ISSN 0869-5911, Petrology, 2006, Vol. 14, No. 3, pp. 247–261. © Pleiades Publishing, Inc., 2006.
Original Russian Text © Sh.K. Baltybaev, O.A. Levchenkov, L.K. Levsky, O. Eklund, T. Kilpeläinen, 2006, published in Petrologiya, 2006, Vol. 14, No. 3, pp. 266–281.

Two Metamorphic Stages in the Svecofennian Domain: Evidence from the Isotopic Geochronological Study of the Ladoga and Sulkava Metamorphic Complexes

Sh. K. Baltybaev[a], O. A. Levchenkov[a], L. K. Levsky[a], O. Eklund[b], and T. Kilpeläinen[b]

[a]*Institute of Precambrian Geology and Geochronology, Russian Academy of Sciences, nab. Makarova 2, St. Petersburg, 199034 Russia*
e-mail: sb@sb2085.spb.ru
[b]*University of Turku, FIN-20014, Turku, Finland*
e-mail: n.n@utu.fi

Received December 8, 2004

Abstract—The Ladoga, Russia, and adjacent Sulkava, southeastern Finland, metamorphic complexes are the two largest “granulite” provinces of the Svecofennian domain. In this area, the domain is composed of outer and inner zones. Sulkava is situated in the inner zone, which principally can be compared to the accretionary arc complex of Southern Finland. Ladoga is situated in the outer zone, which is correlated with the accretionary arc complexes of central and Western Finland. The complexes contain different metamorphic assemblages, which are caused by the different composition of the sedimentary protoliths: the rocks of the Sulkava metamorphic complex are higher in Al and K than those of the Ladoga Complex. Pb–Pb step leaching dating was used to determine the age of prograde sillimanite from both complexes. The dates thus obtained constrain metamorphic peaks for the Sulkava and Ladoga complexes at 1799 ± 19 Ma and 1878 ± 7 Ma, respectively, which is consistent with the U–Pb monazite ages of gneisses from both of the complexes. The differences in the ages of the metamorphic minerals from these complexes reflect the Early Svecofennian (1.89–1.86 Ga) and Late Svecofennian (1.83–1.79 Ga) metamorphic stages in the Fennoscandian Svecofennides.

DOI: 10.1134/S0869591106030039

INTRODUCTION

The Svecofennian domain extends in a wide belt in the southern Fennoscandia over the territory of the Scandinavian and Baltic countries and Russia (Fig. 1). The domain has been subdivided into three distinct arc complexes depending on their ages and litologies: primitive arc complex of Central Finland, accretionary arc complex of Central and Western Finland, and accretionary arc complex of Southern Finland (Korsman et al., 1997). It was generally accepted in the late 20th century that the belt was formed by Early Proterozoic accretionary–collisional processes (Gaál and Gorbatschev, 1987; Glebovitsky, 1993; Nironen, 1997; Korsman et al., 1999; Baltybaev et al., 2000). There are different viewpoints regarding the number of endogenous events during the cratonization of the Svecofennian domain. Some geologists argue for the uninterrupted evolution of the belt, without discrete phases or stages (Gorbatschev and Bogdanova, 1993). Other researchers suggest that the ages of syncollisional magmatism and related metamorphism in some structures testify that the evolution of the Svecofennides was discrete (Korsman et al., 1984; Väisännen et al., 2002). The duration and number of stages were not determined partially because of the spatiotemporal heterogeneity caused by the asynchronous culmination of the same processes in the different parts of the Svecofennian domain (Korsman et al., 1984).

Recent isotopic geochronological data have demonstrated the multistage character of the plutonic–metamorphic evolution of the Svecofennides, which involved, in addition to endogenous activity at 1.89–1.87 Ga (Gaál and Gorbatschev, 1987; Glebovitsky, 1993; Korsman et al., 1999; Baltybaev et al., 2000), repeated tectonothermal activation at ~1.83–1.80 Ga (Väisänen et al., 2002, 2004; Andersson et al., 2004). Our research was aimed at studying two Svecofennian areas: the high-temperature core of the Ladoga metamorphic complex in Russia and the adjacent Sulkava metamorphic complex in southeastern Finland, which are considered to be typical supracrustal areas of the Svecofennian domain. The U–Pb and Pb–LS dates obtained on rock-forming and accessory minerals of aluminous gneisses from these areas also confirm two-stage metamorphism of the Svecofennian domain. The need for such a research was determined by the following factors: (1) the studied areas are the two largest spatially close provinces metamorphosed to the granulite facies, (2) the investigations concern the correlation of the Svecofennides, in particular, identification of the tectonothermal activity of the Late Svecofennian stage, and (3) isotope–geochronological data on the metamor-

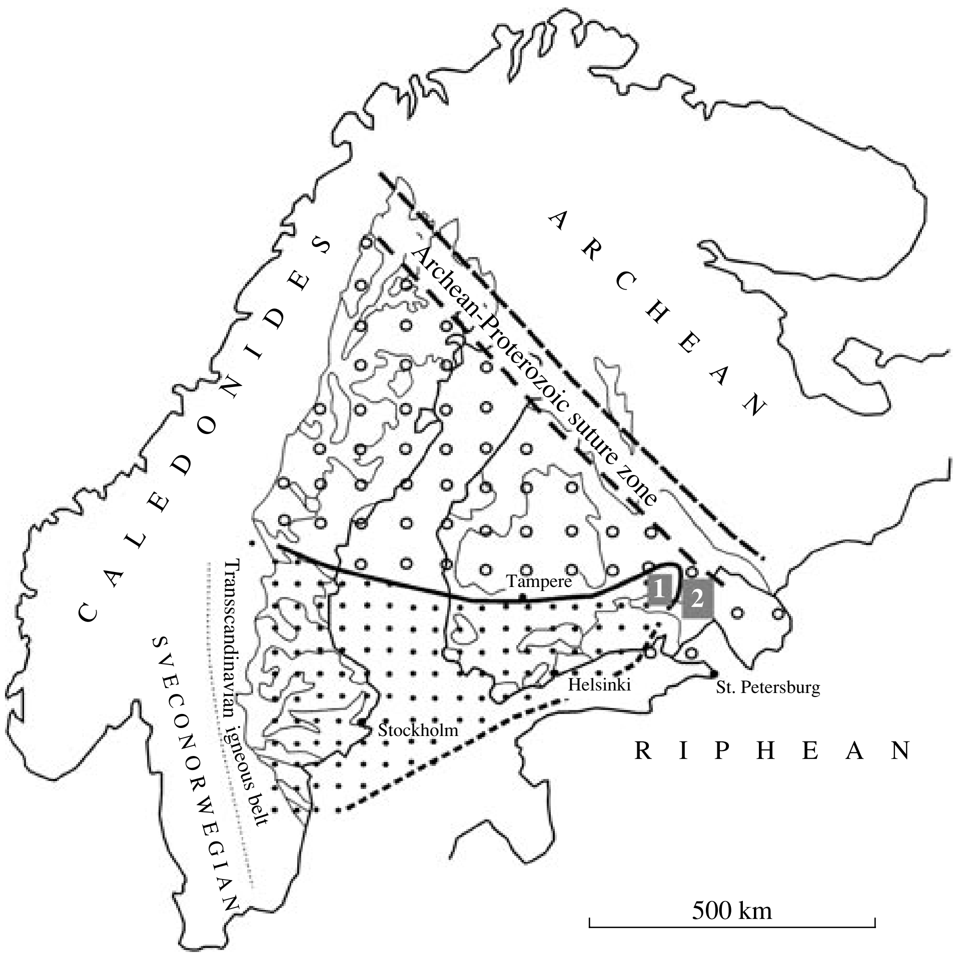

Fig. 1. The Svecofennian domain in the Fennoscandian shield.
Two zones: outer (open circles) and inner (filled circles). Study areas: (1) Sulkava area in Finlad, (2) Ladoga area in Russia. Sea text for explanations.

phism of these complexes are still scarce and often controversial.

GEOLOGICAL SETTING OF THE LADOGA AND SULKAVA METAMORPHIC COMPLEXES

The Svecofennian domain was formed at ~2.0–1.7 Ga. This period of time involved the accumulation of volcanosedimentary sequences and the formation of several plutonic and volcanoplutonic associations. The plutonic events were accompanied by metamorphism, locally, up to the granulite facies (Gaál and Gorbatschev, 1987; Korsman et al., 1999; and others). The Svecofennian basin is marked by an ophiolite association, which is partially preserved at Outocumpu and Jormua in Finland (Kostinen, 1981; Peltonen et al., 1996). This rock association is dated at 1.97–1.95 Ga (Kostinen, 1981; Kontinen, 1987). Ophiolites are overthrust onto the margin of the Archean Karelian craton, assembling with the younger volcanosedimentary rocks into the low-grade structural–metamorphic complexes of the pericratonic zone. Ophiolites are overlain by Kalevian deposits of diverse volcanic rocks and metamorphosed turbidites (Gäal and Gorbatschev, 1987), which were accumulated at 1.96 (1.92)–1.87 Ga (Gaál and Gorbatschev, 1987; Nironen, 1997; Baltybaev et al., 2000; Shul'diner, 2000).

According to composition, metamorphism, and migmatization, the Svecofennian domain can be subdivided into two zones: outer and inner (Fig. 1).

The outer Svecofennian zone is composed of mainly calc-alkaline granitoids and related metamorphic and volcanic rocks with a predominant age of 1.89–1.86 Ga. This interval includes both plutonic and metamorphic events and corresponds to the peak of Svecofennian orogeny, which was mainly responsible for the formation of the Proterozoic crust (Huhma, 1986; Gaál and Gorbatschev, 1987; Nironen, 1997; Rämö et al., 2001; Baltybaev et al., 2004a). The intrusions formed at that time are mostly ascribed to the I type varying from gabbrodiorites and tonalites to plagiogranites and granites. They are spatially associated with tonalitic plagiogneisses and migmatites. These rocks occur in the Skel-

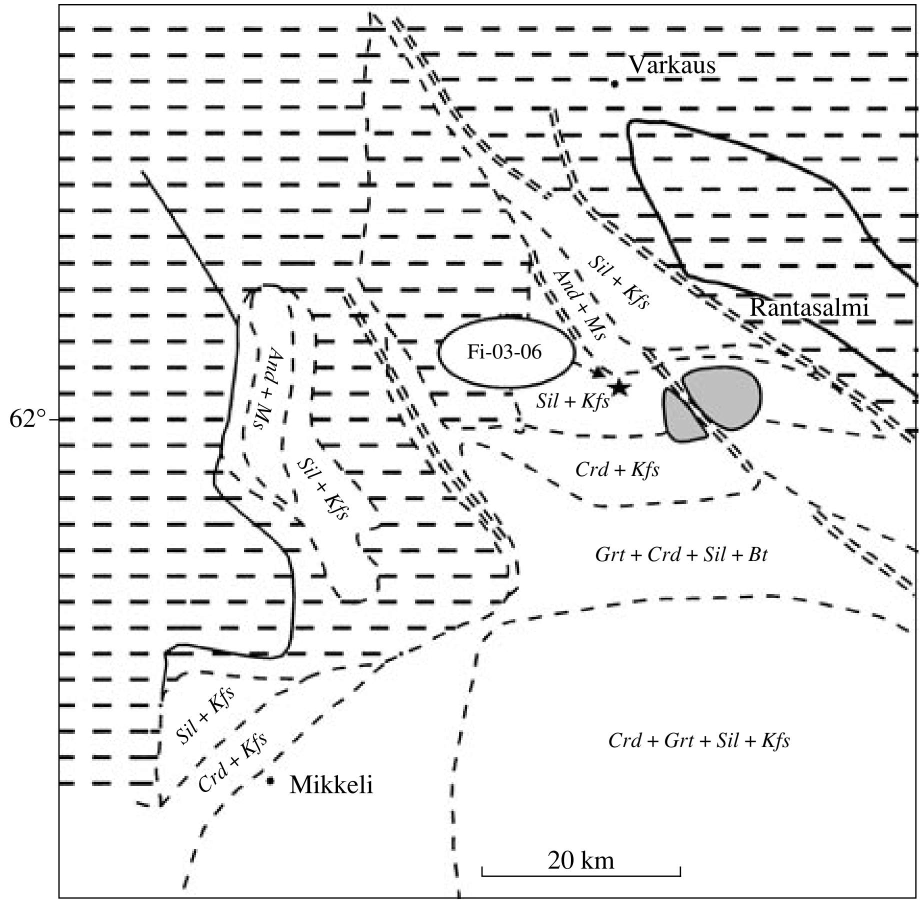

Fig. 2. Schematic map of the Sulkava metamorphic complex with metamorphic zones (modified after: Korsman, 1977; Korsman et al., 1988; Vaasjoki and Kontoniemi, 1991).

The Osikonmäki Massif is shown in gray, intrusive and supracrustal complexes ascribed to the outer Svecofennian zone (see text) are shown by horizontal hatching. The geochronological sampling site is marked by an asterisk.

lefte in Sweden, and Vihanti-Puhäsalmi, Kiuruvesi–Pielavesi–Rautalampi in Finland (Korsman et al., 1999). The complex also includes the Central Finland granitoid massif and Tampere belt (Nironen, 1997; Rämö et al., 2001). The high-temperature core of the Ladoga metamorphic complex is compositionally similar to the complexes of the outer Svecofennian zone (Glebovitsky, 1993; Baltybaev et al., 2000).

The inner Svecofennian zone, known as the Late Svecofennian granite–migmatite zone, is made up mostly of S-type granites with an age of 1.84–1.80 Ga (Ehlers et al., 1993), which are associated with aluminous metapelites and potassium migmatites. These associations include high-temperature complexes of Sulkava, West Uusimaa, and Turku in Finland (Väisänen et al., 2004), and the Berslagen metamorphic complex in Sweden (Andersson et al., 2004).

The tectonic and isotope study of the plutonometamorphism points to a structural–metamorphic disconformity between the outer and inner zones in Finland or, in other words, between the zones of tonalite and potassium migmatization (Korsman et al., 1988; Vaasjoki and Sakko, 1988, etc.). There is evidence that potassium migmatites were deformed and metamorphosed after crustal consolidation in the zone of tonalite migmatites (Korsman et al., 1984, 1988). The regional subdivision of the Svecofennides into potassium (granite) and tonalite migmatites is also confirmed by the presence of deep-seated structures composed of rocks with high electrical conductance, as well as by conformable boundaries of the migmatite areas (Korja et al., 1993).

Sulkava Metamorphic Complex

The Sulkava metamorphic complex is of crucial importance for characteristics of metamorphism in the inner (potassium) Svecofennian zone of Finland. The metamorphic grade of the complex increases southward, demonstrating a subsequent change in the mineral assemblages of the metamorphic rocks (Fig. 2). The complex belongs to the Southern Finland volcano sedimentary complex (Korsman et al., 1984, 1999).

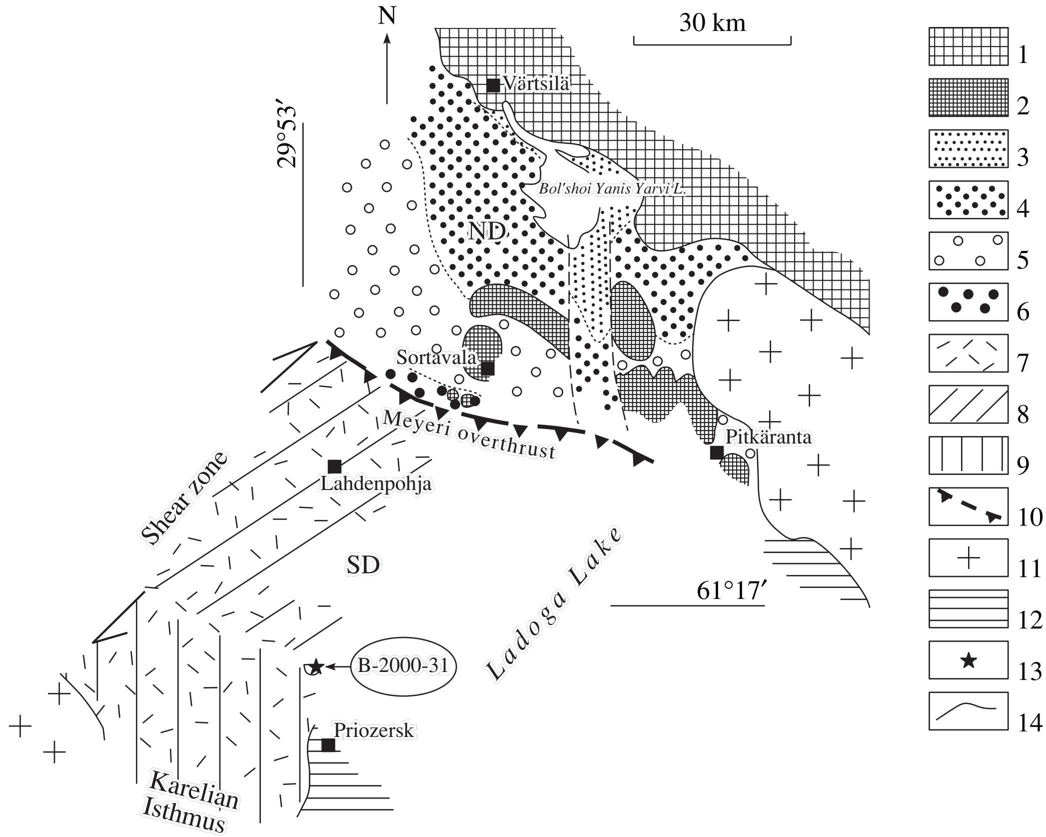

Fig. 3. Schematic map of the metamorphic zoning of the Ladoga metamorphic complex. (1) Archean rocks of the Karelian craton; (2) inliers of the Archean basement in the cores of rimmed gneissic domes; (3–7) metamorphic zones in the Lower Proterozoic volcanosedimenatry cover: (3) biotite and garnet, (4) staurolite–andalusite, (5) sillimanite–muscovite, (6) sillimanite–K-feldspar, (7) hypersthene (garnet–cordierite–K-feldspar); (8, 9) zones of ultrametamorphism and granitization with (8) the predominance of plagiogneisses and plagiomigmatites and (9) with the predominance of garnet–cordierite–K-feldspar gneisses and K migmatites; (10) Meyer thrust separating the Southern and Northern domains; (11) rapakivi granites; (12) Riphean platform cover; (13) geochronological sampling site; (14) other geological boundaries. (ND) Northern domain, (SD) Southern domain. The metamorphic zones are termed with regard for the previous schemes of metamorphic zoning (Velikoslavinskii, 1972; Nagaitsev, 1974).

According to Korsman (Korsman, 1977), the Sulkava complex is subdivided into the andalusite–mica schists, K-feldspar–sillimanite, K-feldspar–cordierite, and garnet–cordierite–K-feldspar zones. This author (Korsman, 1977) proposed simplified reactions to describe the prograde metamorphism. The reactions were written on the basis of the observed transitions from andalusite assemblages into sillimanite, muscovite–quartz into sillimanite-K-feldspar, and biotite–sillimanite into garnet–K-feldspar–cordierite assemblage. The sedimentary textures are gradually obliterated with increasing metamorphic grade; the structural features of turbidite sediments, in particular, their graded bedding, are preserved in mica schists only in the northern part of zoned complex. The sillimanite–K-feldspar gneisses also retained stratified structures with vague graded bedding. The K-feldspar–cordierite zone contains migmatized gneisses, with bedding preserved only locally in the northern part and completely obliterated in the southern part. The southern part contains synmetamorphic hypersthene diorites and hypersthene gneisses (Korsman, 1977).

Ladoga Metamorphic Complex

Geological structures of the Ladoga area (Fig. 3) belong to the Raahe–Ladoga zone, extending for more than 400 km (*Geological Evolution…*, 1970).

The structures are composed of Archean and Lower Proterozoic rocks. The Archean rocks are basement granite gneisses, which compose the cores of mantled gneiss domes (Eskola, 1949). The Archean age (2.7 Ga) of the magmatic rocks is supported by the U–Pb zircon dating of the granite gneisses (Tugarinov and Bibikova, 1980). The Early Proterozoic supracrustal complex is subdivided into the Sortavaala and Ladoga groups. The rocks of the Sortavala Group surround the cores of gneissic domes, being known only within pericratonic zone. U–Pb zircon datings constrained the age of volcanic rocks of the Sortavala Group at 1.96 Ga (Baltybaev et al., 2000; Shul'diner et al., 2000).

The deposits of the Ladoga Group unconformably overlay volcanic rocks of the Sortavala Group. The low-grade rocks of the Ladoga Group retained primary textures, which allowed the identification of sandstones, siltstones with gravelstone interbeds, and, in places, rhythmical graded bedding. Within amphibolite–granulite domain, where all traces of stratification were obliterated, they are distinguished as independent Lahdenpohja metamorphic groups (Shul'diner et al., 1996). U–Pb dates on recently found acid metavolcanic rocks constrain the age of this volcanosedimentary sequence at 1.884 Ga (Baltybaev and Levchenkov, 2005).

The Ladoga region is the area of high-grade (zonal) metamorphism of andalusite–sillimanite series (Predovskii et al., 1967; *Geological Evolution…*, 1970; Velikoslavinskii, 1972; Nagaitsev, 1974). The Ladoga region is confined to the juncture of the Epiarchean Karelian craton and the Svecofennian domain, the two largest structures of the Baltic Shield. In this regard, the Ladoga area is subdivided into two structural blocks or domains.

Northern domain. The northern boundary of the domain is marked by exposures of the granite gneiss basement of the Karelian craton, while the southern boundary (Fig. 3) runs along the outermost mappable units of the Archean rocks and represents the juncture between the Northern and Southern domains; it is distinguished as the Meyeri overthrust (Baltybaev et al., 1996). The structure of the Northern domain is defined by Archean blocks, the largest of which are Kokkaselkä, Sortavala, Kirjavalahti, Impilahti, Pitkäranta, Mursula, and some others. These structures are traced to the northwest in Finland as extended ranges of domes (Oravisalo, Sotkuma, Liperi, Kontiolahti, and others).

Southern domain. This domain extends to the Wiborg rapakivi granite massif (Fig. 3). The Southern domain is subdivided into two lithofacies zones: (1) a zone dominated by metagraywackes, marly rocks, and calc-alkaline metavolcanic rocks (Baltybaev and Levchenkov, 2005); (2) a zone composed of mainly metapelites and moderately aluminous metasiltstones with elevated potassium concentrations (Shul'diner et al., 1997; Baltybaev et al., 2002; Baltybaev and Levchenkov, 2005). The rocks of both zones are strongly migmatized, with the predominance of Na migmatites in the former zone and potassium migmatites in the latter.

The Southern domain differs from the Northern one by higher grade metamorphism and the abundance of ultrametamorphic products and numerous intrusive bodies, including those having no analogues in the Northern domain. The rocks of the Southern domain show neither geological nor geochronological evidence of the presence of Archean rocks. In addition, the Southern domain does not contain analogues of the Sortavala Group. Since metamorphic rocks of the Ladoga area rest on diverse basement complexes, the comparative study of the Svecofennian evolution should be based on the rocks of the Southern domain, a structure of the Svecofennian stage proper.

The Sulkava metamorphic complex differs from the Ladoga Complex by mineral assemblages that define the metamorphic zoning. The Sulkava complex is abundant in aluminous and potassium mineral assemblages and contains almost no staurolite assemblages. The complex is dominated by the K-feldspar–sillimanite assemblage, which is atypical of the moderate-grade metamorphic rocks of the Ladoga region. The aforesaid differences are difficult to explain by possible *P–T* variations; moreover, thermobarometric investigations and paragenetic analysis of minerals showed similar *P–T* conditions in the comparable zones of the complexes (*Geological Evolution…*, 1970; Nagaitsev, 1974; Korsman, 1977; Korsman et al., 1984; *Migmatization and Granite Formation…*, 1985; Baltybaev et al., 2000, 2004a). At the same time, the differences revealed in the mineral assemblages can be explained by differences in the primary lithology and chemical composition of the complexes: higher Al/(Ca + Mg) and K/(Ca + Mg) ratios in the rocks of the Sulkava metamorphic complex as compared to those of the Ladoga Complex (Korsman, 1977; Baltybaev and Levchenkov, 2005).

MAIN RESULTS OF THE PREVIOUS ISOTOPE–GEOCHRONOLOGICAL INVESTIGATIONS OF THE SVECOFENNIDES OF RUSSIA AND ADJACENT FINNISH AREAS

The age of metamorphism of the Sulkava Complex was constrained by U–Pb monazite and zircon dates (Korsman et al., 1984), some of which appeared to be controversial. The accessory monazite and zircon were taken from the migmatite mesosome and the leucosome of garnet–cordierite gneisses. The zircon from leucosome defined an age of 1833 ± 16 Ma, while the zircon from the mesosome showed a younger age of 1810 ± 7 Ma. Monazite yielded the opposite age relations: 1817 ± 4 Ma for the leucosome and ~1840 Ma for the mesosome (Korsman et al., 1984). Later dating of monazite from gneiss of garnet–cordierite zone gave $^{207}Pb/^{206}Pb$ age of 1796 Ma (Vaasjoki and Sakko, 1988). The inconsistent age relations on monazite and zircon from the meso- and leucosome and the high measurement error in zircon age of the leucosome complicated interpretation of these data. Therefore, these results had no decisive significance for the analysis of the Svecofennian metamorphism. Moreover, the proximity and similar sequence of metamorphic zones do not rule out some similarity in the metamorphic isogrades (Shul'diner et al., 1997; Baltybaev et al., 2000). Other geologists (for example, Ehlers et al., 1993) believed that the Ladoga region pertains to the Late Svecofennian granite–migmatite zone.

The high-temperature core of the Ladoga metamorphic complex is regarded as a single Svecofennian block, which was metamorphosed under the granulite (high-temperature garnet–cordierite–orthoclase) facies

1.88–1.86 Ga ago (Baltybaev et al., 2004a). No relations were found in the variations of prograde metamorphism within the core, with the exception of a systematic northward increase in the pressure (toward the Meyeri overthrust, which separates a high-temperature core from lower temperature rocks) (Shul'diner et al., 1997). The structural observations and U–Pb data indicate that the zones of K-feldspar and plagioclase migmatites in the core formed simultaneously at 1.88–1.87 Ga (Baltybaev et al., 2002) under similar *P–T* conditions (Baltybaev et al., 2004a). We have no unambiguous explanation for the disagreement between ages on similar zones in Russian and Finland, in particular, older ages (~1.87 Ga) of some S-type granite intrusions in the potassium zone of the Ladoga area as compared to those of compositionally similar intrusions in southern Finland (Baltybaev et al., 2004b).

Judging from the ages on volcanic rocks and estimates of the early magmatism (Nironen, 1997; Väisänen et al., 2002; Baltybaev et al., 2004a; Baltybaev and Levchenkov, 2005), the compositional difference between Svecofennides of the outer and inner zones is not related to the different accumulation time of the sequences. Petrographic, geochemical, and statistic data on detrital zircons from the Svecofennian paragneisses imply their different provenances (Lahtinen et al., 2002).

FACTUAL MATERIAL, SAMPLING PRINCIPLE, AND CHARACTERISTICS OF THE SAMPLES

The dating of metamorphism is a difficult task because of the genetic heterogeneity and long-term multiple evolution of metamorphic rocks. It is necessary to prove that ages obtained on the accessory minerals correspond to the age of the rock-forming minerals. This research is based on the isotope study of sillimanite, a metamorphic mineral that is directly related to the formation of the corresponding metamorphic assemblages in Svecofennian peraluminous gneisses. Our studies also involved the dating of monazites, which are currently thought to most closely reflect peak or post-peak metamorphic conditions (Daly et al., 2001; Baltybaev et al., 2004a).

The peraluminous gneisses occur in the high-aluminous sequence of the garnet–cordierite–K-feldspar zone of the Southern domain in the Ladoga area (Fig. 3). This sequence consists of unevenly migmatized rocks: monotonous and coarsely banded *Grt* + *Bt* + *Pl* ± *Kfs*, *Grt* + *Bt* + *Crd* + *Sil* + *Kfs* + *Pl*, *Bt* + *Sil* + *Pl* + *Kfs*, *Bt* + *Sil* + *Crd* + *Pl* ± *Kfs* gneisses with scattered, vein, and vein-spotty leucosome.[1] The sequence strikes northwest between 270° and 290° and dips subvertically, as was shown by polydeformational structures. Isoclinal folding with hinge planes dipping S–SE at 40°–60° is well expressed. The mineral lineation (*Bt*, *Sil*) has the same orientation.

The gneiss (sample B-2000-31) is a pinkish–dark gray rock with patches of subequant, often deformed aggregates of pink garnet up to 3–4 cm and larger. Some crystals typically show a skeletal morphology. The matrix consists of *Bt* + *Grt* + *Crd* + *Sil* + *Kfs* + *Pl* + *Qtz*. Sericitization is weak.

According to Baltybaev et al. (2004a), the gneisses were metamorphosed at 780–850°C and 5–6 kbar.

Sillimanite typically occurs as small crystals 0.05–0.1 mm in size, occasionally up to 0.5–1 mm. They are elongated (length to width ratio of 10 : 1), translucent, pale brown, brownish, occasionally colorless, with characteristic birefringence. No heterogeneities are visible under an optical microscope. The dating was conducted on selected grains of different sizes.

Monazite typically forms euhedral and rounded pale yellow crystals, which are rarely prismatic and transparent. The grains are 30–75 μm in size and occasionally contain acicular sillimanite. Fifteen purest crystals were taken for dating.

The peraluminous gneiss (sample Fi-03-06) of the Sulkava Complex was taken from the aluminous sequence in the sillimanite–K-feldspar zone (Fig. 2). The sequence trends nearly sublatitudinally, dips subvertically, and shows practically no traces of migmatization. The gneiss is a dark gray rock with spotty aggregates of *Sil* + *Kfs* up to 1–3 cm in size. The matrix is made up of *Bt* + *Sil* + *Kfs* + *Pl* + *Qtz*.

The garnet–biotite geothermometer was used to estimate the metamorphic temperature for gneisses from the sillimanite–K-feldspar zone (Korsman, 1977; Korsman et al., 1984). The garnet and biotite show wide variations in the different subzones of the sillimanite–K-feldspar zone: X_{Mg}^{Grt} = 0.114–0.223, X_{Mg}^{Bt} = 0.286–0.537. The metamorphic temperature at the sampling site Fi-03-06 was determined as 640–647°C (Korsman, 1977).

Sillimanite. The mineral occurs as <0.05-mm elongated, often felty fibrous aggregated crystals. The grains are light brown, brownish, occasionally colorless and translucent, and optically homogeneous. Ages were determined for sillimanite grains of different size.

Monazite is dominated by rounded pale yellow and brownish grains 50–100 μm in size. Twelve purest grains were taken for dating.

The criteria of sillimanite formation under prograde conditions and evidence for correlation between the monazite age and metamorphic events will be considered below.

In order to characterize the formation of metamorphic mineral in the complexes, we used polished thin sections prepared of our samples and made available for us by courtesy of geologists from the Turku University (several hundreds of polished thin sections). Our

[1] Abbreviations: (*Als*) aluminosilicate Al_2SiO_5, (*Bt*) biotite, (*Crd*) cordierite, (*Grt*) garnet, (*Fsp*) feldspar, (*Kfs*) K-feldspar, (*Mnz*) monazite, (*Ms*) muscovite, (*Pl*) plagioclase, (*Qtz*) quartz, (*Sil*) sillimanite, (V) vapor, (L) liquid.

Table 1. Results of isotopic study of sillimanite (samples B-2000-31 and Fi-03-06) and tourmaline (sample Fi-03-06)

No.	Fraction	$^{206}Pb/^{204}Pb^{a}$	V^{b}, % (2σ)	$^{207}Pb/^{204}Pb^{a}$	V^{b}, % (2σ)	Rho^{c}	$^{208}Pb/^{204}Pb^{a}$	V^{b}, % (2σ)	Rho^{d}	Th/U^{e}
	Ladoga region, B-2000-31, sillimanite									
1	Bulk, I	53.293	0.65	19.627	0.23	0.93	78.662	0.53	0.99	3.2
2	Bulk, II	49.319	1.00	19.156	0.33	0.95	73.326	0.79	0.99	3.1
3	Leachate, I	27.604	0.29	16.671	0.11	0.69	80.906	0.48	0.99	10.4
4	Leachate, II	25.330	0.10	16.417	0.10	0.65	69.348	0.10	0.92	9.5
5	Residue	72.708	1.10	21.858	0.44	0.98	67.576	0.65	0.99	1.6
	Sulkava, Fi-03-06, sillimanite									
1	Bulk	23.554	0.10	16.170	0.10	0.97	41.357	0.13	0.95	2.1
2	Leachate, I	34.970	0.20	17.434	0.10	0.67	51.934	0.16	0.85	2.4
3	Leachate, II	36.357	0.20	17.573	0.11	0.69	53.801	0.16	0.86	2.5
4	Residue, I	22.235	0.41	16.027	0.22	0.69	38.169	0.23	0.36	1.2
5	Residue, II	21.526	0.66	15.952	0.20	0.69	37.220	0.27	0.52	0.9
	Sulkava, Fi-03-06, tourmaline									
1	Bulk	46.654	0.10	18.786	0.09	0.79	81.549	0.14	0.92	4.1
2	Leachate	882.26	6.50	110.96	5.70	0.99	1360.200	6.50	0.99	4.2
3	Residue	19.733	0.06	15.816	0.09	0.97	40.723	0.12	0.98	3.7

Note: (a) Isotopic ratios corrected for the laboratory blank and mass fractionation during mass spectrometric analysis; (b) individual measurement errors of isotopic ratios; (c) correlation coefficients of errors in the $^{207}Pb/^{204}Pb$ and $^{206}Pb/^{204}Pb$ ratios; (d) correlation coefficients of errors in the $^{208}Pb/^{204}Pb$ and $^{206}Pb/^{204}Pb$ isotopic ratios; (e) Th/U ratio at the moment of sillimanite crystallization as calculated from Pb isotopic composition and mineral age.

research was accompanied by microprobe analyses of minerals and chemical analyses of metamorphic rocks. Short-term field works were conducted at the key sites of the two compared regions in 2002–2003.

ANALYTICAL TECHNIQUES

Sillimanite was dated by Pb–Pb step leaching (Pb–LS) (Frei and Kamber, 1995). Sillimanite grains were preliminarily powdered, bulk fractions of the samples weighted 50–70 mg, and the separates taken for leaching weighted 80–100 mg. The sillimanite grains were treated by 6 N HCl at 90°C for six hours to remove the minor admixture of monazite (DeWolf et al., 1996). To minimize the influence of zircon microinclusions, the bulk sample and acid residue were dissolved in concentrated HF mixed with few drops of HNO_3 at 105°C for 48 hours. Pb was extracted on a Pb cation exchanger in HBr. The Pb isotopic composition was analyzed in the bulk sample, acid leachate, and leaching residue. The laboratory blank was less than 0.4 ng for Pb.

The monazite was decomposed and Pb and U were extracted for the analysis following the Krogh technique (Krogh, 1973). The laboratory blanks were less than 0.1 ng for Pb and 0.01 ng for U. The isotopic compositions of U and Pb were measured on a Finnigan MAT-261 mass spectrometer at the Institute of Precambrian Geology and Geochronology, Russian Academy of Sciences. The Pb/U ratio was determined accurate to ±0.5% (2σ). The measured data were processed by the Ludwig program (Ludwig, 1991, 2001).

The composition of rock-forming minerals was studied on an ABT-55 electron microscope equipped with an EDS LINK AN 1000 at the Institute of Precambrian Geology and Geochronology, Russian Academy of Sciences.

RESULTS OF ISOTOPIC INVESTIGATIONS AND INTERPRETATION

The bulk sillimanite fractions of sample B-2000-31 (Table 1, fractions 1 and 2) contain significant amounts of uranogenic and thorogenic Pb. The leachates (fractions 3 and 4) are close to the bulk samples in the $^{208}Pb/^{204}Pb$ ratios and significantly lower in the $^{206}Pb/^{204}Pb$ ratio. The acid residue (fraction 5) is definitely enriched in uranium Pb. The $(Th/U)_{cr}$ calculated for the moment of sillimanite crystallization reflects variations in uranogenic and thorogenic Pb in the five analyzed silliamanite fractions. The $(Th/U)_{cr}$ ratio in leachates are almost three times higher than that in the bulk fractions. This fact might have three explanations.

(1) The low $^{206}Pb/^{204}Pb$ and high $(Th/U)_{cr}$ of 10.4 and 9.5 in the leachates are caused by the dissolution of monazite microinclusions, with Th dominating over U. This assumption is consistent with the fact that the leaching residue (monazite-free) has $(Th/U)_{cr}$ = 1.6,

Table 2. Results of isotopic study of monazite (samples B-2000-31, Fi-03-06)

Characteristics of fraction	Isotopic ratios					Rho	Th/U	Age, Ma			
	$^{206}Pb^{a}/^{204}Pb$	$^{207}Pb^{b}/^{206}Pb$	$^{208}Pb^{b}/^{206}Pb$	$^{206}Pb/^{238}U$	$^{207}Pb/^{235}U$			$^{206}Pb/^{238}U$	$^{207}Pb/^{235}U$	$^{207}Pb/^{206}Pb$	Concordant
Ladoga Complex, B-2000-31											
15 grains	6139	0.11640	6.0920	0.3338	5.256	0.99	16.8	1856.9	1861.8	1867.3 ± 2.1	1860.5 ± 4.4
Sulkava Complex, Fi-03-06											
12 grains	7625	0.10962	1.4900	0.3216	4.860	0.96	4.1	1797.4	1795.4	1793.1 ± 1.7	1794.7 ± 4.6

Note: (a) Isotopic ratios corrected for fractionation and Pb blank; (b) isotopic ratios corrected for fractionation, Pb blank, and common Pb. Th/U ratio was calculated at the moment of monazite crystallization. The method is described in detail in (Rizvanova et al., 2000).

which is two times lower than that in the bulk fractions. Actually, monazite from the studied sample has a high $(Th/U)_{cr}$ ratio of 16.8 (Table 2).

(2) Sillimanite contains only zircon microinclusions. This offers the greatest contribution of Pb from zircon into residue, which is evident from the lowest $(Th/U)_{cr} = 1.6$. Since the applied technique of sillimanite leaching implies no dissolution of zircon microinclusions, the $(Th/U)_{cr}$ ratio in the leachates should reflect that in sillimanite. However, the $(Th/U)_{cr}$ ratios of 10.4 and 9.5 are presumably too high for sillimanite and rather typical of monazite. Therefore, the second assumption seems to be hardly probable.

(3) Sillimanite contains small amounts of both monazite and zircon microinclusions. Heterogeneous distribution of uranogenic and thorogenic Pb in sillimanite is seen from the significant scatter of the data points in the $^{208}Pb/^{204}Pb$–$^{206}Pb/^{204}Pb$ diagram (Fig. 4a). However, MSWD = 0.073 (Fig. 5a) indicates that, if microinclusions of monazite and zircon are present, their U–Th–Pb isotopic systems have been in equilibrium with sillimanite since crystallization.

Optical microscopy and back-scattered electron imaging showed that some grains have a heterogeneous structure caused by specifics of growth and recrystallization. No mineral microinclusions were found in sillimanite.

The level of purity in mineral separates is required to be far better than 0.001%, which cannot be reached by present-day separation techniques (Zhou and Hensen, 1995). Given the required high level of purity and the random distribution of monazite and zircon microinclusions in sillimanite, the preliminary examination of few grains under electron microscope would not rule out the presence of microinclusions of these minerals in the whole analyzed sillimanite separate (DeWolf et al., 1996). Correspondingly, since large aliquots (a few hundred grains) were used for Pb–LS dating, the assumed presence of foreign U- and Th-rich phases in some grains seems to be the most probable.

A five-point Pb–Pb isochron defines an age of 1877.8 ± 6.6 Ma, MSWD = 0.073 (Fig. 5a). Since the mobility of U, Th, and Pb isotopes in mineral microinclusions is defined by their diffusion into the host mineral (Cheong et al., 2000), the age value of 1877.8 ± 6.6 Ma is taken as the closure time of the U–Th–Pb isotopic system of sillimanite (DeWolf et al., 1996).

The sillimanite from sample Fi-03-06 has a significantly less radiogenic Pb composition than that from sample B-2000-31 (Table 1). In addition, the sillimanite from sample Fi-03-06 also has a low $(Th/U)_{cr}$ ratio of 0.9–2.5. However, the variations of $(Th/U)_{cr}$ in the fractions are similar to those in sillimanite from sample B-2000-31. The Pb isotopic ratios in the bulk samples, leachates, and leaching residues of these two sillimanite samples are different. In sillimanite from sample Fi-03-06, the leachates bear the highest content of radiogenic uranogenic Pb, while the leaching residue has the lowest content of radiogenic Pb (Table 1). A five-point Pb–Pb isochron defines an age value of 1799 ± 19 Ma, MSWD = 0.65 (Fig. 5b). A significant scatter of the data points in the $^{208}Pb/^{204}Pb$–$^{206}Pb/^{204}Pb$ diagram (MSWD = 793) presumably points to the presence of two phases with different isotopic compositions of the common Pb and Th/U ratios (Fig. 4b). These phases might be sillimanite and monazite, since the chemical procedures were performed at temperature of 100°C, when zircon is not dissolved. However, if another phase is present, the good correlation of the five data points in the $^{207}Pb/^{204}Pb$–$^{206}Pb/^{204}Pb$ diagram (Fig. 5) indicates that it was in equilibrium with sillimanite during crystallization. The age value of 1799 ± 19 Ma is taken to be the closure time of the U–Th–Pb isotopic system of sillimanite.

The Pb isotopic composition of the tourmaline leachate (sample Fi-03-06) differs in an extremely high $^{206}Pb/^{204}Pb$ ratio, probably testifying to the presence of monazite inclusions. This is also confirmed by the same Th/U ratios in these two minerals (Table 1).

Monazite from garnet–cordierite–sillimanite gneiss (sample B-2000-31) has a concordant age of 1860.5 ±

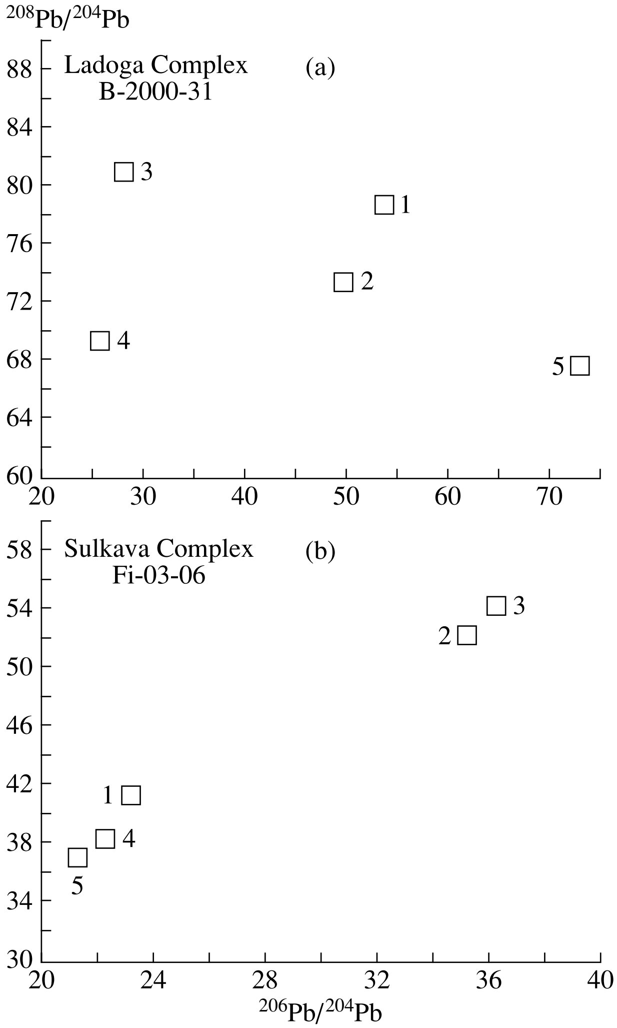

Fig. 4. $^{208}Pb/^{204}Pb$–$^{206}Pb/^{204}Pb$ diagram for sillimanite from the gneisses of (a) the Ladoga and (b) Sulkava complexes.
Numbers near boxes correspond to the numbers of analyzed fractions in Table 1.

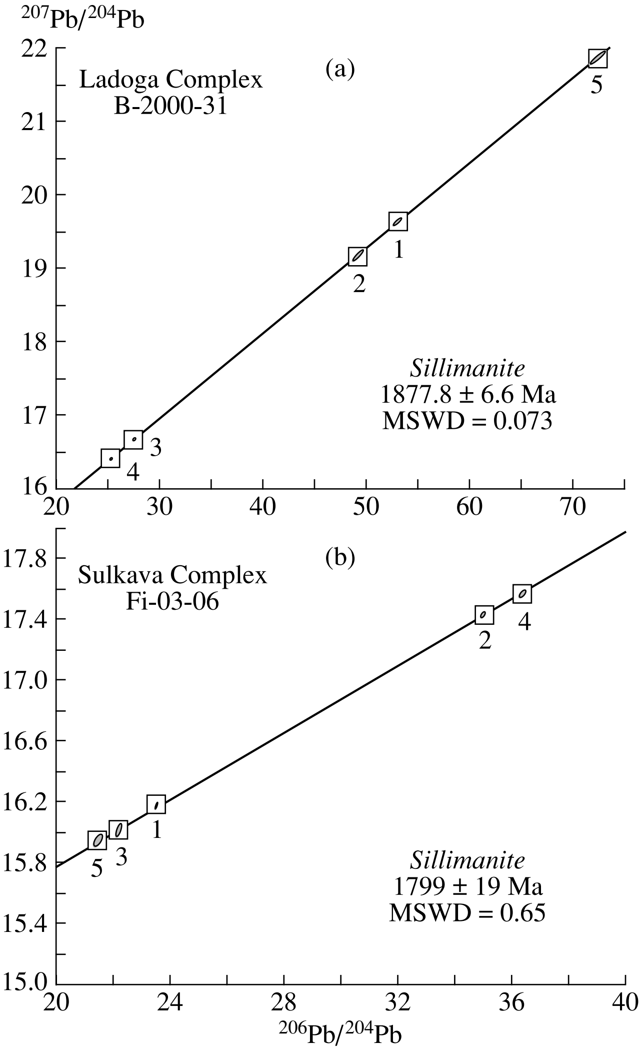

Fig. 5. $^{207}Pb/^{204}Pb$–$^{206}Pb/^{204}Pb$ diagram for sillimanite from the gneisses of the Ladoga (a) and Sulkava (b) complexes.
Numbers near boxes correspond to the numbers of analyzed fractions in Table 1.

4.4 Ma (with probability 0.28 at 95% significance level, Table 2, Fig. 6a). The concordant age (probability 0.39 at 95% confidence level) of monazite from K-feldspar–sillimanite gneiss (sample Fi-03-06) is 1794.7 ± 4.6 Ma (Table 2, Fig. 6b).

DISCUSSION

Our data showed a significant difference in the ages of monazites and sillimanites from gneisses of two complexes. Evidently, the difference (>60 Ma) significantly exceeds the analytical error of the dating methods. Two questions must first to be answered in order to interpret the isotopic data. First, it is necessary to show that the studied sillimanites was formed by the prograde metamorphic stage, because this mineral could, in principle, be also formed during the retrograde stage. Second, it should be explained as to why the ages of sillimanite and monazite are coincident in the Sulkava Complex and differ in the Ladoga Complex. The second question must be solved to determine the monazite position in the metamorphic evolution of the rock.

The studied gneiss from the Sulkava Complex occurs in the sillimanite–K-feldspar zone, i.e., was formed immediately after the decomposition of the quartz–muscovite rocks of the previous zone, thus indicating a prograde formation of sillimanite. Samples from the high-temperature (for example, granulite) zone were not used because of difficulties in determining the sillimanite position in the metamorphic evolution of the high-temperature Sulkava Complex. Thus, the prograde genesis of sillimanite in the Sulkava Com-

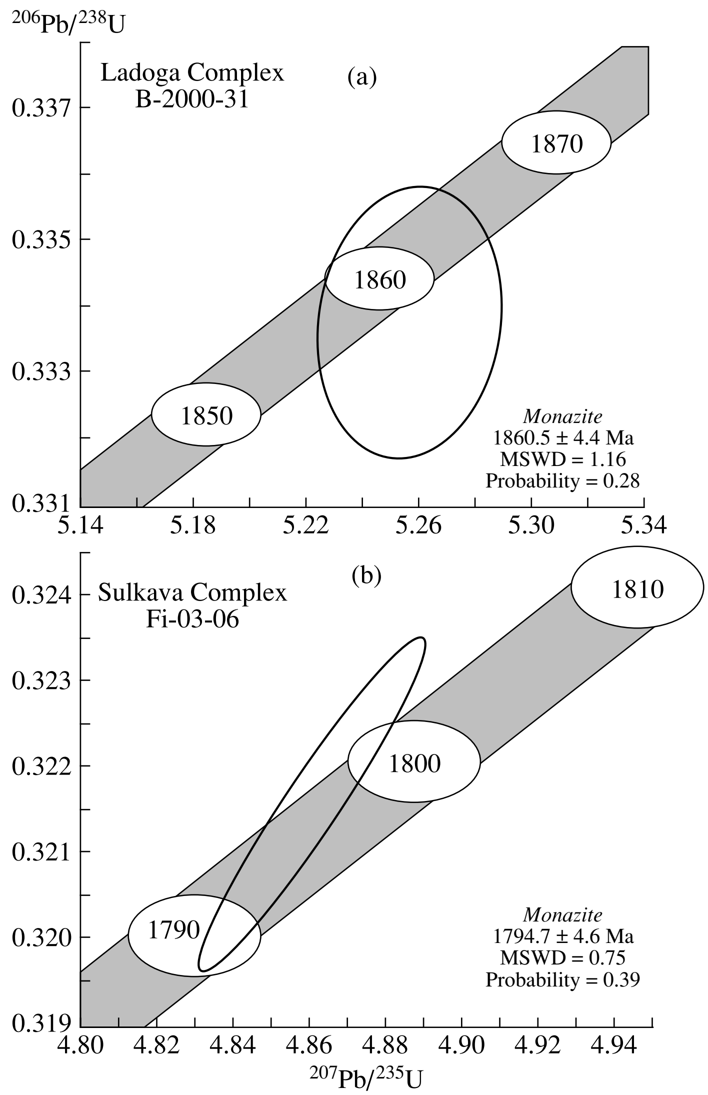

Fig. 6. Concordia diagram for monazites from the gneisses of the Ladoga (a) and Sulkava (b) complexes.

plex is evident from muscovite decomposition by the reaction

$$Ms + Qtz \longrightarrow Sil + Kfs + H_2O\uparrow.$$

This reaction is well confirmed by the petrography (Fig. 7a) and mapping of the prograde series of the metamorphic zones (Korsman, 1977).

The presence of sillimanite inclusions in garnet and cordierite in our sample from the Ladoga area (Figs. 7b, 7c), considered together with the development of K-feldspar in these rocks, suggests the proceeding of the following reactions occurring with increasing temperature (Korikovsky, 1979):

$$Bt + Sil + Qtz \longrightarrow Crd + Kfs + \mathrm{V(L)}\uparrow,$$

$$Bt + Sil + Qtz \longrightarrow Grt + Crd + Kfs + \mathrm{V(L)}\uparrow.$$

These data indicate that sillimanite already existed during prograde stage.

Thus, the position of the sillimanites suggest that their age estimates mark the timing of prograde metamorphism in each of the compared complexes.

Monazite as an informative gechronometer was studied in magmatic (Williams et al., 1983; Copeland et al., 1988) and metamorphic (Smith and Barreiro, 1990; Ross et al., 1991; Spear and Parrish, 1996) rocks. The mineral typically defines concordant or near-concordant ages, reflecting the time of magmatic or metamorphic crystallization, but can also record an inherited age (Copeland et al., 1988). Monazite occurs as detrital grains in sediments (Parrish, 1990; Ross et al., 1991), more often in aluminous rocks, and is rare in calcic rocks, which contain allanite, apatite, and titanite in place of this mineral (Parrish, 1990). Monazite is unstable during diagenesis and low-temperature metamorphism (Sawka et al., 1986; Smith and Barreiro, 1990). At low temperatures, it is decomposed into hydrophos-

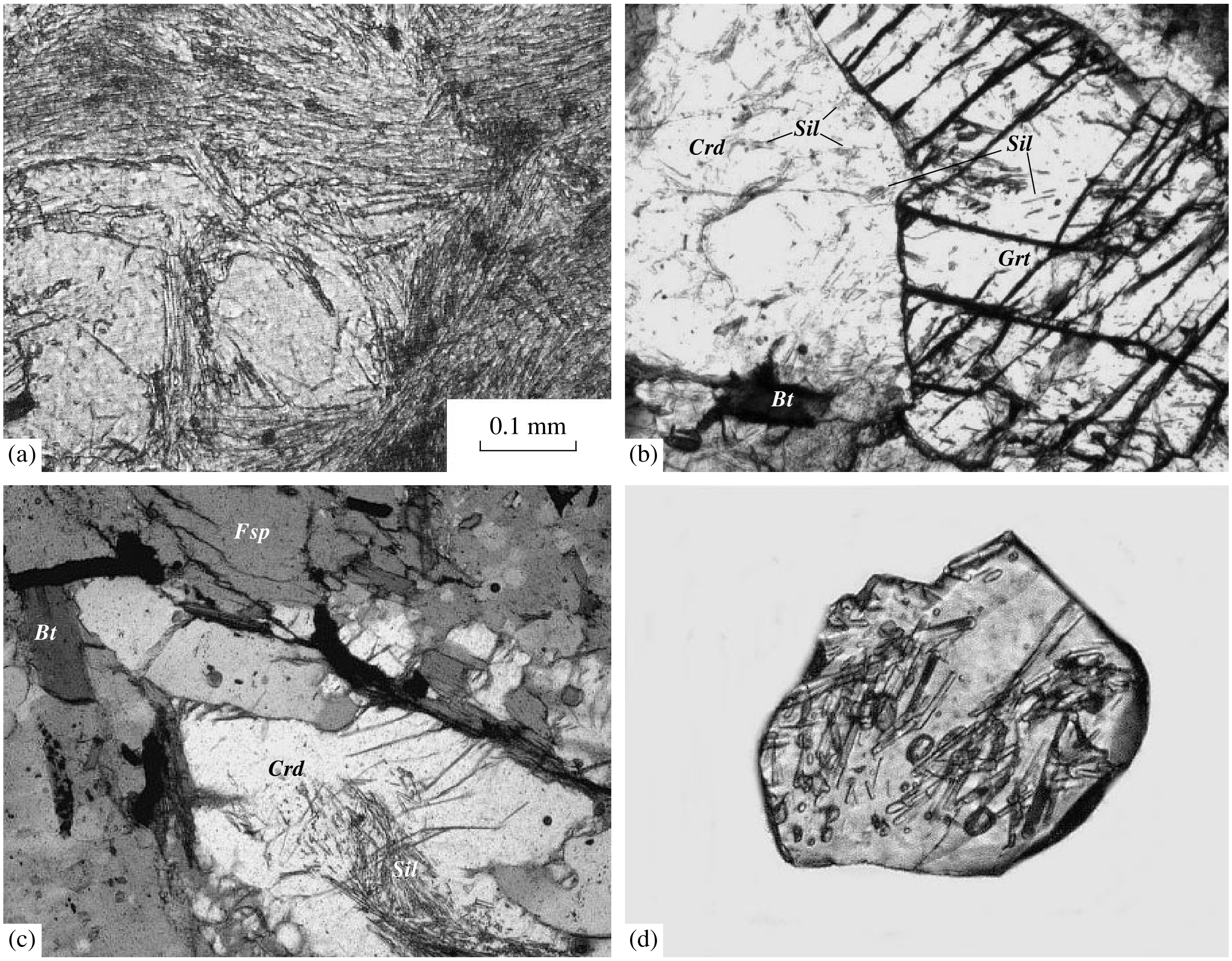

Fig. 7. Photographs of the polished thin sections of the studied high-Al gneisses of the Sulkava and Ladoga complexes.
(a) Sillimanite–K-feldspar aggregates in high-Al gneisses from the Sulkava complex; (b–d) fine sillimanite needles in cordierite and garnet (b), cordierite (c), and monazite (d) from the high-Al gneiss of the Ladoga Complex.

phates, which release hydroxyl groups with increasing temperature and are again transformed into monazite (Sawka et al., 1986; Smith and Barreiro, 1990). This restricts the use of monazite for deciphering the sedimentary and diagenetic stages of rock formation. Hence, monazite in high-temperature metamorphic rocks presumably records metamorphic crystallization. It was shown (Pyle and Spear, 2003) that a metamorphic rock can contain several monazite generations with geochemical signatures reflecting certain metamorphic processes. Geochronological investigations sometimes showed the presence of monazite crystals of different ages in the same rock (Hermann and Rubatto, 2003).

The closure temperature of the U–Pb system in monazite is not known exactly. Based on the correlation of monazite dates and *P–T* evolution of the rocks, the closure temperature was roughly estimated at 700°C (Copeland et al., 1988), which is about 100°C lower than that of zircon (Rubatto et al., 2001). These estimates are well consistent with our empirical data (Baltybaev et al., 2004a).

The experimental study of Pb diffusion in monazite showed that the closure temperature of the U–Pb system in slowly cooling lithospheric blocks (1–10°C/Ma) depends on the grain size and lies within 550–700°C (Smith and Giletti, 1997). Some works point to a higher closure temperature, up to 800°C (Spear and Parrish, 1996).

With regard to the aforesaid data, we arrived at the conclusion that the U–Pb age of monazite in high-temperature rocks presumably recorded metamorphic crystallization or recrystallization at temperature 700°C and more. Under less favorable conditions for the U–Pb isotopic system (small grain sizes, very slow cooling, high fluid activity), the age of initially high-temperature monazite corresponds to the closure of the system during the retrograde stage, reflecting "rejuvenation"

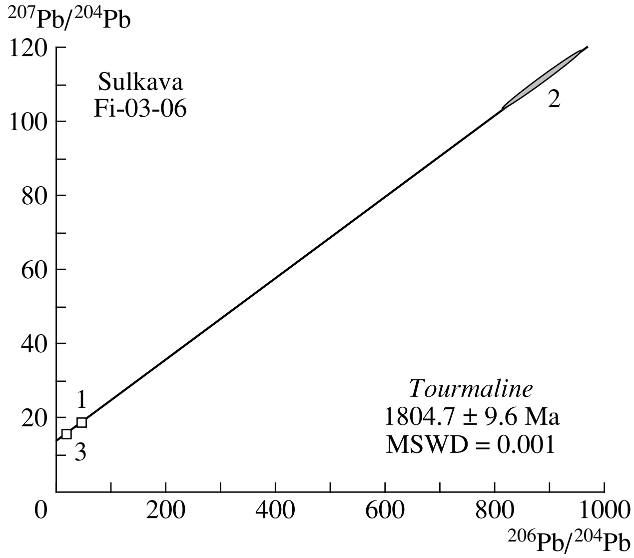

Fig. 8. $^{207}Pb/^{204}Pb$–$^{206}Pb/^{204}Pb$ diagram for tourmaline from the high-Al gneiss of the Sulkava complex.
The numbers near the boxes and ellipse correspond to the numbers of analyzed fractions in Table 1.

owing to diffusion losses or fluid–thermal influence. Then, the age of the high-temperature (amphibolite or granulite facies) monazite presumably corresponds to the retrograde stage, the facies of two-mica gneisses or even a lower temperature. The rejuvenation problem for low- and medium-temperature monazites with crystallization temperature lower than the closure temperature of the U–Pb system is less acute. Rather, the sample should be tested for the presence of an old component, such as relict grains, protocores or other parts of the crystals.

Such a behavior of the U–Pb system of monazite suggests that the age difference between monazites from the two compared complexes could not result from the different evolution of initially coeval monazites. Conversely, our high-temperature rock contains older monazite than that in the low- and medium-temperature gneiss. The age of the monazite from the Sulkava Complex can hardly correspond to the timing of the local fluid reworking. There is no corresponding geological and mineralogial–petrologial evidence. This is also inconsistent with the finding of monazites of similar age in rocks from other zones of this complex (Korsman et al., 1984; Vaasjoki and Sakko, 1988). It should be added that tourmaline (Table 1, Fig. 8) from the same sample (Fi-03-06) appeared to have an age similar to those of the sillimanite and monazite. Naturally, the monazite from the Ladoga gneiss is younger than sillimanite from this sample, as follows from monazite overgrowths on sillimanite (Fig. 7d). It is difficult to determine whether the obtained age corresponds to the crystallization time or recorded the later closure of the U–Pb system. The dates on monazite from other rocks of the Ladoga region fall into a narrow range of 1876–1850 Ma, indicating that these events were perhaps temporally close (Baltybaev et al., 2004a).

Thus, the studied metamorphic complexes ascribed to the inner and outer Svecofennian zones differ in both the composition and the age of prograde metamorphism.

Svecofennides of the outer zone border the Sulkava Complex in the north (Korsman et al., 1984, 1999; Hölttä, 1988), with the Kolkonjärvi lineament as the juncture zone (Korsman et al., 1984). In the considered region of Finland, the junction zone between the outer and inner Sveconfennian zones is represented by suture complicated by folds and faults. The time of fault initiation can be inferred from (Vaasjoki and Kontoniemi, 1991). These authors showed that the intrusions, which are cut by the faults and have similar monazite age of ~1800 Ma (Vaasjoki and Sakko, 1988), define different U–Pb zircon dates: 1887 ± 5 Ma for the Osikonmäki Massif, 1850 ± 7 Ma for the Hiltula Massif, and 1815 ± 2 Ma for the Pirilä Massif. This leads to the conclusion that the U–Pb system of monazites from the intrusions was equilibrated later, during the initiation of the fault zone, which presumably formed during Late Svecofennian tectonothermal events. The high-temperature rocks of the Southern domain of the Ladoga area presumably show the same tectonic relations with adjacent rocks of the Sulkava metamorphic complex west of the area. The junction zone is very poorly exposed. Geophysical data confirm the existence of a structural–tectonic disconformity. For example, there is a sharp gravity step along the boundary band, while magnetic anomalies extend northeast, parallel to the state boundaries.

Thus, the Sulkava thermal structure (Korsman, 1977; Korsman et al., 1984, 1988) pertains to the Late Svecofennian granite–migmatite zone (Ehlers et al., 1993) (or inner zone, according to our terms), which is confirmed by the structure, composition, and our isotopic data. This zone extends through southern Finland to Sweden, where it is made up of metamorphic complexes locally reaching the granulite facies. Judging from numerous data (Väisänen et al., 2002, 2004; Andersson et al., 2004; present paper), the rocks of this zone underwent *Late Svecofennian* metamorphism at 1.83–1.79 Ga. The Svecofennian block of the Ladoga area (Southern domain) is ascribed to the outer Svecofennian zone consolidated at ~1.86–1.86 Ga (Baltybaev et al., 2004a, present paper). The outer Svecofennian zone also experienced metamorphism to the granulite facies, with the metamorphic peak reliably dated at 1.89–1.87 Ga (Korsman et al., 1984, 1988, 1999; Huhma, 1986; Hölttä, 1988; Vaasjoki, Sakko, 1988; Mouri et al., 1999, etc.). With regard for our data, this metamorphic stage can be regarded as the *Early Svecofennian* stage.

The two stages of endogenous activity, which occurred in the Svecofennides of the outer and inner zones and defined the lateral heterogeneity of the Svecofennian domain, require revision of the concepts suggesting the formation of the latter during a single stage of Svecofennian *orogeny*. This is of crucial importance, since the age gap between the indicated stages (from ~1.86 to ~1.83 Ga) was marked by a drastic decrease in endogenous activity throughout the Svecofennian domain. During this period, the volcanic activity was absent, while plutonic activity was expressed in the formation of single intrusive massifs within the Svecofennides of Finland and Russia. In the Svecofennides of Sweden, the magmatic activity also was reduced during this period, but later, after 1.83–1.84 Ga, it was resumed to form the Transscandinavian igneous belt (Andersson, 1991 etc.).

CONCLUSIONS

(1) The Ladoga and Sulkava metamorphic complexes represent, respectively, the inner and outer zones of the Svecofennian domain, which differ in composition and crystallization time of rock-forming and accessory minerals.

(2) The differences between the metamorphic mineral assemblages of the two studied complexes were predetermined by the lithochemical composition of the sedimentary protoliths: higher Al_2O_3 and K_2O contents in the rocks of the Sulkava metamorphic complex than those of the Ladoga Complex.

(3) The difference in crystallization time of the minerals of the compared complexes was caused by the regional occurrence of the Early Svecofennian (1.89–1.86 Ga) and Late Svecofennian (1.83–1.79 Ga) metamorphic stages in the Svecofennian domain. The age gap between these stages is characterized by a relative decrease in endogenous activity.

(4) Combined dating of rock-forming and accessory minerals in metamorphic rocks and petrological observations make it possible to constrain the position of the dated minerals in the metamorphic evolution of the rocks, thus placing reliable constraints on the timing of metamorphism.

ACKNOWLEDGMENTS

We thank S. Z. Yakovleva, A. F. Makeev, and N. I. Ro (Institute of Precambrian Geology and Geochronology, Russian Academy of Sciences) for help in isotopic investigations of minerals. We are grateful to S. P. Korikovsky and I. V. Chernyshev (Institute of Geology of Ore Deposits, Petrography, Mineralogy, and Geochemistry, Russian Academy of Sciences) for critical reviewing of the manuscript. This study was supported by the Russian Foundation for Basic Research (project nos. 02-05-65343, 03-05-64779, 04-05-64856, 05-05-65308) and the Federal Program for the Support of the Leading Scientific Schools (project no. 615.2003.05), Program of the Division of Earth Sciences of RAS (project no. 7), and the NorFA sponsored Nordic network “Transition from Orogenic to Anorogenic Magmatism in the Fennoscandian Shield.”

REFERENCES

1. U. B. Andersson, “Granitoid Episodes and Mafic-Felsic Magma Interaction in the Svecofennian of the Fennoscandian Shield, with Main Emphasis on the ~1.8 Ga Plutonics,” Precambrian Res. **51**, 127–149 (1991).
2. U. B. Andersson, K. Högdahl, H. Sjöström, et al., “Magmatic, Detrital, and Metamorphic Ages in Metamorphic Rocks from South–Central Sweden,” in *Proceedings of 26th Nordic Geological Wintermeeting, Uppsala, Sweden, 2004* (Geologiska Föreningens Stockholm Förhandlingar, Uppsala, 2004), Vol. 126, pp. 16–17.
3. Sh. K. Baltybaev and O. A. Levchenkov, “Volcanics in Svecofennides of the Ladoga Region and Results of U–Pb and Pb–Pb Dating of Rocks and Minerals as a Basis for Correlation of Svecofennian Events,” Stratigr. Geol. Korrelyatsiya **13** (2), 3–19 (2005) [Stratigr. Geol. Correlation **13** (2), 119 (2005)].
4. Sh. K. Baltybaev, E. B. Sal’nikova, V. A. Glebovitsky, et al., “The Kuznechensky Massif of Potassic Porphyritic Granites: Results of U–Pb Dating and Substantiation of the Tectonic Position (Baltic Shield),” Dokl. Akad. Nauk **398** (4), 519–523 (2004b) [Dokl. Earth. Sci. **398** (7), 1024 (2004b)].
5. Sh. K. Baltybaev, O. A. Levchenkov, N. G. Berezhnaya, et al., “Age and Duration of Svecofennian Plutono-Metamorphic Activity in the Ladoga Area, Southeastern Baltic Shield,” Petrologiya **12** (4), 373–392 (2004a) [Petrology **12** (4), 330 (2004a)].
6. Sh. K. Baltybaev, V. A. Glebovitskii, I. V. Kozyreva, et al., “The Meyeri Thrust: The Main Element of the Suture at the Boundary between the Karelian Craton and the Svecofennian Domain in the Ladoga Region of the Baltic Shield,” Dokl. Akad. Nauk **348** (3), 353–356 (1996) [Dokl. Earth Sci. **347** (4), 581 (1996)].
7. Sh. K. Baltybaev, V. A. Glebovitskii, I. V. Kozyreva, et al., *Geology and Petrology of the Svecofennides in the Ladoga Area* (S.-Peterb. Gos. Univ., St. Petersburg, 2000) [in Russian].
8. Sh. K. Baltybaev, V. A. Glebovitskii, O. A. Levchenkov, et al., “Age Relationship between Potassic and Sodic Migmatites of Svecofennides, Ladoga Region, Baltic Shield,” Dokl. Akad. Nauk **383** (4), 523–526 (2002) [Dokl. Earth. Sci. **383A** (3), 292 (2002)].
9. C. S. Cheong, S.-T. Kwon, and K.-H. Park, “Pb and U Isotope Constraints on Paleoproterozoic Crustal Evolution of the Northern Yeongnam Massif, South Korea: Tectonic Implication,” Precambrian Res. **102**, 207–220 (2000).
10. P. Copeland, R. R. Parrish, and T. M. Harrison, “Identification of Inherited Radiogenic Pb in Monazite and Implications for U–Pb Systematics,” Nature **333**, 760–763 (1988).
11. J. S. Daly, V. V. Balagansky, M. J. Timmerman, et al., “Ion Microprobe U–Pb Zircon Geochronology and Isotopic Evidence for a Transcrustal Suture in the Lapland–Kola Orogen, Northern Fennoscandian Shield,” Precambrian Res. **105**, 289–314 (2001).

12. C. P. DeWolf, C. J. Zeissler, A. N. Halliday, et al., "The Role of Inclusions in U–Pb and Sm–Nd Garnet Geochronology: Stepwise Dissolution Experiments and Trace Uranium Mapping by Fission Track Analysis," Geochim. Cosmochim. Acta **60**, 121–134 (1996).

13. C. Ehlers, A. Lindroos, and O. Selonen, "The Late Svecofennian Granite–Migmatite Zone of Southern Finland—A Belt of Transpressive Deformation and Granite Emplacement," Precambrian Res. **64**, 295–309 (1993).

14. P. E. Eskola, "The Problem of Mantled Gneiss Domes," Geol. Soc. London Quart. J. **104** (4), 461–476 (1949).

15. R. Frei and B. S. Kamber, "Single Mineral Pb–Pb Dating," Earth Planet. Sci. Lett. **129**, 261–268 (1995).

16. G. Gaál and R. Gorbatschev, "An Outline of the Precambrian Evolution of the Baltic Shield," Precambrian Res. **35** (1), 15–25 (1987).

17. *Geological Evolution of the Deep-Seated Zones of Mobile Belts, Northern Ladoga Region*, Ed. by N. G. Sudovikov, V. A. Glebovitskii, A. S. Sergeev, et al. (Nauka, Leningrad, 1970) [in Russian].

18. V. A. Glebovitskii, "Tectonics and Regional Metamorphism of the Early Precambrian in the Eastern Part of the Baltic Shield," Reg. Geol. Metallogeny, No. 1, pp. 7–37 (1993).

19. R. Gorbatschev and S. Bogdanova, "Frontiers in the Baltic Shield," Precambrian Res. **64**, 3–21 (1993).

20. J. Hermann and D. Rubatto, "Relating Zircon and Monazite Domains to Garnet Growth Zones: Age and Duration of Granulite Facies Metamorphism in the Val Malenco Lower Crust," J. Metamorph. Geol. **21**, 833–852 (2003).

21. P. Hölttä, "Metamorphic Zones and the Evolution of Granulite Grade Metamorphism in the Early Proterozoic Pielavesi Area, Central Finland," Bull. Geol. Soc. Finl. **344** (1988).

22. H. Huhma, "Sm–Nd, U–Pb and Pb–Pb Isotopic Evidence for the Origin of the Early Proterozoic Svecokarelian Crust in Finland," Bull. Geol. Surv. Finl. **337** (1986).

23. T. J. Koistinen, "Structural Evolution of an Early Proterozoic Strata-Bound Cu–Co–Zn Deposit, Outokumpu, Finland," Trans. R. Soc. Edinb. Earth Sci. **72**, 115–158 (1981).

24. A. Kontinen, "An Early Proterozoic Ophiolite—The Jormua Mafic-Ultramafic Complex, Northeastern Finland," Precambrian Res. **35**, 313–341 (1987).

25. S. P. Korikovsky, *Metamorphic Facies of Metapelites* (Nauka, Moscow, 1979) [in Russian].

26. T. Korja, "Electrical Conductivity Distribution of the Lithosphere in the Central Fennoscandian Shield," Precambrian Res. **64**, 85–108 (1993).

27. K. Korsman, "Progressive Metamorphism of Metapelites in the Rantasalmi–Sulkava Area, Southeastern Finland," Bull. Geol. Surv. Finl. **290** (1977).

28. K. Korsman, P. Hölttä, T. Hautala, et al., "Metamorphism as an Indicator of Evolution and Structure of Crust in Eastern Finland," Bull. Geol. Surv. Finl. **328** (1984).

29. K. Korsman, R. Niemela, and P. Wasenius, "Multistage Evolution of the Proterozoic Crust in the Savo Schist Belt, Eastern Finland," Bull. Geol. Surv. Finl. **343**, 89–96 (1988).

30. K. Korsman, T. Koistinen, J. Kohonen, M. Wennerström, et al., *Bedrock Map of Finland 1 : 100000*; Geol. Surv. Finl. (Espoo, Finland, 1997).

31. K. Korsman, T. Korja, M. Pajunen, and GGT/SVEKA Working Group, "The GGT / SVEKA Transect: Structure and Evolution of the Continental Crust in the Paleoproterozoic Svecofennian Orogen in Finland," Int. Geol. Rev. **41**, 287–333 (1999).

32. T. E. Krogh, "A Low-Contamination Method for Hydrothermal Decomposition of Zircon and Extraction U and Pb for Isotopic Age Determinations," Geochim. Cosmochim. Acta **37**, 485–494 (1973).

33. R. Lahtinen, H. Huhma, and J. Kousa, "Contrasting Source Components of the Paleoproterozoic Svecofennian Metasediments: Detrital Zircon U–Pb, Sm-Nd and Geochemical Data," Precambrian Res. **116**, 81–109 (2002).

34. K. R. Ludwig, "PbDat 1.21 for MS-DOS: A Computer Program for IBM-PC Compatibles for Processing Raw Pb–U–Th Isotope Data. Version 1.07," U.S. Geol. Surv. Open-File Rept. 88-542 (1991).

35. K. R. Ludwig, "Isoplot/Ex Rev. 2.49. A Geochronological Toolkit for Microsoft Excel", Berkeley Geochronology Center, Spec. Publ., No. 1a, 2001.

36. *Migmatization and Granite Formation in Different Thermodynamic Regimes* (Nauka, Leningrad, 1985) [in Russian].

37. H. Mouri, K. Korsman, and H. Huhma, "Tectono-Metamorphic Evolution and Timing of the Melting Processes in the Svecofennian Tonalite–Trondhjemite Migmatite Belt: An Example from Luopioinen, Tampere Area, Southern Finland," Bull. Geol. Soc. Finl. **71**, 31–56 (1999).

38. Yu. V. Nagaitsev, *Petrology of the Metamorphic Rocks of the Ladoga and Belomorian Complexes* (Leningr. Gos. Univ., Leningrad, 1974) [in Russian].

39. M. Nironen, "The Svecofennian Orogen: a Tectonic Model," Precambrian Res. **86** (1–2), 21–44 (1997).

40. R. R. Parrish, "U–Pb Dating of Monazite and Its Application to Geological Problems," Can. J. Earth Sci. **27**, 1431–1450 (1990).

41. P. Peltonen, A. Kontinen, and H. Huhma, "Petrology and Geochemistry of Metabasalts from the 1.95 Ga Jormua Ophiolite, Northeastern Finland," J. Petrol. **37**, 1359–1383 (1996).

42. A. A. Predovskii, V. P. Petrov, and O. A. Belyaev, *Geochemistry of Ore Elements in the Precambrian Metamorphic Series: Evidence from the Northern Ladoga Region* (Nauka, Leningrad, 1967) [in Russian].

43. J. M. Pyle and F. M. Spear, "Four Generations of Accessory-Phase Growth in Low-Pressure Migmatites from SW New Hampshire," Am. Mineral. **88**, 338–351 (2003).

44. O. T. Rämö, M. Vaasjoki, I. Manttari, et al., "Petrogenesis of the Post-Kinematic Magmatism of the Central Finland Granitoid Complex: I. Radiogenic Isotope Constraints and Implications for Crustal Evolution," J. Petrol. **42**, 1971–1993 (2001).

45. N. G. Rizvanova, O. A. Levchenkov, A. E. Belous, et al., "Zircon Reaction and Stability of the U–Pb Isotope System during Interaction with Carbonate Fluid: Experi-

mental Hydrothermal Study," Contrib. Mineral. Petrol. **139**, 101–114 (2000).

46. G. M. Ross, R. R. Parrish, and F. Ö. Dudås, "Provenance of the Bonner Formation (Belt Supergroup), Montana: Insights from U–Pb and Sm–Nd Analyses of Detrital Minerals," Geology **19**, 340–343 (1991).
47. D. Rubatto, I. S. Williams, and I. S. Buick, "Zircon and Monazite Response to Prograde Metamorphism in the Reynolds Range, Central Australia," Contrib. Mineral. Petrol. **140**, 458–468 (2001).
48. W. Sawka, J. F. Banfield, and B. W. Chappell, "A Weathering-Related Origin of Widespread Monazite in S-Type Granites," Geochim. Cosmochim. Acta **50**, 171–175 (1986).
49. V. I. Shul'diner, I. V. Kozyreva, and Sh. K. Baltybaev, "Geochronology and Formation Subdivisions of the Lower Precambrian in the Northwestern Ladoga Region," Stratigr. Geol. Korrelyatsiya **4** (3), 11–22 (1996) [Stratigr. Geol. Correlation **4** (3), 220 (1996)].
50. V. I. Shul'diner, O. A. Levchenkov, S. Z. Yakovleva, et al., "The Late Karelian in the Stratigraphic Scale of Russia: Determination of Its Lower Boundary and Regional Units in the Stratotype Area," Stratigr. Geol. Korrelyatsiya **8** (6), 20–33 (2000) [Stratigr. Geol. Correlation **8** (6), 544 (2000)].
51. V. I. Shul'diner, Sh. K. Baltybaev, and I. V. Kozyreva, "Metamorphic Evolution of Garnet-bearing Granulites in the Western Ladoga Area," Petrologiya **5** (3), 253–277 (1997) [Petrology **5** (3), 223 (1997)].
52. H. A. Smith and B. Barreiro, "Monazite U–Pb Dating of Staurolite Grade Metamorphism in Pelitic Schists," Contrib. Mineral. Petrol. **105**, 602–615 (1990).
53. H. A. Smith and B. J. Giletti, "Lead Diffusion in Monazite," Geochim. Cosmochim. Acta **61**, 1047–1055 (1997).
54. F. S. Spear and R. R. Parrish, "Petrology and Cooling Rates of the Valhalla Complex, British Columbia, Canada," J. Petrol. **37**, 735–765 (1996).
55. A. I. Tugarinov and E. V. Bibikova, *Geochronology of the Baltic Shield According to Zircon Isotopic Dating* (Nauka, Moscow, 1980) [in Russian].
56. M. Vaasjoki and M. Sakko, "The Evolution of the Raahe–Ladoga Zone in Finland: Isotopic Constraints," Bull. Geol. Surv. Finl. **343**, 7–32 (1988).
57. M. Vaasjoki and O. Kontoniemi, "Isotopic Studies of the Proterozoic Osikonmaki Gold Prospect at Rantasalmi, Southeastern Finland," Geol. Surv. Finland. Spec. Pap. **12**, 53–57 (1991).
58. M. Väisänen, I. Manttari, and P. Hölttä, "Svecofennian Magmatic and Metamorphic Evolution in Southwestern Finland As Revealed by U–Pb Zircon SIMS Geochronology," Precambrian Res. **116**, 111–127 (2002).
59. M. Väisänen, U. B. Andersson, H. Huhma, et al., "Age of Late Svecofennian Regional Metamorphism in Southern Finland and South–Central Sweden", in *Proceedings of 26th Nordic Geological Wintermeeting, Uppsala, Sweden, 2004* (Geologiska Föreningens Stockholm Förhandlingar, Uppsala, 2004), Vol. 126, pp. 40–41.
60. D. A. Velikoslavinskii, *Comparative Characteristics of the Medium- and Low-Pressure Regional Metamorphism* (Nauka, Leningrad, 1972) [in Russian].
61. I. S. Williams, W. Compston, and B. W. Chappell, "Zircon and Monazite U–Pb Systems and the Histories of I-Type Magmas, Berridale Batholith, Australia," J. Petrol. **24**, 76–97 (1983).
62. B. Zhou and B. J. Hensen, "Inherited Sm/Nd Isotope Components Preserved in Monazite Inclusions within Garnets in Leucogneiss from East Antarctica and Implications for Closure Temperature Studies," Chem. Geol. **121**, 317–326 (1995).