented mostly by subalkaline to calc-alkaline rhyolites associated with subordinate dacites, andesites and basalts. Rhyolites are enriched in LREE, Zr, Th, Cr but depleted in LILE, Ti, Sc, Y and Ni. The REE spectra are LREE-enriched in rhyolites, less fractionated in dacites, and flattened in basalts. Basaltic rocks displaying negative Nb and Ti anomalies are enriched in LILE and depleted in HFSE (Neogi et al., 1996). The SHMS of the Ukrainian Shield is likely represented by metaporphyrites (dacitic andesites and rhyolites) of the Novograd–Volhyn Group and by basaltic to andesitic lavas of the Teterev Group, all being 2.4 Ga old (Shcherbak et al., 1989).

Thus, the Early Paleoproterozoic SHMS magmatism dominated in all the ancient shields, being responsible for volcanic activity in rift structures and for origin of dike swarms and large layered mafic–ultramafic intrusions within large igneous provinces. Granitoid magmatism was of subordinate significance at that time.

MAGMATISM OF THE SECOND STAGE (2.3–2.05 Ga)

Characteristic of the second stage was the first mass appearance on the Earth of Fe-Ti picrites and basalts with normal to high content of alkalies, which became typical of the Phanerozoic intraplate magmatism. Simultaneously, it was a time of gradual cessation of the SHMS magmatism. Transition to the new type of magmatic activity was asynchronous in different cratons. In some cratons, the SHMS magmatism gradually dying out was still in action, whereas vast areas of other shields represented arenas for continental rifting-related magmatism of the Phanerozoic type. Changes in the character of magmatism had not been related however with transformation of tectonic processes, which functioned as they did before.

In the *Baltic Shield*, magmatism of the second stage is exemplified by volcanics and sills occurring in the Jatulian and equivalent successions of linear rift-like structures and separate broader synforms and grabens, where these successions discordantly overlie either the Sumian-Sariolian rocks or the Archean crystalline basement (Golubev and Svetov, 1983; *Geology of Karelia*, 1987; Smolkin et al., 1995). Geochronological dates obtained in the Karelian craton for the Jatulian rocks are not numerous: metavolcanic rocks near the Jatulian base yielded the Pb–Pb date of 2331 ± 98 Ma; dates obtained for the layered sills in the Jatulian quartzites fluctuate around 2150 Ma; the Sm–Nd age of basalts in the succession upper part is 2090 ± 70 Ma, and the U–Pb zircon age of Upper Jatulian silicic lavas corresponds to 2062 ± 2 Ma (Laajoki, 1980; Huhma, 1990; Pekkarinen and Lukkarinen, 1991). Among volcanic rocks, there are normal and Fe-Ti tholeiitic, subalkaline and picritic basalts containing up to 3 wt % TiO_2 and confined usually to the Upper Jatulian horizons. Owing to subaerial character of eruptions, lavas are often intercalated with tuff and tuff-breccia horizons.

In the Pechenga–Varzuga belt of the Kola craton, equivalents of the Jatulian are represented by the Kuetsjarvi and Umba groups of volcanogenic-sedimentary rocks, and in the lower part by the Kolasjok Group mostly composed of volcanics (Smolkin et al., 1995). Rocks prevailing in upper horizons of two former correlative groups are subalkaline basalts (mugearites) frequently enriched in titanium, basaltic trachyandesites, dacites, and scarce picrites. Volcanics of the Kuetsjarvi Group are 2214 ± 54 Ma old (Rb–Sr dating), and comagmatic dikes and sills are dated at 2200 Ma (Smolkin et al., 1995). The Kolasjok Group (2114 ± 52 Ma) of the Pechenga structure comprises pillow lavas corresponding in composition mostly to tholeiitic and sometimes to picritic basalts. In the northwest of the Kola craton, volcanics of the Jatulian age are represented by Fe-Ti picrite-basaltic series of the Karasjok belt (Krill et al., 1985) and by high-Ti mafic–ultramafic lavas of the Lapland greenstone belt (Hanski et al., 2001b). The Sm–Nd dates on these volcanics are 2.27–2.09 and 2.06 Ga, respectively. Tholeiitic and Fe-rich tholeiitic diabases are widespread outside grabens in Karelia, East and North Finland, being as old as 2.2–2.1 Ga (Vuollo et al., 1995). In distinction from tholeiitic diabases with modestly high concentration level of compatible elements, which are not enriched in TiO_2 and HFSE, rocks of the second type with the same concentrations of compatible elements are enriched in TiO_2, FeO, and HFSE (Stepanova, 2005).

Thus, a vast area of the Jatulian igneous rocks in the east of the Baltic Shield was arena of basaltic magmatism, and relevant magmas relatively depleted in Mg and Al had variably high relative concentrations of many other elements: Fe, Ti, Mn, P, alkalies, LREE, Cr, Ni, Co, Cu, and Ba. Parameter εNd(T) ranging from –2.83 in subalkaline volcanics of the Kuetsjarvi Group to +5 in volcanic rocks of the Karasjok Group suggests the juvenile origin of parental magmas in the last case and a considerable contamination by crustal components in the former one (Smolkin et al., 1995; Krill et al., 1985).

In the *Canadian Shield*, equivalents of the Jatulian volcanics are confidently established in the Labrador trough only, where magmas corresponding in composition to alkaline basalts erupted 2.17–2.14 Ga ago (Wardle and van Kranendonk, 1996). In the same shield, there are widespread dike swarms and diabase sills of comparable age, the oldest ones being 2.23–2.21 Ga old; these are the Nipissing sills of the Southern Province, the Maguire, Senneterre and Klotz dikes in the south of Superior craton, and the Malley-McKay dikes of the Slave craton (Vogel et al., 1998). Their distinctive features are the low Ti but high Mg content and undifferentiated REE spectra. The younger dike swarms (2.17–2.14 Ga) are known in the Kapuskaising belt of the Superior Province and in the Griffin belt of the western Churchill Province (Boily and Ludden, 1991). The high-Ti, slightly alkaline rocks of the Griffin belt are enriched in Nb and La, being depleted in HREE; εNd (T) varies from +1.0 to –0.1. Magmatic rocks of similar composition and age are known in the other regions as well, for instance in the Transbaikal region. The Paleoproterozoic Udokan Group of volcanogenic-sedimentary rocks is intruded here by the Chinei pluton of high-Ti mafic rocks dated at about 2.2 Ga (Konnikov, 1986) and by dike swarms of Ti-picrites, which are 2202 ± 41 Ma old (Sm–Nd dating method); εNd(T) = +1.6 (Puchtel and Zhuravlev, 1992).

The SHMS magmatism still developed in local domains of certain shields. In the South African Shield, for instance, magmatic rocks of the stage under consideration are represented by the SHMS lavas and pyroclastic horizons of basaltic-andesite composition in the 2223 ± 13 Ma Pretoria Group (Oberholzer and Eriksson, 2000). The famous Bushveld Complex over 29000 km^2 in area, which is composed of platinum-bearing mafic–ultramafic rocks, intruded here ca. 2.06 Ga ago the lower sedimentary succession of the Rooiberg Group (2061 ± 2 Ma) that comprises the SHMS volcanics as well. As is assumed, plutonic rocks crystallized in a magma chamber that existed at the formation time of the group upper part (Oberholzer and Eriksson, 2000), and based on their compositional similarity with comagmatic volcanics, Buchanan et al. (1999) distinguished the Bushveld magmatic province of the shield. Similar situation is recognizable in other regions, e.g., in the Voronezh crystalline massif, where the SHMS magmatic complexes (2.1 to 2.05 Ga) are represented by large layered intrusions bearing PGE–Cu–Ni mineralization and by dike swarms and subvolcanic sills (Chernyshev, 2004). In the Archean block of southwestern Greenland, there are two ~ 2.1 Ga dike swarms (Halls and Hughes, 1990), one of which is close in composition to tholeiitic dolerite, while the other one composed of norites and gabbronorites is of the SHMS affinity. The 2241 Ma norite dikes are known as well in the West Antarctic Shield.

Thus, the second stage of Paleoproterozoic magmatism produced the widespread Fe-Ti picrites and normal to subalkaline and alkaline basalts, which are characteristic of the Phanerozoic intraplate magmatism. Gradual cessation of the SHMS magmatism, which was manifested in local areas only, was another interesting feature of this stage. Despite the obvious change in composition of magmatic melts, there was no cardinal transformation in tectonic regimes: as before, the rifting regime was dominant, and volcanism was confined to structures inherited from the Early Paleoproterozoic ones, where its products continue upward the former successions.

Rocks of the Birrimian Group (2.21–2.10 Ga) in West Africa are of particular interest. They are known in several greenstone belts between Cote d'Ivoire and the Ahaggar massif. Dominant in the belts are basaltic rocks associated with subordinate intermediate and silicic volcanics, and plutonic facies are represented by small mafic–ultramafic intrusive bodies and granitoid intrusions (Hirdes et al., 1996; Beziat et al., 2000). In distinction from most concurrent groups of the other shields, rocks of the Birrimian Group have geochemical characteristics of island-arc magmatism: enrichment in LILE and PB, depletion in HFSE, negative Nb and Ti anomalies in spidergrams, high Ce/Nb and Th/Nb ratios, and εNd(T) = +3.0 ± 1. The oldest rocks of the group (Pb–Pb zircon age 2211 ± 3 Ma) are established in the Niani Group of lavas and tuffs of dacitic to rhyodacitic composition; some of the rocks are of adakitic affinity that may be indicative of their origin in suprasubduction settings (Lahondere et al., 2002). In addition, relicts of oceanic plateaus are distinguished in some belts based on geochemical data (Sylvester and Attoh, 1992). It is possible that here we observe for the first time the events typomorphic of the third stage of Paleoproterozoic magmatism.

MAGMATISM OF THE THIRD STAGE (2.05–1.8 Ga)

Formation of first mobile belts and orogens of Phanerozoic type with ophiolite suites satisfying definition of the Penrose Conference (a succession of mantle ultramafic, gabbroid, sheeted dike complexes crowned by tholeiitic pillow lavas) commenced ca. 2.05 Ga ago. It is likely that at this time exactly the Earth went into new period of its development lasting until present (Bogatikov et al., 2000).

Bright examples of magmatic complexes, which originated at the third stage, are known in the *Baltic Shield*. In particular, this is the Ludicovian Group composed predominantly of picritic basalts and subaerial pyroclastic deposits, which are 1.7 km thick in total and discordantly overlap older volcanics in the rift structures of Karelia (*Geology of Karelia*, 1987). Concurrent neck facies, dikes and sills are widespread in proximity. The Konchozero sill is 1975 ± 24 Ma old according to the known Sm–Nd date, εNd(T) = +3.2 ± 0.1 (Kulikov et al., 1999), that is close to the published Re–Os age value of 1969 ± 18 Ma (Puchtel et al., 1999). A comparable age (1965 ± 10 Ma) is established for tholeiitic diabase dikes crosscutting the Jatulian succession in eastern Finland (Vuollo et al., 1995). Many of the rocks in question are enriched in Fe, Ti, alkalies and LREE. The Pilgujarvi Group of the Pechenga–Varzuga belt (PVB) is an age analogue of the Ludicovian in Karelia (Smolkin et al., 1995). Its lower formation is composed of turbidites (~1 km), and the upper one (1.9–2.0 km) comprises tholeiitic and scarcer subalkaline pillow lavas, tuffs, hyaloclastites, and lenticular interlayers of black shales and tuffaceous siliciliths. This rock succession is intruded by gabbro-diabase sills now metamorphosed and by coarsely layered mafic–ultramafic plutons with Cu–Ni sulfide mineralization of economic value. In composition, volcanics of the Pilgujarvi Group are of two types (Smolkin, 1992). In majority, they are close to mid-ocean ridge basalts (MORB) with their characteristic REE patterns, while the other rocks correspond in composition to Fe–Ti picritic and subalkaline basalts distinctly enriched in LREE, being therefore comparable with Phanerozoic volcanics of the intraplate settings, i.e., with basalts of oceanic islands (OIB) and continental rifts. The same REE patterns are established in chilled margins of plutons bearing Ni mineralization and representing therefore intrusive equivalents of aforementioned ferropicrites. Besides, the group of Fe-enriched rocks differs from associated tholeiitic basalts by higher concentrations of incompatible elements (Zr, Ba, Sr, U, Th, P, F), and parental magmas of two rock types originated therefore in separate magmatic reservoirs as in the present-day oceans and rift structures of the Red Sea type (Wilson, 1989).

According to numerous U–Pb, Pb–Pb, Sm–Nd and Os–Ir dates, ferropicrites and intrusions with Ni mineralization of the Pechenga province are 1990 to 1970 Ma old (Smolkin, 1992). The $^{87}Sr/^{86}Sr$ initial ratio in these rocks corresponds to 0.70303 ± 0.00027; εNd(T) = +1.6 ± 0.4. The ledge of Archean basement between the Pechenga and Imandra–Varzuga structures is intruded by the Gremyakha-Vyrmes differentiated mafic–ultramafic massif (ca. 100 km^2 in area) with Ti mineralization and central core composed of nepheline syenite. In geochemical characteristics, this massif is close to the Pechenga intrusions, though being differentiated to a much higher extent. The Sm–Nd isochron age of the massif is 1926 ± 74 Ma; εNd = +0.8 (Savatenkov et al., 1998).

The $^{207}Pb/^{204}Pb$ initial ratio in the considered rocks is relatively low as compared to the average global value calculated for the geological time mark of 2.0 Ga. This means that ferropicritic magmas have not been contaminated by radiogenic crustal Pb at that time, and parameter εNd(T) = +1.6 characterizes isotopic composition of mantle reservoir enriched in LREE shortly before the magma generation. The calculated model age of the depleted mantle enrichment is 2.2 Ga (Hanski et al., 2001b), and this event was likely a result of mantle metasomatism, as it is evident from relatively high percentages of TiO_2, P_2O_5, and F in the rocks. Consequently, first enrichment of ancient depleted mantle under the Kola craton took place ca. 2.2 Ga ago, when first alkaline Fe–Ti basalts appeared in the Kuetsjarvi Group.

Thus, geological and petrological data imply that evolution of supracrustal complexes of the PVB was characterized by transition from the continental (Kuetsjarvi Group) to oceanic (Kolasjok and Pilgujarvi groups) rifting. Intense underwater eruptions of undifferentiated basalts and deposition of abyssal black shales and turbidites apparently indicate that a relatively deep uncompensated basin existed at the site of the PVB. The basin formation could be related either to transition from the continental rifting to oceanic spreading, or to development of the back-arc spreading behind the Main Lapland Suture with the MORB-type magmatism in the basin and calc-alkaline volcanism along its periphery (Sharkov and Smolkin, 1997). Regardless of the PVB interpretation, this structure corresponded in general to a zone of new oceanic crust formation in the Paleoproterozoic time.

In the eastern Baltic Shield, subalkaline Fe–Ti picrites and basalts with associated diabase dikes ranging in age from 2.05 to 1.90 Ga are widespread in the Kola region, North Karelia and Lapland (*Mafic dike*…, 1989; Smolkin et al., 1995; Vuollo et al., 1995). They exemplify numerous volcanic conduits under rift structures, in particular under the PVB. Data on core section SG-3 recovered by superdeep drilling in the Pechenga structure show that rocks of Proterozoic dikes represent no less than 12–15% in its Archean basement (Vetrin et al., 2003). In addition, there are dikes of Ti-rich hornblende peridotite and olivine gabbro, which are dated at 1941 ± 3 Ma in the Kola Peninsula (Smolkin et al., 2003).

Opening of the Svecofennian oceanic basin in the Baltic Shield central zone commenced ca. 2.0 Ga ago (Gaal and Gorbatschev, 1987) almost concurrently with the considered events. The succession of Svecofennides includes turbidites and tectonic sheets of typical ophiolites – the Jormua and Outokompu complexes of northeastern Finland, which are as old as 1953 ± 2 Ma (Kontinen, 1987; Peltonen et al., 1996). The Re–Os isotopic systematics of serpentinites and chromites from the Jormua Complex showed broad γ_{Os} variations from –5.1 to +3, and formation of ophiolites involved, consequently, the ancient subcontinental mantle mixed with younger MORB-type (chondritic) reservoir at the time of rifting (Tsuru et al., 2000). After the rifting in Archean craton, obduction of the Jormua Complex took place at the time of collision between the craton and Svecofennides. A present-day analogue of that Paleoproterozoic structure is probably the Red Sea rift.

The Svecofennides occupying almost entirely the western part of the Baltic Shield are represented predominantly by granitoids, volcanic arc complexes, and by migmatized metaturbidites; proportion of volcanics among these rocks is not greater than 10%. All the rocks experienced progressive metamorphism predominantly under conditions of amphibolite facies. The oldest intrusive and volcanic rocks (1.93–1.92 Ga) form a primitive volcanic arc composed of metamorphosed low-K tholeiites, basaltic andesites, low-K rhyolites, and comagmatic tonalites. The Central Finland Granitoid Complex (CFGC) of the Svecofennian orogen is formed by large synkinematic intrusions of 1.89–1.87 Ga I-granites. The Earth crust of the CFGC is very thick, up to 50–60 km. In composition and deep structure, this complex is similar to Mesozoic batholiths of active continental margins, e.g., to the Sierra Nevada batholith of California (Ducea and Salliby, 1998). In the Tampere belt, island-arc volcanics ranging in age from 1904 to 1889 Ma are represented predominantly by andesites and dacites, which overlie the 1.95 Ga pillow lavas and sediments. Southward, the island-arc rocks are replaced by mafic volcanics characteristic of rifting settings. Finally, the South Finland Complex of volcanogenic-sedimentary rocks consists of migmatized turbidites and island-arc volcanics corresponding in age to 1.89–1.88 Ga. Cratonization of the orogen that culminated ~ 1.80 Ga ago resulted in formation of contemporaneous Svecofennian geoblock (domain).

To accomplish the analysis of tectono-magmatic processes, which were in progress 2.05–1.80 Ga ago, we should emphasize that they were very different in western and eastern parts of the shield. At the development time of the Svecofennian mobile belt in the west, the eastern part represented arena of basaltic eruptions, which were nevertheless of comparable chemical and isotopic composition. Consequently, magmatism of both regions can be united and attributed to the Ludicovian igneous province of the latest Jatulian and Ludicovian magmatic rocks related in origin to one mantle plume (Kulikov et al., 1999). P. Escola (1963) was first to recognize this province as large as ca. 600000 km^2 in area, comparable in dimensions with Phanerozoic trap provinces.

Intense tectono-magmatic processes of similar scope and character developed at the same time in the *Canadian Shield* (Gower et al., 1990). The Wopmey, Trans-Hudson, Penokean, Labrador, and Ungava (Fig. 2) orogenic belts, which separated here slightly reworked rigid cratons (Hoffman, 1988; Henger et al., 1989), terminated their evolution by 1.85–1.8 Ga (Lewry and Collerson, 1990). In structure, these orogens are very similar to Svecofennides of the Baltic Shield, being composed of migmatized metasedimentary and metavolcanic rocks altered under conditions of medium-grade amphibolite facies. Tholeiitic to subalkaline Fe–Ti basaltic volcanism frequently followed with time by deep-water sedimentation was characteristic of rift structures in cratons. In the northern part of Trans-Hudson orogen, along its contacts with the Archean Superior craton (Circum-Ungava foldbelt), there were formed continental rift structures with fields of tholeiitic and subalkaline Ti-enriched basalts, which are set in the Archean basement. As is assumed, it was continental margin of an oceanic basin, and the 1998 ± 2 Ma Purtuniq ophiolites represent fragments of that basin lithosphere obducted onto craton (Scott et al., 1992). Rifting that resumed 1.92–1.87 Ga ago was responsible for eruptions of tholeiitic basalts in the orogen northern margin and subalkaline and tholeiitic basalts, scarcer silicic volcanics and associated rocks of banded iron formation in the eastern zone.

Approximately at the same time (1.91–1.88 Ga), the Lynn Lake belt representing structural collage of volcanic-arc complex, MORB pillow lavas and OIB, which are 1.9 Ga old, originated in central zone of the Trans-Hudson orogen (Baldwin et al., 1987). Approximately 1.85 Ga ago, these rocks were intruded by granite batholiths of the Andean type, the belt of which originated above subduction zone (Bickford et al., 1990). The Flin Flon belt in the south of the Trans-Hudson orogen also represents a collage of island-arc rocks and basalts of the back-arc basin and oceanic plateau types (Stern et al., 1995) with characteristic underwater association of pillow lavas and rhyolite cupolas. Pillow lavas correspond in composition to the MORB and back-arc basin basalts with juvenile εNd(T) parameters from +3.3 to +5.4. The Fe–Ti basalts occurring here are 1904 to 1901 Ma old, having εNd(T) parameters of +2.2 to +3.4 comparable with those of OIB magmas.

Similar processes took place in the Wopmey orogen that developed 2.1–1.85 Ga ago (Cook et al., 1998) and in the Labrador orogen situated between the Superior and North Atlantic cratons. History of the latter commenced with events of continental rifting converted later on into oceanic spreading with associated deep-water sedimentation and tholeiitic magmatism that was active 1.89–1.87 Ga ago (Wardle and van Kranendonk, 1996). Closure of the basin and termination of the orogen development took place ~ 1.8 Ga ago. Dike swarms of Fe–Ti tholeiites and alkaline basalts, for instance the Kapuskaising dikes, originated in nearby cratons 2.1–1.9 Ga ago (Boily and Ludden, 1991).

Rocks of the Middle Paleoproterozoic stage of magmatism are established in the other regions as well. Tholeiitic and subalkaline Fe–Ti basalts with associated small intrusive bodies and dikes of gabbroids are described in the Indian Shield, where the West Himalayan large igneous province existed 2.0–1.8 Ga ago (Ahmad et al., 1999). The Mt. Bundey province of high-K and calc-alkaline magmatism confined to the Pine Creek ledge of crystalline basement in Northwest Australia originated 1831 ± 6 Ma ago (Sheppard, 1995), and a vast belt (100 × 700 km) of silicic high-K magmatism was formed 1865–1820 Ma ago along southern and eastern margins of the Kimberly craton (Griffin et al., 2000). The Middle Paleoproterozoic succession comparable with that of the Baltic Shield is also established in the Ukrainian Shield, where synorogenic volcanism of calc-alkaline series developed 2.08–1.96 Ga ago in its western part (Esipchuk et al., 2000). In eastern areas near the Sea of Azov, there were formed dike-shaped bodies of phlogopite peridotites and pyroxenites, nepheline syenites and carbonatites, which are 1920 ± 80 Ma old, representing rocks typical of the alkaline K–Na series (Shcherbak et al., 1989). Numerous small intrusive bodies of mafic–ultramafic rocks with Cu–Ni–PGE mineralization are confined here to a zone of deep-seated faults. They range in age from 2.02 to 2.0 Ga, having the $^{87}Sr/^{86}Sr$ initial ratio of 0.7026 and εNd value of +3.6. In addition, there are detected intrusions of gabbro–monzonite–syenite–granite as old as 2.02–1.98 Ga (Esipchuk et al., 2000). Like in the other regions, culmination of orogeny was accompanied here by intense amphibolite metamorphism and migmatization.

Thus, the third stage was marked by opening of oceanic basins and by formation of large Phanerozoic-type orogens almost in all the Precambrian shields. Zones of continental rifting inherited from the previous stage were still under development along periphery of the orogens in marginal areas of rigid Archean cratons. The magmatic activity was locally so intense that resulted in origin of large igneous provinces, for instance, in the east of the Baltic and in the north of the Indian shields. Developing through the time, many rifts turned into structures of the Red Sea type with the MORB magmatism and deep-water sedimentation, as is established in the Pechenga–Varzuga belt of the Baltic Shield and in analogous structures of the Canadian Shield. The stage terminated ca. 1.82–1.80 Ga ago by general stabilization of the Earth crust.

MAGMATISM OF THE FOURTH STAGE (1.8–1.5 Ga)

In all the ancient platforms, the terminal Paleoproterozoic through the initial Mesoproterozoic (1.8–1.5 Ga) was the appearance time of vast intraplate belts with prevailing magmatism of the K-series that became active after stabilization of the Early Paleoproterozoic orogens ca. 1.8 Ga ago and culminated about 1.7–1.6 Ga ago. This magmatism developed predominantly within the Paleoproterozoic orogens of the Baltic, Ukrainian, Greenland, and Canadian shields, along southern margins of the Siberian platform, in the Sino-Korean, African, Indian, Brazilian shields, and in the West Antarctic (Semikhatov, 1974; Rämö and Haapala, 1996). Large unmetamorphosed anorthosite–rapakivi granite complexes (ARGC) are associated with these belts.

The multiphase batholiths of the ARGC occupy areas up to dozen thousands square kilometers, and many of them were formed during 20–30 m.y., as one can judge from the isotopic dates obtained (Rämö and Haapala, 1996). The intrusive phases of batholiths range in composition from mafic (ultramafic sometimes) rocks to granites associated with scarce rocks of intermediate composition (monzonite, quartz monzonite, diorite, etc.) The ARGC are actually the bimodal rock associations dominated by granitoids and deprived of mafic rocks, especially gabbro-anorthosites in certain cases. Closely associated with batholiths are predominantly silicic and scarcer mafic volcanic rocks, and siliciclastic red beds filling in the roof pendants of granite intrusions and fringing the latter.

The *Baltic and Ukrainian shields* are classical structures of the ARGC development (Sharkov, 1999). The submeridional belt of relevant large intrusions is localized within the Svecofennian domain spanning the entire western part of the East European craton; it is over 2000 km long and about 900 km wide, penetrating occasionally into some marginal areas of the craton (Fig. 3). In the domain, there are numerous relicts of abnormally thick crust, as thick as 50–60 km at present and 70–80 km thick at the time of the ARGC formation (Sharkov, 1999). The crust was formed during the Svecofennian orogeny, and its fragments are traceable from the Baltic Shield, via the Russian plate basement, almost to the Black Sea. The ARGC formation was accompanied by emplacement of diabase and quartz porphyry dike swarms. The diabase dikes intruding the rapakivi granites are crossed in turn by later granitic dikes. Injections of basaltic melt into granitic magma chambers resulted sometimes in magma mingling with formation of basite "fountains" chilled in a cooler granitic magma. The ARGC formation was associated consequently with generation of magmas in the mantle and crust.

The ARGC rocks of the Baltic Shield originated 1.65–1.53 Ga ago. Silicic magmatism of the vast Trans-Scandinavian volcano-plutonic belt (Gorbatschev and Bogdanova, 1993) is somewhat older (1.78–1.59 Ga), represented mostly by volcanics of dacite–rhyolite composition, some of which are presumably the extrusive equivalents of the ARGC. The oldest ARGC rocks (1.78–1.74 Ga) are known in massifs of the Ukrainian and Greenland shields (Amelin et al., 1996b; Harrison et al., 1989).

The rapakivi granites of K-series are relatively enriched in iron and alumina. Characteristic of them are high Ti, Zn, Pb, Zr concentrations, elevated concentration level of Be, Sn, Y, Nb, Rb, F, W, and Mo, while considerable Li and U concentrations are established only in the late albite-rich differentiates. Mafic rocks of the ARGC relatively depleted in Mn, Ni, Cr, V, Cu, and Y reveal elevated Ce, Sn, Mo, La, Fe, Ti, P, and Pb concentrations (Velikoslavinskii et al., 1978). In their geochemical parameters, rapakivi granites are close to the A-type intraplate granites, especially to the Phanerozoic Li–F rare-metal granites and their extrusive equivalents (ongonites). A comprehensive isotopic-geochemical investigation of rapakivi granites in Fennoscandia showed that their parental magmas were generated in the Svecofennian crust of granodiorite composition (Rämö, 1991; Neymark et al., 1994). Feldspars from gabbro-anorthosites and granites under consideration are identical in terms of Nd, Sr and Pb isotopic composition: εNd(T) from –6.5 to –8.2, $^{87}Sr/^{86}Sr$ initial ratio from 0.7052 to 0.7057. Consequently, parental magmas of mafic rocks were generated by interaction of mantle-derived melts with crustal rocks. The REE concentrations in norites and gabbronorites of the Salmi pluton are close to those typical of the intraplate Fe–Ti subalkaline basalts originating under influence of mantle plumes. In these rocks modestly enriched in LREE, parameter εNd(T) ranges from –1.2 to +1.6 (Rämö, 1991). As is shown for the ARGC mafic rocks of the Suwalki massif in Poland, which are 1560 Ma old, they have high $^{187}Os/^{186}Os$ ratios and parameters γOs (from +640 to +902), whereas εNd is low (–6.0). These data also suggest a considerable contribution of crustal material to parental magmas of the rocks (Morgan et al., 2000).

As is noted above, the ARGC originated from magmas melted out simultaneously from the mantle and crust. The rock complexes as such corresponded to upper parts of trans-crustal magmatic systems, which originated above protuberances of mantle up to 10–20 km high (Orovetskii, 1990; Elo and Korja, 1993). At present, those systems are represented by thick alternating layers of mafic and silicic rocks, which suggest that large sill-like bodies of mafic composition (apparently Fe–Ti basalts similar in composition to dike rocks) intruded different levels of abnormally thick sialic crust and caused melting of crustal material above them (Sharkov, 1999). The intraplate activity resumed in the region 1.8–1.5 Ga ago was likely the main factor that controlled the magmatism under consideration. As one can judge from Fe–Ti basaltic magmatism characteristic of mantle plumes, the mantle protuberances beneath the ARGC areas represented the plume heads. The outlined long-lived magmatic systems (centers), composition and occurrence settings of Riphean deposits in the East European craton suggest a stabilized state of the latter.

In distinction from Svecofennides, endogenic processes 1.8–1.5 Ga ago in the craton eastern part with normal crust about 40 km thick were much less intense, responsible for events of continental rifting and emplacement of small basaltic or lamprophyric dike swarms. Silicic magmatism of the fourth stage corresponded apparently to a specific manifestation of intraplate magmatic activity in areas of anomalously thick crust, and mantle-derived basaltic magmas resided in mass within the crust, triggering an intense melting of sialic material.

The belt of predominantly silicic magmatism, as wide as 1000 km and extending from the Labrador to California for a distance of about 6000 km, is situated along southeastern and southern margins of the Mid-continent in North America (Semikhatov, 2002; Khain and Bozhko, 1988; Gower et al., 1990). Being formed 1.8–1.3 Ga ago, it probably comprises several belts different in age. The oldest in the belt are granite batholiths of the Churchill and Manitoba provinces and K-granites with associated subordinate quartz monzonites of the Trans-Labrador belt 500 km long. The older rocks were formed here 1.80–1.72 and about 1.65 Ga ago (Kerr and Fryer, 1990), while the bulk of plutons originated later, 1.51–1.3 Ga ago (Gower et al., 1990),

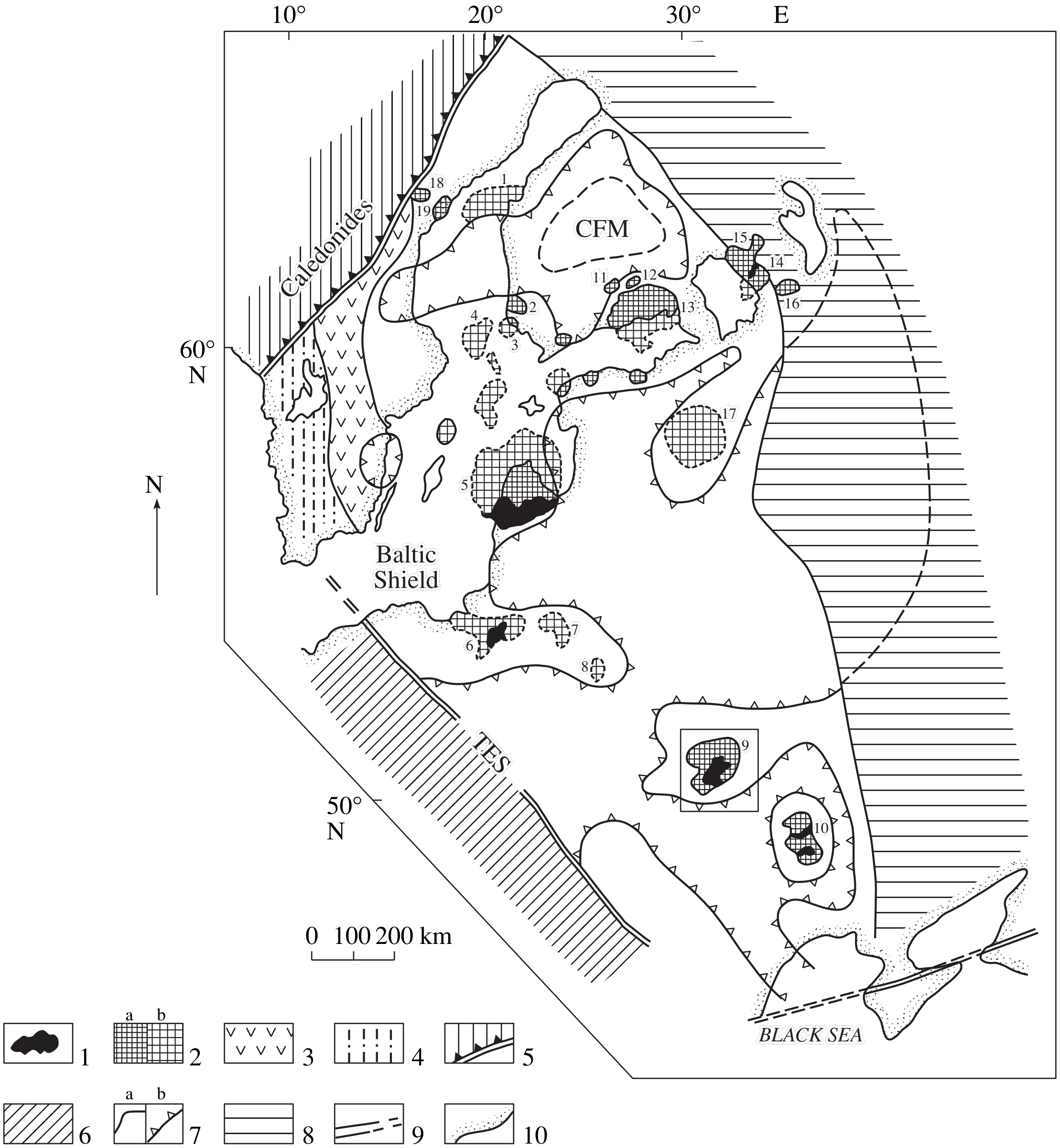

Fig. 3. Localities of anorthosite–rapakivi granite complexes in the west of the East European craton (after Sharkov, 1999, with modifications): (1) gabbronorite-anorthosite, anorthosite; (2) rapakivi granite observed (a) and inferred from geophysical data (b); (3) Trans-Scandinavian igneous belt; (4) Góta domain; (5) Caledonides of Norway and their boundary; (6) Phanerozoic foldbelts; (7) crust of Svecofennian orogen < 45 km thick (a) and > 45 km thick (b); (8) Archean to Early Paleoproterozoic crust of the East European craton; (9) marginal sutures of the East European craton; (10) coastline; (CFM) Central Finland massif; (TES) Trans-European suture. Plutons by names: (1) Bothnia; (2) Laitilla; (3) Vehmaa; (4) Aland; (5) Riga; (6) Suwalki; (7) Mazurki; (8) Belarusian; (9) Korosten; (10) Korsun-Novomirgorod; (11) Ahvenisto; (12) Suomenniemi; (13) Vyborg; (14) Salmi; (15) Ulyaleg; (16) Lodeinoe Pole; (17) Novgorod; 18) Ragunda; (19) Nordingra.

being represented by large bodies of anorthosites, anorthosite–diorites, and rapakivi granites.

The Trans-Siberian volcano-plutonic belt of similar type extends in the South Siberian platform for a distance of 3000 km from the Sea of Okhotsk to the Yenisei Ridge (Larin et al., 2003), being formed in post-orogenic environments ca. 1.9–1.7 Ga ago. Silicic volcanics, the dominant rocks of the belt, are tightly associated with siliciclastic red beds, rapakivi granites, subalkaline granites, and anorthosites. The belt is superimposed on diverse older structures of the Archean crystalline basement and Paleoproterozoic foldbelts. The U–Pb age of rocks in the belt increases in westward direction from 1.7 to 1.9; prevailing values fall in the period of 1.7–1.84 Ga. Crustal εNd(T) parameters ranging from –0.2 to –9.2 and T_{DM} model ages of 2.2 to 2.6 Ga feature isotopic-geochronological characteristics of silicic rocks. According to these characteristics, magmatic rocks of the Trans-Siberian belt are fairly close to rocks of North American belts, being indicative of variable contribution of crustal components to magma generation, but they are different from comparable rocks of East European belts. According to geochemical specialization (high K, Rb, Li, Be, Sn, W, Nb, Mo, Ta, Zr, REE, and F concentrations), silicic rocks under consideration correspond to typical A-granites of post-orogenic stages and/or intraplate settings. Analogous belts of silicic magmatism that was active 1.8–1.5 Ga ago and culminated at ca. 1.7 Ga are known in South America (Rio Negro–Juruena belt), Australia (Gawler Range belt), and the West Antarctic craton. The ~ 1.7 Ga rapakivi granites are detected as well in the Sino-Korean Shield and Northeast China (Rämö and Haapala, 1995).

GEOCHEMICAL EVOLUTION OF PALEOPROTEROZOIC MAGMATISM

As is evident from presented materials, Paleoproterozoic magmatic rocks reveal a directional trend of compositional changes through time, and this trend controlled respectively the composition of substance flux into biosphere. Principal trends in isotopic-geochemical evolution of magmatic rocks in the well-studied Baltic and Canadian shields throughout the Paleoproterozoic are illustrated in Figs. 4–7. The diagrams depict secular changes in bulk compositions and trace-element geochemistry (indicative geochemical ratios included) primarily of mafic volcanic rocks (up to 55% SiO_2), which are most confidently dated based on geological and isotopic-geochronological data.

Interpreting the diagrams, we should emphasize the following trends in geochemical evolution of Paleoproterozoic magmatism. (1) One of two maximums in Mg distribution through time is recorded at ~ 2.1 Ga, at the time of widespread eruptions of picritic magmas, and the other, less distinct one at 2.5–2.4 Ga, when magmatism of boninite-like series was intense. (2) Enrichment of mafic rocks in Fe and Ti commenced about 2.25 Ga ago, then reached the two relative maximums 2.1 and 2.0 Ga ago, and afterward concentrations of both elements were gradually reducing. The first maximum in concentration of both elements is interrelated with the grown intensity of Fe–Ti picritic and basaltic volcanism; the second one with manifestations of magmatism confined to continental rifts. (3) The P concentrations were highest 2.2 and ~ 1.9–1.85 Ga ago; they reflect widespread eruptions of subalkaline Fe–Ti basalts and manifestations of orogenic volcanism. (4) Four, relatively small peaks at 2.4, 2.2, ~1.9 and 1.6 Ga are detectable in secular variations of alumina content. Three earlier peaks were connected respectively with eruptions of boninite-like lavas, subalkaline basalts and calc-alkaline series in suprasubduction settings, while the fourth, less distinct peak marks the outburst of the ARGC magmatism. (5) The rise of CaO content at 2.1–2.0 Ga positively correlating with variations of CaO/Al_2O_3 ratio, Ti and Fe concentrations is indicative of juvenile Fe–Ti picrite-basaltic volcanism that became widespread at that time. (6) Insignificant maximums of Mn concentration at 2.1–2.05, 1.9–1.85 and 1.6 Ga ago are correlative with relative peaks of Fe-Ti basaltic volcanism. (7) Secular variations of silica content are rather irregular. The least silicic rocks were formed ~ 2.1 Ga ago; more silicic ones are confined to orogens of the third stage and to intraplate settings of the fourth stage.

Thus, we can see three main periods, when distribution of above components in the rocks was controlled by prevailing development of particular magmatism: 2.5–2.3 Ga ago, it was the SHMS volcanism represented by high-Mg rocks with low to modest concentration level of titanium; 2.1–1.9 Ga ago, it was Fe–Ti picritic to basaltic volcanism that gave place with time to calc-alkaline volcanism in orogens; and 1.7–1.5 Ga ago, it was intraplate magmatism resumed and superimposed on orogens. Also remarkable is a weak manifestation of juvenile volcanism 1.8–1.7 Ga ago, in the time span between terminated stabilization of Paleoproterozoic orogens and onset of intraplate magmatic activity of the Late Paleoproterozoic–Early Mesoproterozoic. Magmas, which were commonly depleted in Fe, Ti and Mg, characterized the period of a less intense magmatic activity 2.3–2.2 Ga ago.

The established distribution trends of Cr and Ni, the elements typical of high-Mg mantle derived magma (Figs. 5, 6), are correlative with distribution of major petrogenic components. Maximums of Cr and Ni concentrations at 2.1–1.9 Ga, which characterize Paleoproterozoic boninite-like rocks and Middle Paleoproterozoic volcanics, are almost concurrent to Fe, Ti and Mg peak concentrations at 2.1–2.0 Ga and to P_2O_5, SiO_2 and Al_2O_3 maximums at ~ 1.9 Ga. The highest Cu concentrations in the 2.1 Ga rocks are correlative with Cr and Ni concentration peaks, and lower though elevated Cu concentrations are established in rocks of the first and fourth stages. The Zn behavior is different: this element showed the main concentration peak at 1.9–1.8 Ga, at the time of widespread suprasubduction magmatism, and two lower peaks, one corresponding to the Early Paleoproterozoic boninitic volcanism and the other one to volcanism of the Trans-Scandinavian belt. The Co-concentration maximum at 2.3 Ga is followed by a long-term decline with subsequent minor peak at 1.8 Ga. The maximal enrichment in V is evident at 2.1–2.0 Ga, and lower extreme points characterize eruption time of boninitic lavas (that is consistent with presence of vanadium titanomagnetite in late differentiates of layered intrusions) and basaltoids of the Early Paleoproterozoic. Finally, the bimodal distribution is characteristic of Th: one maximum is recorded in the Early Paleoproterozoic boninitic rocks, being separated by the time span of 2.1–2.05 Ga, when the concentration

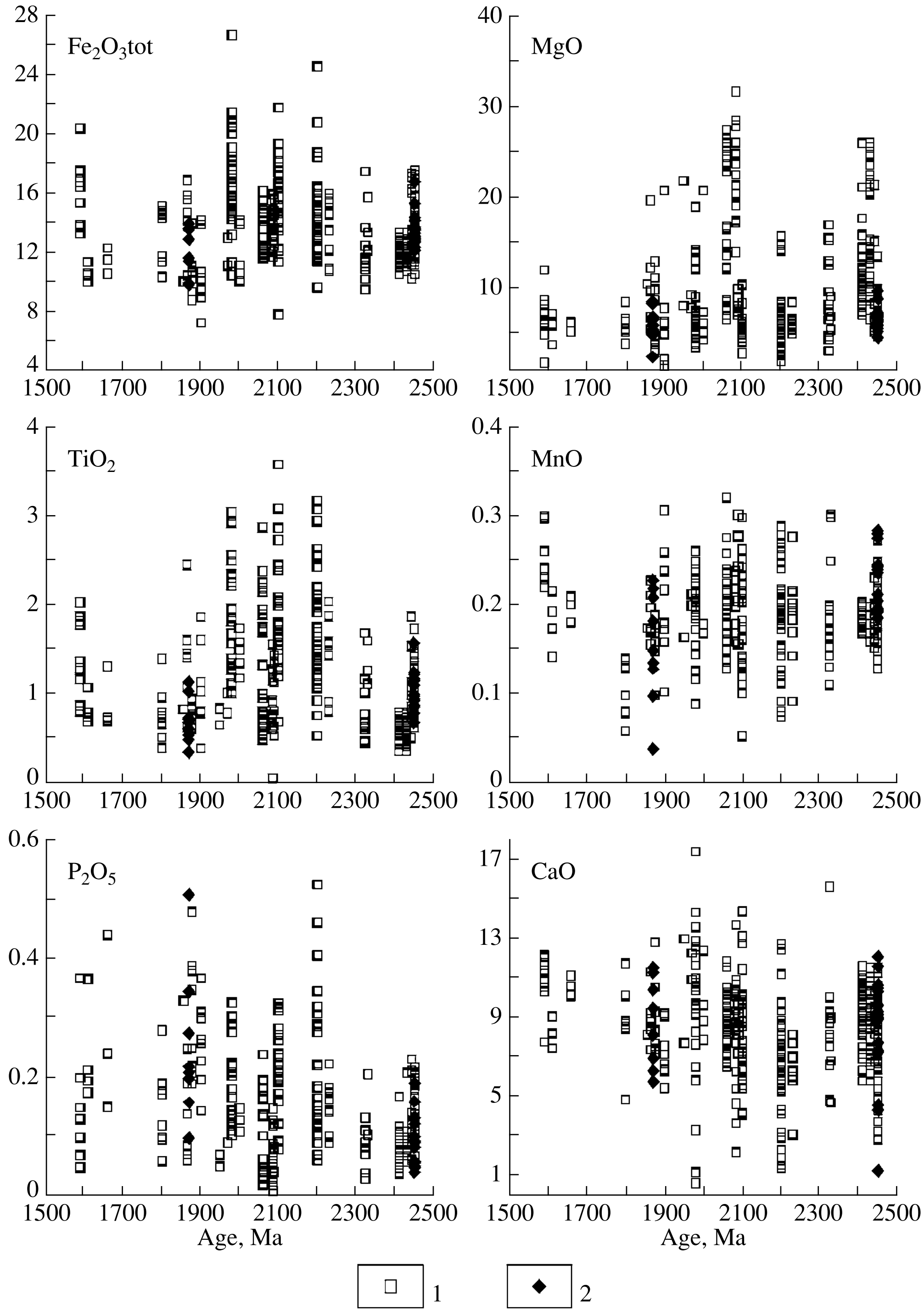

Fig. 4. Secular variations of major oxides (wt %) in Paleoproterozoic volcanics of the Baltic (1) and Canadian (2) shields.

level of femaphilic elements was relatively high, from the other maximum characteristic of the Late Paleoproterozoic basalts. In the intervening time span, contamination of mafic magmas by crustal material was likely at the minimum.

Comments to REE can be restricted by consideration of La and $(La/Yb)_N$ variations characterizing the relative fractionation of REE distribution patterns. As one can see in Fig. 8, there are two La concentration peaks, one correlative with manifestations of boninitic magmatism; the other one with eruptions of Fe–Ti picritic magmas 2.0–1.9 Ga ago. The first peak apparently reflects assimilation of Archean crustal material by mantle-derived magmas, while Fe–Ti picrites are

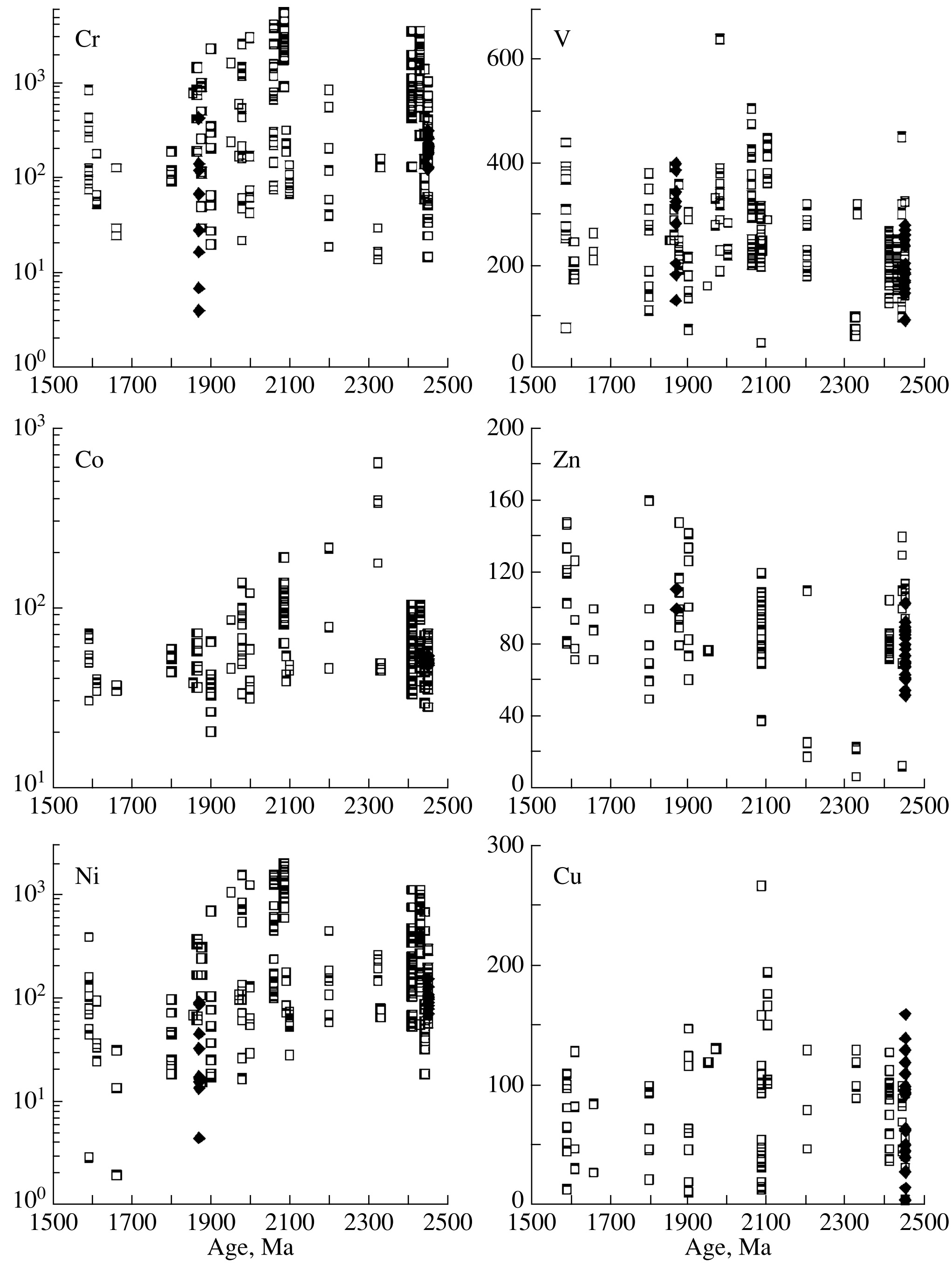

Fig. 5. Secular variations of major oxides (wt %), trace elements (ppm), and geochemical ratios in Paleoproterozoic volcanics of the Baltic (1) and Canadian (2) shields.

likely related in origin to enriched mantle. Variations of $(La/Yb)_N$ ratios, which are intricate in general, reveal nevertheless a minor peak at 2.0 Ga.

The Nd and Os isotopic parameters most stable in response to recurrent alteration processes are suitable properly for subsequent consideration that is also based, in view of limited data on volcanic rocks, on isotopic systematics of comagmatic dikes and intrusive bodies or on concurrent granitoids. According to data presented in Fig. 7, εNd values increased gradually 2.5–2.1 Ga ago from –6 that is typical of crustal material to juvenile level of +5, which corresponds to expectable parameters of depleted mantle at that time. This trend suggests that proportion of crustal components in magmas was apparently decreasing toward the Middle Paleoproterozoic, and ca. 2.1 Ga ago, at the time of most intense pulse of magmatic activity, it closely approached zero. Afterward, εNd values were

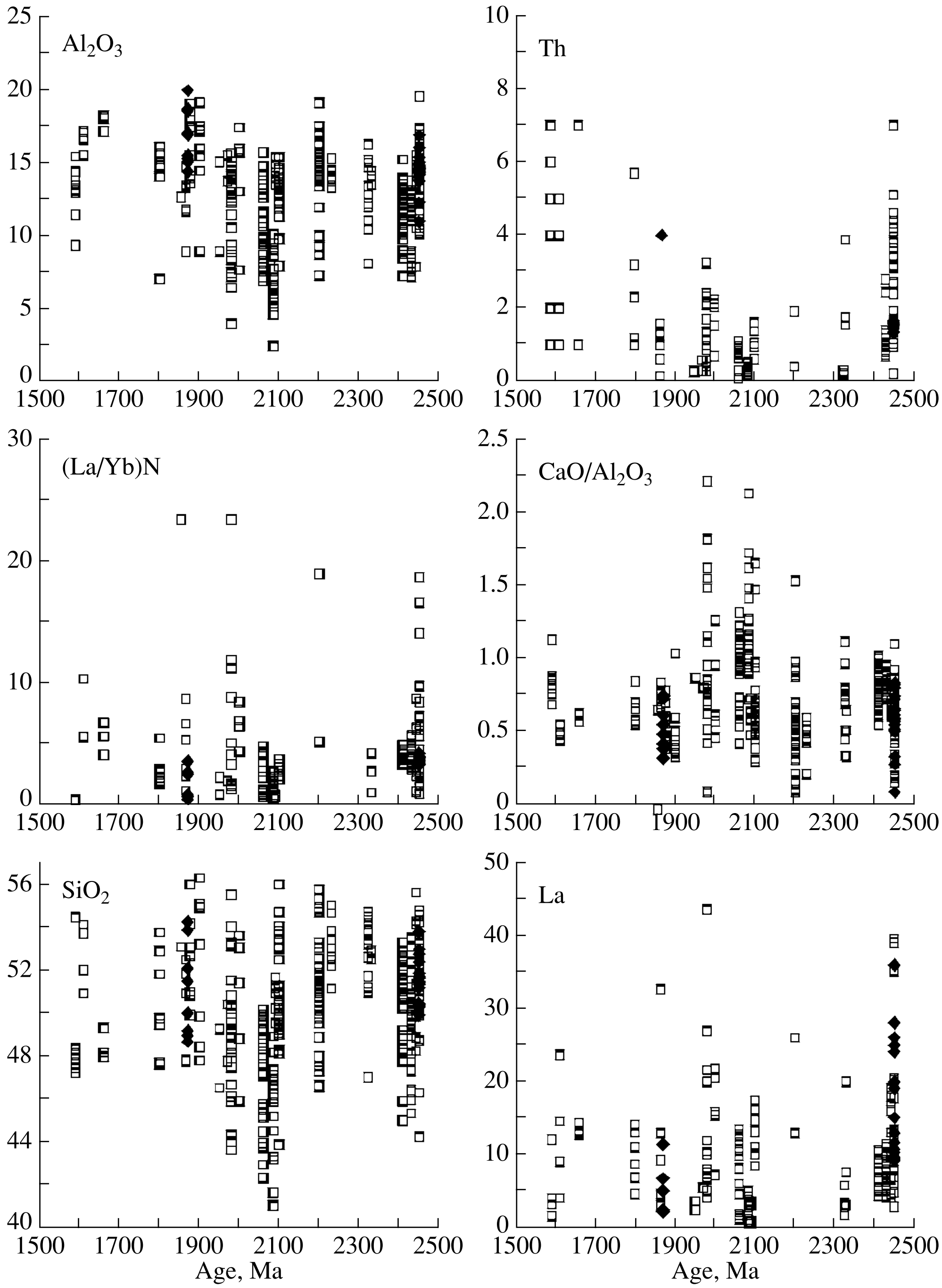

Fig. 6. Secular variations of trace elements (ppm) in Paleoproterozoic volcanics of the Baltic (1) and Canadian (2) shields.

declining a little though positive in general except for separate cases of crustal contamination (komatiitic lavas of the Onega structure and anorthosites).

Data on rocks series of the Baltic Shield, which originated 2.50–2.4, 2.05–1.75 and 1.5 Ga ago, exemplify secular changes in the Os isotopic parameters of magmatism. Volcanics of the Vetreny belt (2.44 Ga) to be considered first reveal a narrow range of γOs values from –0.43 to –0.07, which imply that relevant magmas have been derived from an upper-mantle plume depleted in Os as long before formation of volcanics as 150 m.y. (Puchtel, 2001). A wider range of γOs values (–6 to +1) characteristic of rocks from the Akanvaara and Koitelainen layered massifs (2.45 Ga old) suggests contamination of parental magmas by crustal material, when they ascended from mantle into crust (Hanski, 2001a). A much greater variability of Os isotopic parameters is recorded in rocks formed 2.05–1.9 Ga

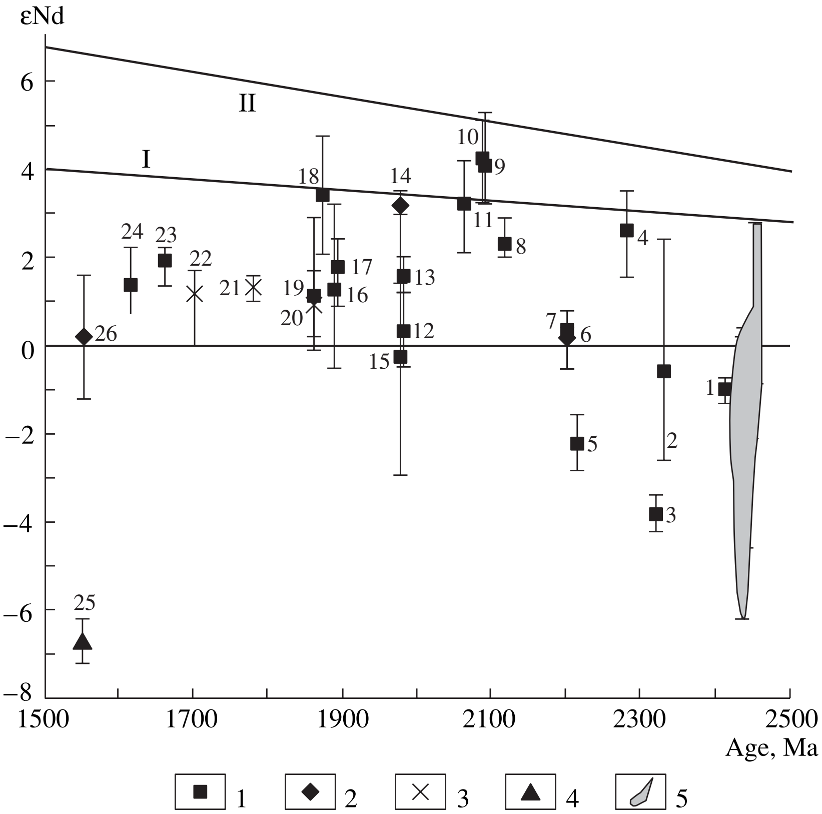

Fig. 7. Secular variations of parameter εNd(T) in Paleoproterozoic rocks of the Baltic and Canadian shields: (1) basic volcanics; (2) basic dikes and sills; (3) granitoids; (4) anorthosites; (5) layered mafic–ultramafic intrusions of the Baltic Shield. Lines I and II depicting isotopic evolution of depleted mantle are after De Paolo (1981) and Goldstein and Jacobsen (1988). Numbers in the figure denote the following objects; 1, Vetreny belt (Puchtel et al., 1998); 2, Perapohja belt, Runkaus volcanics (Huhma et al., 1990); 3, Pechenga, volcanics of Maajarvi Formation, Ahmalahta Group (Skuf'in et al., 2004); 4, Kautokeino greenstone belt, Kaskejas amphibolites (Krill et al., 1985); 5, Pechenga, volcanics of Pirtijarvi Formation, Kuetsjarvi Group (Skuf'in et al., 2004); 6, Hurwitz Group, Canada (Sandeman, 2003); 7, Perapohja belt, Runkausvaara sill (Huhma et al., 1990); 8, Pechenga, Zapolyarnyi Formation, Kolosjok Group (Skuf'in et al., 2004); 9, Perapohja belt, Jotiaappa volcanics (Huhma et al., 1990); 10, Karasjok komatiites (Krill et al., 1985); 11, Central Lapland greenstone belt (Hanski et al., 2001); 12, Pechenga, Matert Formation, Pilgujarvi Group (Skuf'in et al., 2004); 13, ditto, ferropicrites (Smolkin et al., 1992); 14, Onega Plateau (Puchtel et al., 1998); 15, Konchozero sill (Puchtel et al., 1998); 16, South Finland, Pellinki volcanics ((Patchett and Kuovo, 1986); 17, Central Sweden, western Berslagen (Valbracht, 1991); 18, Flin Flon (Thom et al., 1990); 19, Svecofennian basic volcanics, Sweden (Patchett et al., 1987); 20, Early Svecofennian silicic to intermediate volcanics and granitoids (Patchett et al., 1987); 21, 22, Trans-Scandinavian igneous belt, granitoids (Patchett et al., 1987); 23, Horred Formation (Brewer et al., 1998); 24. Amal Formation (Brewer et al., 1998); 25, anorthosite of Salmi batholith; 26, diabase dikes in the same batholith (Neymark et al., 1994).

ago. For instance, γOs values vary in the Ludicovian basalts of the Onega Plateau and picritic dolerites of the Konchozero sill from +0.77 to –0.61 (Puchtel et al., 1999). Ferropicritic lavas of the Pechenga region (Walker et al., 1997) have more radiogenic Os-isotope composition (γOs = +6.1). This and other isotopic-geochemical parameters suggest, being compared with isotopic characteristics of the Onega Plateau volcanism, that pertinent event of magmatism was rooted in a deeper hot area of mantle plume, which probably experienced influence of material derived from outer core, being contaminated by the Belomorian gneisses at the terminal stage of the belt formation (Puchtel et al., 1999). The widest range of γOs values (–5.1 to +3.0) is characteristic of the 1.95 Ga Jormua ophiolites, comprising rocks related in origin to the ancient subcontinental lithospheric mantle and a younger MORB-type source of magmas (Tsuru et al., 2000). Consequently, mantle of that time was already quite heterogeneous in terms of Os isotopic systematics.

Summarizing the above isotopic-geochemical data, we suggest that the Ludicovian igneous province and Svecofennian orogen originated under influence of one superplume that ascended from the core/mantle interface, and juvenile material transported from there was responsible for development of isotopic heterogeneity in the mantle ca. 2.0 Ga ago. As for extremely high γOs values (+640 to +902) established in anorthosites of the Suwalki massif (Poland), they and low negative εNd parameters imply a considerable contamination of parental mantle-derived magma by crustal material.

CONCLUSIONS

Data considered in this work show that history of Paleoproterozoic magmatic processes is divisible into four stages, each of specific character.

The *2.5–2.3 Ga* corresponds to formation period of siliceous high-Mg series (SHMS) of boninitic rocks in rifting-related volcanogenic-sedimentary belts and of large dike swarms and layered mafic–ultramafic plutons. The corresponding large igneous provinces of Archean cratons signify existence of mantle superplumes at that time composed of depleted ultramafic material. Magmatic activity of this kind was manifested already in the Late Archean, when rocks close in composition to boninites appeared in some greenstone belts, e.g., in the Whitney Township volcanic group of the Abitibi belts (Kerrich, 1998) or in the Khizovaara belt of the Baltic Shield (Shchipanskii et al., 1999). Concurrent emplacement of large layered intrusions was characteristic of stabilized craton zones. Well-known examples are the Stillwater Complex (2.7 Ga) in the Wyoming craton (Lambert et al., 1994) or the Great Dike (~2.6 Ga) in Africa (Mukasa et al., 1998). The high-grade metavolcanic rocks used to be considered as exemplifying the island-arc magmatism, but the layered plutons cannot be interpreted of course in terms of such a model.

In geochemical aspect, the SHMS magmas were characterized by elevated to high contents of silica, Mg, Cr, Ni, Co, Cu, V, PGE, and some incompatible trace elements, LREE included. Simultaneously, they were depleted in Fe, Ti, Nb and had negative εNd parameters. Consequently, the SHMS magmas were of a mixed mantle–crustal genesis. Being derived from ultramafic material of asthenospheric mantle depleted during previous events of magma generation, magmatic melts assimilated components of Archean crust during their ascent from the depth to the surface. The assimilation explains their similarity with island-arc volcanic series of the Phanerozoic, in particular with boninitic ones. Intrusive magmatism of K-granite composition was of a limited significance at the first stage.

The *second stage (2.3–2.05 Ga)* is exemplified by subalkaline Fe–Ti picritic and basaltic lavas enriched in characteristic geochemical components and erupted in mass, largely in the second stage half, in all the shields. This change in character of magmatism was not controlled by tectonic factors, because lavas erupted in the same rift structures, where they overlie the older successions, and at the site of the SHMS igneous province, there was formed similarly large province of predominantly tholeiitic basalts frequently enriched in titanium. Judging from data on rocks in the Pechenga structure (Smolkin et al., 1995) and central areas of the Aldan Shield (Puchtel and Zhuravlev, 1992), magmatism of the new type reflected influx of fluids enriched in Fe, Ti, alkalies, Zr, Ba, and LREE from the depth to superplume. Influence of identical fluids is postulated for recent superplumes, which take origin at the core/mantle interface (Condie et al., 2004).

High concentrations of alkalies, Fe, Ti, Nb, Mn, P, incompatible elements, LREE in particular, and relatively elevated concentration level of Cr, Ni, Co, Cu, and Ba characterized geochemistry of magmas at the second stage, which were simultaneously depleted to some extent in SiO_2, Mg, and Al. Positive εNd parameters typical of them reached the maximum (+5) 2.1 Ga ago. Magmas modestly enriched in Ti and alkalies like basalts of large igneous provinces of the Phanerozoic (traps and oceanic plateaus), prevailed at the beginning of the stage, and magmatic activity of the former type, the SHMS magmatism included, was characteristic of some separate regions (South America, Voronezh crystalline massif, the Antarctic, etc.).

The *third stage (2.05–1.8 Ga)* was signified by opening of first oceans with subduction zones and back-arc basins along their periphery and by formation of orogens resembling those of Phanerozoic time with relevant magmatism. Continental rifting with associated picrite-basaltic volcanism was characteristic of the orogens' periphery and adjacent marginal zones of Archean cratons.

The initial period of the stage (2.05–2.0 Ga) was marked by intense rifting-related volcanism responsible for eruptions of picrite-basaltic lavas with high Mg, Fe, Ti, Cr, Ni, Cu, V but extremely low Th concentrations. In many rift structures, there is recorded concurrent transition from continental to deep-water facies with tholeiitic pillow lavas corresponding to MORB in composition. Suprasubduction magmatism characteristic of the stage second half gave birth to rocks with very low CaO/Al_2O_3 ratio, relatively enriched in Si, Al, P, Zn, and Th. These rocks usually have positive εNd parameters except for particular cases, when they are contaminated by crustal material. As is known, volcanism of this kind is often explosive, ejecting into atmosphere the pyroclastic material and volcanic gases, mostly H_2O, CO, CO_2, sulfur compounds (mainly SO_3), F, Cl, and some other volatiles. Study of volcanic gases in the Kurile island arc active at present showed that they are enriched in many trace elements, such as Re, Mo, W, Cu, Co, I, Bi, Cd, B, and Br (Taran et al., 1995). Gases of similar composition could be typical of Paleoproterozoic volcanoes in the same tectonic settings.

Evolution of magmatism during the third stage was probably determined by increasing influx of fluids from the mantle interiors into magma-generating reservoirs of superplumes. The Os isotopic systematics suggests that mantle superplumes of that time could take origin at the core/mantle interface (Puchtel et al., 2001), as it is assumed for the recent superplumes (Condie et al., 2004). It is reasonable to think that influence of fluids could decrease viscosity and density of mantle material, and heads of superplumes were capable to penetrate higher into the lithosphere and to provoke changes in tectono-magmatic processes, opening of oceanic basins, and further development of orogens. In general, orogeny of the third stage is comparable, for instance, with orogenic development of the Alpine foldbelt (Sharkov and Svalova, 2005). Detectable in this belt are systems of volcanic arcs, back-arc basins with oceanic crust, and former seamounts, the sites of OIB magmatism, while continental rift zones with associated Fe–Ti basaltic volcanism are confined to the belt periphery (rifts in Central and western Europe, the Atlas, and elsewhere).

The *fourth stage (1.8–1.5 Ga)* commenced after the large-scale stabilization of the Earth crust, and relevant magmatism in intraplate settings was responsible for widespread development of volcano-plutonic belts, where eruptions of silicic K-enriched magmas dominated over basaltic volcanism. Intermediate magma chambers, which existed here under volcanoes, are apparently exemplified by anorthosite and rapakivi granite batholiths characteristic of the belts. Magmatism of the stage was predominantly superimposed on Paleoproterozoic orogens with thick to anomalously thick crust (up to 70–80 km at that time). Owing to this situation, the bulk of basaltic magmas was trapped in the crust that triggered melting of crustal material and gave rise to mass appearance of silicic magmas. In domains with crust of normal thickness (~40 km), there were concurrently formed continental rifts of a limited extension. Geochemical peculiarities of magmatism under consideration are enrichment in alkalies (primarily in K), Ti, Zn, Pb, Zr and relatively high Be, Sn, Y, Nb, Rb, F, Cu, W, Mo, sometimes Li and U concentrations. The negative εNd parameters and relative high Th and Zn concentrations most frequently observable in relevant rock imply that magmas were considerably contaminated by crustal components at the terminal stage of Paleoproterozoic evolution.

Thus, the Paleoproterozoic was an era of significant changes in tectonic and magmatic processes approaching the present-day status. Global trends and changes of magmatic evolution show that magma-generating reservoirs in the mantle, which were extremely depleted in terms of geochemistry at the beginning, turned into enriched later on. The most meaningful modifications in magmatic evolution happened 2.0–1.9 Ga ago at the onset of formation of first oceanic basins and orogens of Phanerozoic type. A considerable crustal contamination of early magmas evident from their isotopic parameters gave way to intense juvenile magmatism related in origin to mantle plumes, and then, by the end of Paleoproterozoic era, magmatism of crustal origin was of prime significance. The distinguished stages of magmatic evolution are traceable in all the Precambrian shields thus being of a global significance.

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