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Paleozoic Rocks in Infrastructure of the Metamorphic Core, the Greater Caucasus Main Range Zone

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Abstract—Granitoid orthogneisses and migmatites are widespread in the lower, deeply metamorphosed gneiss–migmatite complex of the pre-Alpine basement (infrastructure) exposed within northern part of the Greater Caucasus Main Range zone. Like the other rocks of the complex, they have been traditionally attributed to the Proterozoic, but the U–Pb dating revealed the Late Paleozoic age of migmatites and Devonian age of orthogneiss protolith. Bodies of blastomylonitic apogranite gneisses, which are confined to boundary between gneiss-migmatite complex and overlying Makera Complex of supracrustal rocks, turned out to be of the Late Paleozoic age as well. The dating results suggest synchronism and, apparently, genetic interrelations between the high-T/low-P metamorphism and granite formation in the Main Range zone of the Greater Caucasus.

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Key words: Greater Caucasus, U–Pb dating, orthogneisses, migmatites, Paleozoic.

INTRODUCTION

The deeply eroded part of the Greater Caucasus fold structure corresponds to the Main Range zone, where metamorphic rocks and granitoids of the pre-Alpine basement are exposed at the surface (Fig. 1). Two metamorphic complexes are distinguished in the basement of the northern (El'brus) subzone. The Makera Complex of supracrustal rocks is represented by metapelites and, to a lesser extent, by gneisses of epidote-amphibolite metamorphic grade, while the Gondarai Complex of lower gneisses, metapelites and migmatites of high-T amphibolite facies is attributed to infrastructure (Baranov, 1987; Grekov and Lavrishchev, 2002).

One of most intriguing problems in geology of the Greater Caucasus concerns the rock ages in both complexes and their structural and geochronological relations. According to most widespread opinion, both subdivisions are of Proterozoic age (Baranov, 1987; Grekov and Lavrishchev, 2002, Gamkrelidze and Shengelia, 2005), although it is unclear whether this means the age of protoliths for metamorphic rocks or the time of metamorphism as well. As is suggested in most recent monographic synthesis (Gamkrelidze and Shengelia, 2005), the progressive high-T metamorphism of gneiss-migmatite complex took place in the Grenvillian epoch or even earlier, being followed by diaphthoresis of biotite-sillimanite gneiss facies in the Late Caledonian epoch, while the Hercynian tectonic events were responsible only for retrograde metamorphism of greenschist facies. Lebed'ko et al. (2002) also attributed metamorphic rocks of the Main Range zone to the Precambrian. In opinion of Potapenko (1982) that is lacking desirable substantiation however, the Precambrian was a time of the greenschist facies metamorphism only, and high-T metamorphic transformations, migmatization included, developed in the Hercynian epoch. These and some other ideas about age of metamorphic rocks in the Main Range zone are based on old, now doubtful isotopic–geochronological information, being related to misleading concept, still popular among some geologists, that age of metamorphic rocks is proportional to their metamorphic grade. On the other hand, the viewpoints under consideration are engendered by geological data on the Bechasyn zone of the Greater Caucasus extrapolated to the Main Range zone without serious substantiation. In the Bechasyn zone, metamorphic rocks are transgressively overlapped by terrigenous-carbonate sedimentary cover spanning stratigraphic interval of the Vendian–Cambrian (?)–Silurian, as was suggested (Potapenko, 1982). Presence of rocks of the same lithologic composition has not been established with confidence in the Main Range zone, which is separated in addition from the Bechasyn zone by the Fore Range zone with high-P metamorphic rocks in crystallinicum of very peculiar ensimatic type (*Petrology of Metamorphic...*, 1991; Somin, 1997). As it was established recently, metamorphic igneous rocks of crystallinicum are of the Middle Paleozoic age (Somin and Lavrishchev, 2005), and cor-

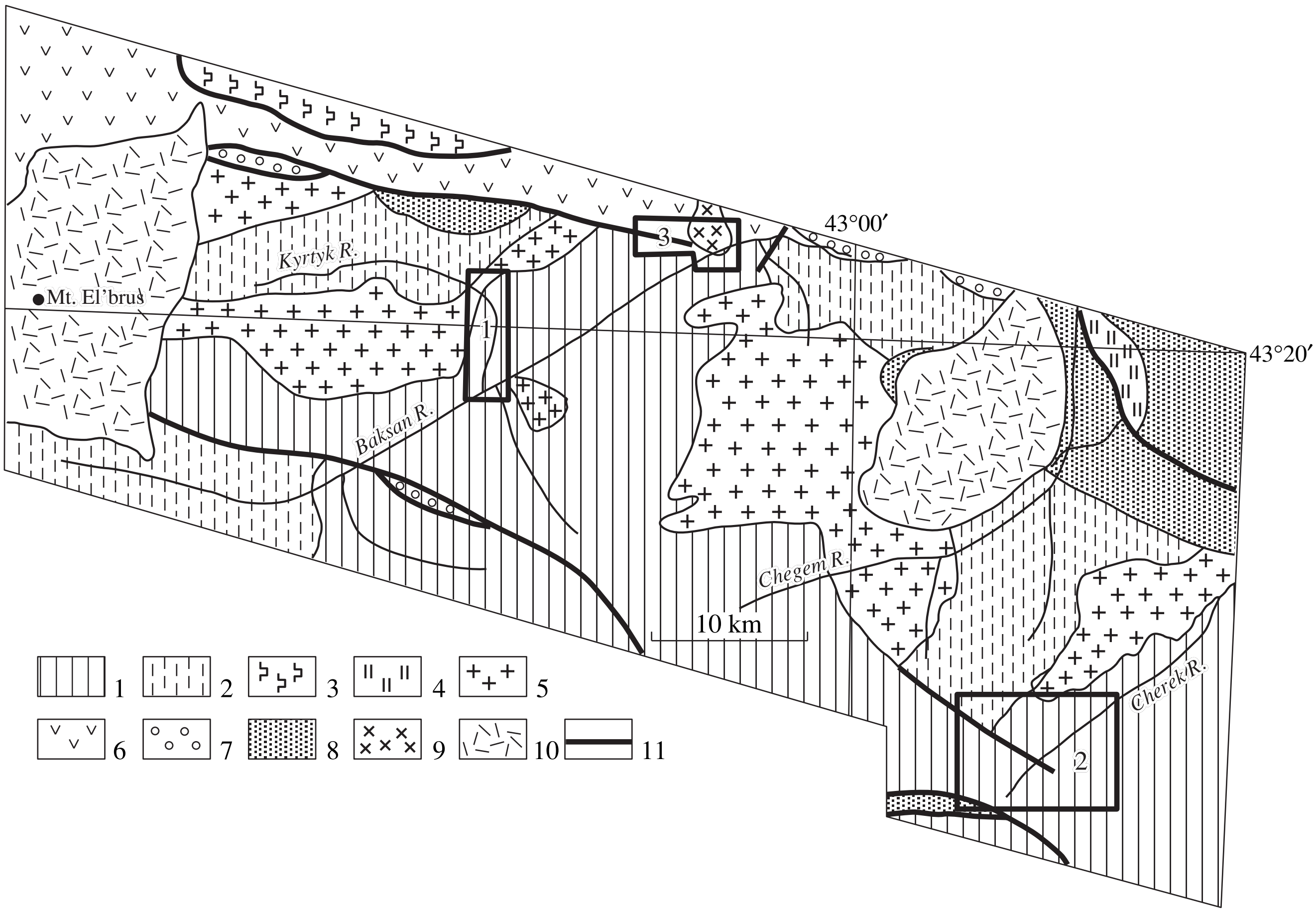

Fig. 1. Geological scheme of the Baksan–Cherek Bezengiiskii interfluve area. Complexes: (1) Gondarai gneiss-migmatite, (2) Makera, (3) Bechasyn and (4) Blyb–Labardan complexes; (5) granites of the Main Range zone; (6) Middle Paleozoic volcanogenic-sedimentary deposits of the Fore Range; (7) Upper Paleozoic molasses; (8) Lower–Middle Jurassic deposits; (9) El'dzhurta granite; (10) volcanics of the El'brus and Chegem upland; (11) major faults; numbered in the figure are the Kyrtyk (1), Mukulan (2) and Ulluchiran (3) study areas.

relation between metamorphites of the Main Range and Bechasyn zones should be proved by independent geochronological dating. Results of investigations in this direction, i.e., the U–Pb dates obtained for orthogneisses and migmatites from the gneiss-migmatite complex of the El'brus subzone and for blastomylonitic apogranite orthogneisses at the boundary of this complex with the Makera Complex are discussed in this work.

GEOLOGICAL SITUATION AND OBJECTS OF GEOCHRONOLOGICAL STUDY

In the El'brus subzone, metamorphic rocks of gneiss-migmatite complex crop out from below the Makera Complex, being separated from the latter by a slightly inclined zone of tectonic displacement that is of unclear origin and by bodies of Late Paleozoic granites (Fig. 1). Rocks of gneiss-migmatite complex are represented by para- and orthogneisses, metapelites, and subordinate metabasites. Metamorphic grade of all the rocks corresponds to low-P amphibolite facies. A complete petrological characterization of the complex has been expounded earlier (*Petrology of Metamorphic…,* 1991; Gamkrelidze and Shengelia, 2005).

Orthogneisses studied in this work are remarkable and widespread rocks of gneiss-migmatite complex, although they have been distinguished in the complex composition not long ago (Bibikova et al., 1991; Hanel et al., 1993), being taken before for the ghost migmatites. These are banded to lenticular-banded rocks corresponding in composition to granite, granodiorite, or plagiogranite sometimes. Their tabular bodies are from a few to hundreds meters thick. Orthogneisses with granoblastic microstructure contain sometimes the relics of zoned magmatic plagioclase. Xenoliths of host rocks (paragneisses and metabasites) occasionally occur in marginal zones of large orthogneiss bodies.

Outcrops of orthogneisses have been studied in canyons of the Adyl-su and Kyrtyk rivers and in the Bol'shoi Mukulan Gorge. Bibikova et al. (1991) who already dated orthogneisses from the Adyl-su canyon

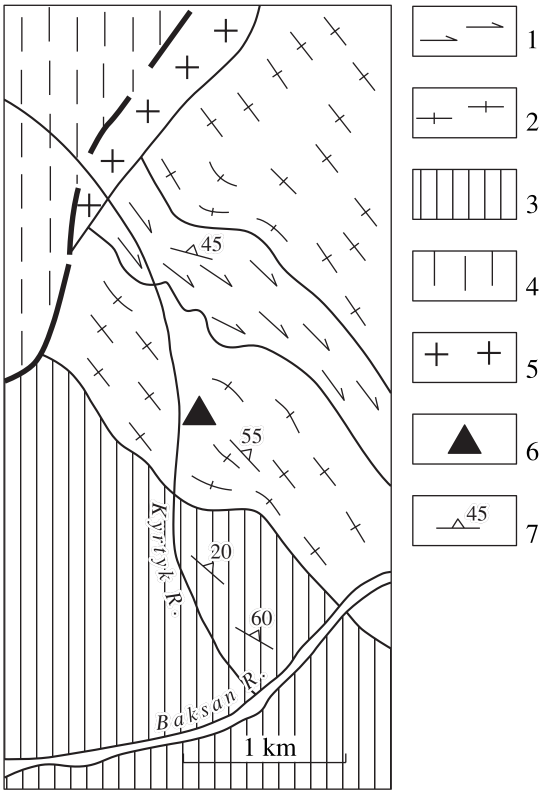

Fig. 2. Geological scheme of the Kyrtyk River lower reaches: (1) amphibolites; (2) orthogneisses; (3) undivided gneiss-migmatite complex; (4) Makera Complex; (5) granites; (6) sampling site 02-24; (7) attitude of metamorphic foliation.

obtained age value of 400 ± 10 Ma. On the other hand, the U–Pb age of about 1235 Ma, the unusual one for the Greater Caucasus, was recently determined by Somin et al. (2004) for orthogneiss (metaplagiogranite) from the Sofiya River left bank, which is of a lesser metamorphic grade as compared to gneisses of the Adyl-su and Kyrtyk canyons. This result gave rise to doubts, because homogeneity of material sampled by colleagues from the North Caucasus was unclear. Zircons from a new sample of the same metaplagiogranite, which was taken by Somin and analyzed by SHRIMP II U–Pb method at the Isotopic Research Center of All-Russia Research Institute of Geology, yielded value of 333 ± 5 Ma. It is likely that the former date at the level of Middle Riphean was obtained for clastic zircons from garnet-sillimanite paragneisses. In the sampled outcrop, these paragneisses are in contact with gneissoid metaplagiogranites, and by first sampling, both rocks were probably mixed in the field so that paragneisses represented prevailing part of analyzed material.

Considered below are geochronological results obtained for from Kyrtyk River canyon orthogneisses, Bol’shoi Mukulan Gorge orthogneisses, and Ulluchiran Glacier area migmatites.

Fig. 3. Exposure of orthogneisses, left wall of the Kyrtyk Canyon.

Tabular body of orthogneisses exposed in lower reaches of the Kyrtyk River canyon (left tributary of the Baksan River) is a few hundred meters thick (Figs. 2, 3). In the southwest, these rocks are in contact with migmatites, paragneisses and granites; in the northeast with uniform coarse-grained amphibolites (gabbro-amphibolites). Schistosity of orthogneisses is oriented parallel to sharp contacts with surrounding rocks and to schistosity in the latter. There are also visible signs of migmatization in orthogneisses. It is remarkable as well that angular amphibolite xenoliths derived from the main body and biotitized along periphery have been encountered in marginal zones of the orthogneiss body.

Like orthogneisses of the Adyl-su site, granoblastic massive orthogneisses from the Kyrtyk canyon reveal coarse to lenticular banding, being noticeably enriched in microcline (up to 50%). The other minerals of the rocks are plagioclase (not more than 15%), quartz (about 25%), and brown biotite (10%).

Orthogneisses of the other type, essentially different as compared to orthogneisses of the Kyrtyk and Adyl-su localities, were discovered on the Baksan River left side in the right wall of the Bol’shoi Mukulan Gorge (Somin et al., 2004) at the altitude from 2700 to 3100 m (Figs. 4, 5, 7). These are classical, mostly coarse-grained apogranite blastomylonites of the low-T amphibolite facies. The rocks have augen porphyroclastic structure and typical S/C structure produced by orientated scales of biotite, muscovite, and granulated quartz. These minerals envelope porphyroclasts of microcline and plagioclase, which are crushed and transformed into pelitic material in narrow marginal zones but retain magmatic zoning in central areas. The rocks are lacking signs of ultrametamorphism in distinction from orthogneisses of the Adyl-su and Kyrtyk localities. Because of intense mylonitization in places (e.g., in outcrops of the gorge floor), orthogneisses in question have coarse-banded pseudolayered structure, and their fine-grained bands are close in texture to ultramylonites.

Fig. 4. View at right wall and headwaters of the Bol'shoi Mukulan Gorge from the Tyrnyauz quarry: light band in the slope upper part corresponds to blastomylonitic orthogneisses; dark area below to schists of the Makera Complex; gneiss-migmatite complex is exposed higher, near the ridge top.

Fig. 5. Blastomylonitic augen orthogneiss; large light grains are feldspar porphyroclasts (Bol'shoi Mukulan site).

The main body of Mukulan orthogneisses is 250 to 300 m thick. In the northeast, it is in contact with chloritized but not migmatized quartz-two-mica schists of the Makera Complex, which contain garnet and andalusite. In the southwest, orthogneisses are overlain by rocks of the gneiss-migmatite complex, predominantly by biotite and biotite-sillimanite gneisses. Contacts of rock complexes are sharp, presumably tectonized and deformed in synmetamorphic conditions. One xenolith, a large amphibolite block, was encountered in orthogneisses. Thin lenticular bodies of leucocratic fine-grained gneisses, presumably metamorphic equivalents of granite porphyries, are confined to contact of orthogneisses with schists.

Migmatites, the other object of our study, are represented in infrastructure of the El'brus subzone by broad spectrum of varieties, from thin arterites to ghost migmatites of coarse-grained texture, the lenticular restite in which is composed of mica, garnet, and sillimanite. Leucosome of migmatites consists of quartz, plagioclase, and microcline; there are also plagiomigmatites. Like in other localities, the extent of migmatization depends on substrate: metapelites and paragneisses are migmatized to a greater extent than orthogneisses, and metabasites are almost deprived of migmatization. In many places, migmatites enclose variably oriented blocks of para- and orthogneisses, which are separated by coarse-banded migmatized material locally penetrating into blocks along schistosity planes. Migmatites experiences a multiphase folding. Near the Ulluchiran Glacier in the Cherek Bezengiiskii riverhead, where Sample 0-91L was taken for geochronological analysis (Fig. 6), migmatites are deformed into isoclinal folds of two generations that proves their origin under influence of regional (not contact) metamorphism.

Data considered above imply that orthogneisses from the Kyrtyk River canyon and Bol'shoi Mukulan Gorge belong to different structural and age groups. Protoliths of orthogneisses from the Kyrtyk site were emplaced before ultrametamorphism that affected rocks of gneiss-migmatite complex, whereas orthogneisses from the Bol'shoi Mukulan Gorge likely originated after or concurrently with metamorphic events. In other words, geochronological investigation of orthogneisses and migmatites of different types and ages is aimed at dating the protolith material and the last high-

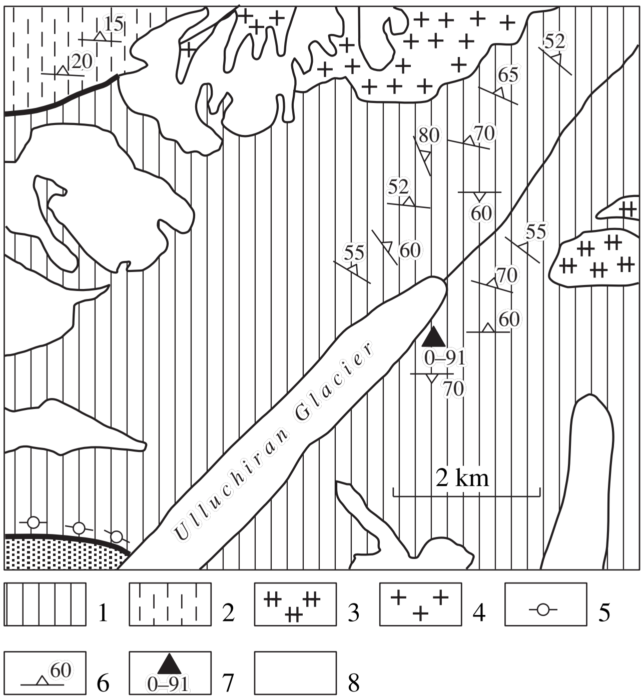

Fig. 6. Geological scheme of the Ulluchiran Glacier area: (1) gneiss-migmatite and (2) Makera complexes; (3) granites of the Belorechensk type; (4) granites of the Ullukam type; (5) mylonites; (6) attitude of metamorphic foliation; (7) sampling site 0-91L; (8) glaciers.

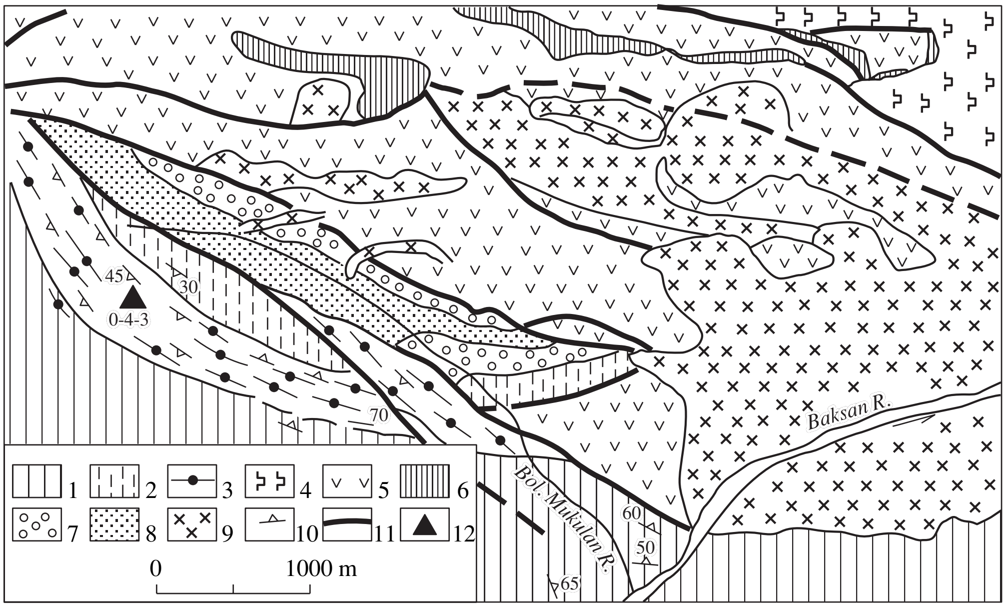

Fig. 7. Geological scheme of the Bol'shoi Mukulan area: (1) gneiss-migmatite and (2) Makera complexes; (3) blastomylonitic orthogneisses; (4) Bechasyn metamorphic complex; (5) Middle Paleozoic volcanogenic-sedimentary deposits of the fore ridge; (6) ultramafic rocks; (7) Upper Paleozoic molasses; (8) Jurassic deposits; (9) El'dzhurta granites, Neogene; (10) attitude of metamorphic foliation; (11) faults; (12) sampling site 0-4-3.

T metamorphic event that affected infrastructure of metamorphic basement in the El'brus subzone of the Greater Caucasus Main Range zone.

ANALYTICAL METHODS

Standard separation method in heavy liquids has been applied to extract accessory zircons from rock samples. The U–Pb dating is performed for zircon fractions of about 0.4 mg in weight and for lesser amount of zircon grains (9–16) separated from orthogneisses of the Kyrtyk (Sample 02-24) and Bol'shoi Mukulan (Sample 04-3) sites. Optical and thermoluminescence microscopy was applied to study inner structure of individual zircon grains. Surficial contamination of zircon crystals selected for geochronological analysis has been incrementally cleaned up in alcohol, acetone and 1M HNO_3. After each step, crystals have been washed in superpure water. The modified Krogh's method (Krogh, 1973) has been used to decompose crystals and to extract Pb and U. Total blank was not greater than 30 pg for Pb and 5 pg for U. Isotopic composition is determined in static mode on multicollector mass spectrometer Finnigan MAT 261 equipped with electronic multiplier (discrimination coefficient of multiplier for Pb is 0.32 ± 0.11% per amu).

Zircon fraction separated from leucosome of migmatite sample from the Ulluchiran site was not more than 0.24 mg in weight. Like in the first case, the modified Krogh's method (Krogh, 1973) has been used to decompose crystals and to extract Pb and U. Blank was not greater than 0.1 ng for Pb and 0.01 ng for U. Isotopic composition is determined on multicollector mass spectrometer Finnigan MAT 261. Fractionation coefficients are estimated at 0.10% for Pb and 0.08% for U per amu. Variation coefficient 0.90% (2σ) is determined for measured $^{206}Pb/^{238}U$ and $^{207}Pb/^{235}U$ isotopic ratios based on results of parallel analyses of zircon standard THA-16, age 475.6 Ma.

Experimental data are processed using programs PbDAT (Ludwig, 1991) and ISOPLOT (Ludwig, 1999). Standard U decay constants (Steiger and Jager, 1976) are used to calculate age values. Correction for common lead is consistent with that accepted in model by Stacey and Kramers (1975). All uncertainties are quoted at 2σ level.

RESULTS OF U–Pb GEOCHRONOLOGICAL STUDY

Orthogneiss from the Kyrtyk site. Zircon crystals from Sample 02-24 are of two morphological types. The first one is represented by subidiomorphic semitransparent pink crystals of prismatic and short-prismatic habit. In inner structure of these crystals, there are distinguishable fissured metamict cores or their relicts surrounded by transparent zoned rims (Figs. 8c, 8d, 8e). Pale-pink zircon crystals of the second type are idiomorphic, prismatic or long-prismatic in habit, having zoned inner structure (Figs. 8a, 8b). Crystals range in size from 30 to 200 μm, elongation coefficient 1.5 to 3.5. For the first analytical run, we selected most transparent zircon crystals from size fraction -85 + 53 μm (no. 1 in the Table). Then, from size fractions >100 μm

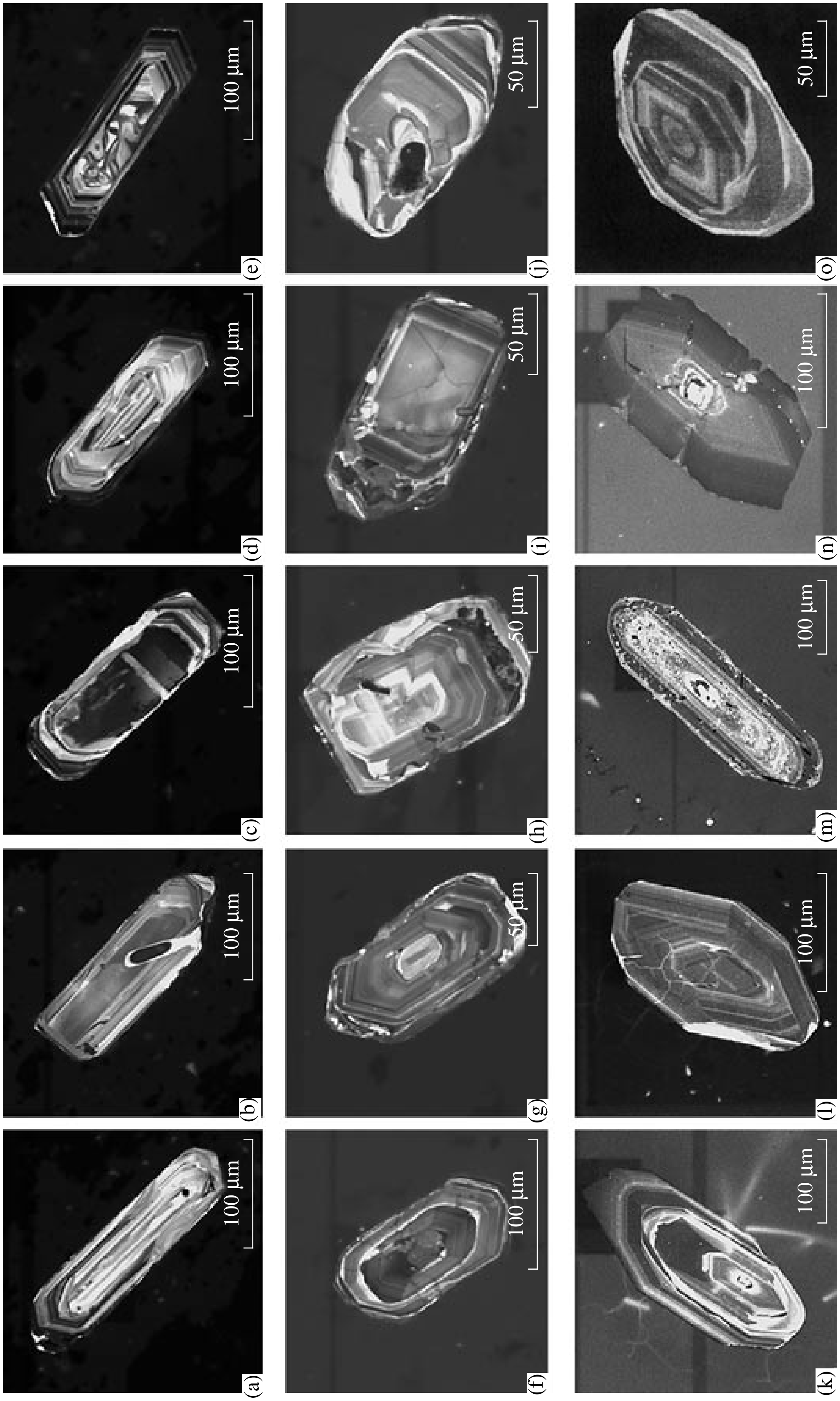

Fig. 8. Cathodoluminescence images of zircon shot under scanning electron microscope CamScan (accelerating voltage 15 kW): (a–e) zircons from orthogneisses of the Kyrtyk site; (f–j) zircons from orthogneisses of the Bol'shoi Mukulan site; (k–o) zircons from leucosome of migmatite from the Ulluchiran Glacier site.

and -85 + 53 μm, we analyzed aliquots of 16, 9 and 13 long-prismatic crystals of the second type (nos. 2, 3 and 4 in the table), and aliquot of 9 crystals was subjected to preliminary air abrasion.

As one can see in Fig. 9a, data points characterizing the above crystal aliquots plot on discordia with lower intercept corresponding in age to 386 ± 5 Ma and upper one to 1234 ± 92 Ma, MSWD = 1.1. It is remarkable that data points are concentrated near the lower intercept. The point most remote from that intercept corresponds to aliquot of zircon crystals subjected to air abrasion, and these crystals contain therefore a considerable proportion of inherited radiogenic lead. Deviation of isotopic composition characterizing zircons from size fraction -85 + 53 μm presumably means a partial loss of radiogenic lead in recent time. Morphological characteristics of zircons from orthogneisses imply their magmatic origin. Consequently, the age value of 386 ± 5 Ma estimated for the discordia lower intercept can be regarded as characterizing crystallization time of the orthogneiss protolith.

Orthogneiss from the Bol'shoi Mukulan site (Fig. 5). Blastomylonitic apogranite orthogneisses from this site (Sample 0-4-3) predominantly contain transparent to semitransparent idiomorphic zircon crystals of light pink color and prismatic to short-prismatic habit. The crystals occur as inclusions in feldspar and mica. They reveal fine internal zoning and presence of fissured translucent rims. By cathodoluminescence, rims lacking zoning are divided into segments of variable, sometimes unusually low luminosity (Figs. 8f to 8j). In some crystals, aggregates of dusty mineral inclusions are confined to relict cores (Fig. 8g). Zircon grains range in size from 30 to 150 μm, elongation coefficient 2.0 to 3.5.

Three aliquots of most transparent and idiomorphic zircon crystals have been selected for isotopic analysis from size fractions -85 + 53 and >85 μm. One of the aliquots (no. 7 in the table) was subjected to preliminary air abrasion treatment (Krogh, 1982). In addition, we analyzed 10 long-prismatic zircon crystals (no. 8 in the table). Data points of untreated aliquots plot on discordia (Fig. 9b) with upper intercept at 305 ± 8 Ma, while its lower intercept practically corresponds to zero, MSWD = 0.4. The $^{207}Pb/^{206}Pb$ age estimated for residue after air abrasion is older, equal to 424 Ma that probably indicates presence of inherited radiogenic lead. According to morphological characteristics and position of zircon crystals in the rock, they are of magmatic origin. Thus, there are grounds to suggest that the date 305 ± 8 Ma characterizes crystallization time of the rock protolith from parental magma.

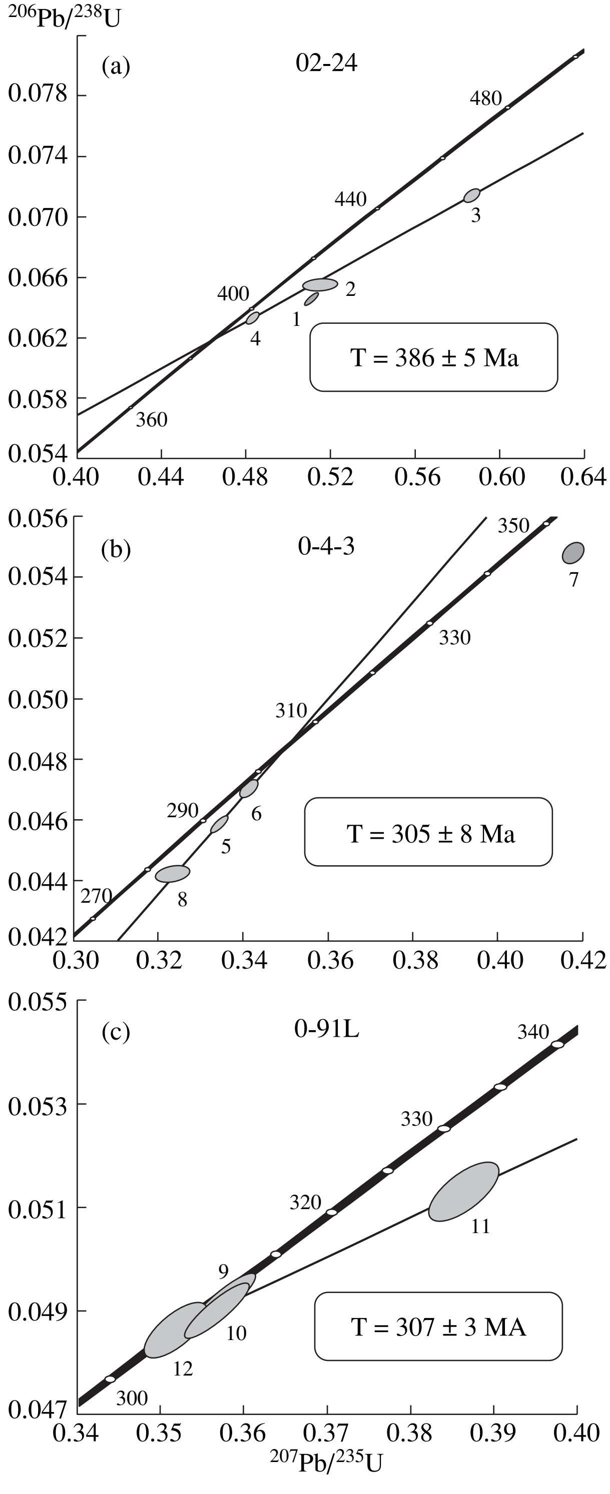

Fig. 9. Diagrams with concordia for studied zircons: (a) orthogneiss from the Kyrtyk site; (b) orthogneiss from the Bol'shoi Mukulan site; (c) leucosome of migmatite from the Ulluchiran Glacier site.

Leucosome of migmatite from the Ulluchiran site. Zircon separated from leucosome of migmatite (Sample 0-91L) is represented by ideally transparent and opaque long- and short-prismatic crystals of pale brown color. Acicular microinclusions frequently observable in central areas of opaque short-prismatic crystals are relatively rare in long-prismatic grains. By cathodoluminescence, crystals of both types reveal distinct magmatic zoning (Figs. 8k, 8l, 8o). In some grains, central resorbed areas apparently represent frag-

Results of U-Pb study of zircons from orthogneisses and migmatites of the Greater Caucasus Main Range zone

No.	Size fraction, µm, and its characterization	Weight, mg	Concentration, ppm		Isotopic ratios					Rho	Age, Ma		
			Pb	U	^{206}Pb/^{204}Pb	^{207}Pb/^{206}Pb^{a}	^{208}Pb/^{206}Pb^{a}	^{207}Pb/^{235}U	^{206}Pb/^{238}U		^{207}Pb/^{235}U	^{206}Pb/^{238}U	^{207}Pb/^{206}Pb
Orthogneiss, Kyrtyk River (Sample 02-24)													
1	–85 + 53	0.35	52.1	830	3687	0.05738 ± 4	0.0593 ± 1	0.5114 ± 10	0.0646 ± 1	0.89	419.4 ± 0.8	403.7 ± 0.8	506.3 ± 1.4
2	>100, 16 grains	–	U/Pb* = 15.0		285	0.05701 ± 76	0.0741 ± 1	0.5154 ± 70	0.0656 ± 3	0.2	422.0 ± 5.7	409.4 ± 1.6	491.8 ± 29
3	–85 + 53, A 20%, 9 grains	–	U/Pb* = 14.0		786	0.05956 ± 26	0.0728 ± 1	0.5871 ± 31	0.0715 ± 2	0.57	469.0 ± 2.4	445.2 ± 1.4	587.7 ± 9.3
4	–85 + 53, 13 grains	–	U/Pb* = 16.3		1430	0.05530 ± 22	0.0628 ± 2	0.4833 ± 25	0.0634 ± 2	0.67	400.3 ± 2.1	396.2 ± 1.4	424.3 ± 8.7
Orthogneiss, Bol'shoi Mukulan Gorge (Sample 0-4-3)													
5	–85 + 53	0.47	26.5	575	1295	0.05288 ± 5	0.0644 ± 1	0.3345 ± 7	0.0459 ± 1	0.82	293.0 ± 0.6	289.2 ± 0.6	323.5 ± 2.0
6	>85	0.38	10.4	218	942	0.05262 ± 11	0.0687 ± 1	0.3415 ± 9	0.0471 ± 1	0.56	298.3 ± 0.8	296.5 ± 0.6	312.5 ± 4.8
7	>85, –85 + 53, A 20%	0.18	11.4	212	1101	0.05530 ± 24	0.0800 ± 1	0.4180 ± 21	0.0548 ± 1	0.38	354.6 ± 1.8	344.0 ± 0.7	424.3 ± 9.9
8	<100, 10 grains	–	U/Pb* = 22.0		417	0.05285 ± 38	0.0666 ± 1	0.3225 ± 25	0.0443 ± 2	0.41	283.8 ± 2.2	279.2 ± 0.9	322.6 ± 16
Leucosome of migmatite, Ulluchiran Glacier (Sample 0-91L)													
9	>130	0.24	80.8	1803	11460	0.05270 ± 3	0.0016 ± 1	0.3575 ± 5	0.0492 ± 1	0.89	310.2 ± 2.4	309.4 ± 2.7	315.9 ± 1.3
10	100–130	–	U/Pb* = 22.0		3048	0.05280 ± 3	0.0055 ± 1	0.3566 ± 5	0.0490 ± 1	0.89	309.7 ± 2.4	308.3 ± 2.8	320.1 ± 1.3
11	85–100	–	U/Pb* = 21.0		5590	0.05464 ± 7	0.0098 ± 1	0.3865 ± 6	0.0513 ± 1	0.69	331.8 ± 2.6	322.5 ± 2.8	397.7 ± 2.8
12	85–100	0.19	64.6	1438	3154	0.05249 ± 6	0.0076 ± 1	0.3517 ± 6	0.0486 ± 1	0.71	306.0 ± 2.4	305.9 ± 2.7	306.7 ± 2.7

Notes: (^{a}) isotopic ratio corrected for blank and common lead; (A 20%) percentage of substance removed by aero-abrasive treatment; (*) concentration not determined; uncertainty corresponds to last significant figures.

ments of old cores (Figs. 8m, 8n). Morphological and crystallographic features suggest magmatic origin of separated crystals.

Four zircon fractions of different size (table) have been analyzed by U–Pb dating method. The analyzed zircons are relatively enriched in U and have low Th/U ratios (0.002–0.029) that is typical of zircons from leucosomes of migmatites.

As one can see from the Table, three zircon fractions (nos. 9 to 11) revealed a straight age discordance: $t(^{206}Pb/^{238}U) < t(^{207}Pb/^{235}U) < t(^{207}Pb/^{206}Pb)$, that is an evidence of ancient inherited Pb present in their isotopic systems. Fraction no. 12 (table) yielded the concordant age value of 306 ± 2 Ma (MSWD = 0.006, concordance probability 0.94). In the concordia diagram (Fig. 9c), data points characterizing the studied fractions fit to discordia with lower intercept at 307 ± 3 Ma and the upper one at 1340 ± 310 Ma (MSWD = 0.38). With due account for morphology of crystals having distinct internal zoning, we can regard the date 307 ± 3 Ma as age of leucosome crystallization in migmatites.

DISCUSSION AND CONCLUSIONS

The dating results suggest that high-T metamorphism of gneiss-migmatite complex, or to say correctly, of rocks from this complex, took place in the Greater Caucasus Main Range zone at the Late Paleozoic time. It is less clear, however, how old were protoliths of rocks under consideration. Nevertheless, we may state that in the complex, there are orthogneisses dated at 386 ± 5 Ma (Middle Devonian), and to a first approximation this date determines the upper age limits of host rocks: they should be older than 386 Ma. Consequently, the regional metamorphism should be younger than this date. As is mentioned above, orthogneisses from the Adyl-su site, which are comparable in composition and geological position with orthogneisses of the Kyrtyk site, are dated by U–Pb zircon method at 400 ± 10 Ma (Bibikova et al., 1991) that is close to the new date obtained. The other fact known from publications is the U–Pb date of about 422 Ma (Bakuradze et al., 1990) determined for gneisses (probably orthogneisses) in the Upper Svanetia, which also belong to gneiss-migmatite complex. Unfortunately, the last date was published without necessary analytical comments and geologo-petrological data. Besides, Hanel et al. (1993) reported that protolith of orthogneisses from the Adyl-su site was crystallized in the Late Paleozoic (500 ± 40 Ma ago), and attributed their host rocks to the Paleoproterozoic (~2000 Ma). The dates presented in their work are estimated, however, based on measured $^{207}Pb/^{206}Pb$ ratios, which usually point to presence of ancient inherited components in Phanerozoic rocks but do not determine the real age of the latter.

The aforementioned date of 1235 Ma has been determined at the discordia upper intercept for zircons from rocks of the Sofiya River site; the lower intercept corresponds therewith to zero age. This date is close to age of 1234 ± 92 Ma corresponding to upper intercept in diagram plotted for Sample 02-25 of orthogneiss from the Kyrtyk site, and it is reasonable to consider this age as reflecting presence of old inherited components in analyzed zircons. Consequently, our assumption that the Middle Riphean date obtained for rocks of the Sofiya River site is obtained for mixed clastic zircons containing the same components appears to be correct.

The Middle–Late Carboniferous age established for migmatites is unexpected to some extent. In the Greater Caucasus, this or even earlier (according to some researchers) age is characteristic of undeformed granites, which crosscut migmatites in the Main Range zone or incorporate their xenoliths (Baranov and Chernov, 1965). However, the intrusive contacts alone do not define the time span between events of migmatization and granite formation. Its is also admissible to assume a quick transition, in terms of geological time, from migmatization events to formation of large granite bodies. As has been noted long ago (Somin, 1965), tabular bodies of Late Paleozoic granites and surrounding frame structure are almost conformal. Being set in infrastructure, these bodies do not exert influence on contact rocks, which had, therefore, a high temperature at the intrusion time of granitoids. Similarly remarkable is the parallel orientation of mica flakes in marginal zones of both the granitoid bodies set in infrastructure and neighboring leucosome of migmatites. In addition, there are examples of mutually intertongued leucosome and small granite bodies containing skialiths of country rocks with diffuse boundaries. All these facts deny a considerable time gap between migmatization and granite formation suggesting genetic links between the two processes.

The date of 305 ± 8 Ma obtained for protolith of medium-T apogranite blastomylonite from the Bol'shoi Mukulan Gorge is an additional evidence in favor of metamorphic events in the Main Range zone at the time of the Middle–Late Carboniferous transition. On the other hand, transformation of Late Paleozoic granites into blastomylonites was a specific phenomenon characteristic of granites in the narrow marginal belt of the Main Range zone that is adjacent to the Fore Range zone, where mylonites proper were discovered long ago, but their formation age remained unknown.

New geochronological results allow a new interpretation of K–Ar dates known for infrastructure rocks, which mostly correspond to the Late Paleozoic interval (*Petrology of Metamorphic…*, 1991). The K–Ar dates for biotite–muscovite pairs from migmatites of gneiss-migmatite complex (potassium concentration above 7%) showed peaks of occurrence frequency near 280 and 310 Ma (Bibikova et al., 1991). These age values used to be considered as a result of thermal impacts exerted by Late Paleozoic granites, but now it is more reasonable to suggest that they correspond to cooling time of rocks after regional metamorphism responsible for origin of migmatites and granites in the Main Range zone.

There are some additional points to be discussed. For instance, it is known from geological works on the Greater Caucasus that granitoid clasts derived from the Main Range zone are widespread in the Upper Carboniferous molasses. If not a quick tectonic denudation, it is impossible to understand how migmatites and granites formed at the depth of 5 km at least reached the level of erosion so soon after their formation. Consequently, infrastructure of the Main Range zone may represent a metamorphic core risen up from below suprastructure, when the latter was detached along gently inclined tectonic planes, as it is suggested for metamorphic cores of the Cordilleran type.

Thus, it is possible to conclude that within the Greater Caucasus Main Range zone, the pre-Alpine basement of the El'brus subzone experienced the high-T regional metamorphism of the Variscan time. U–Pb ages of migmatites and metamorphosed granites, which originated at that time, correspond to the Bashkirian and Moscovian stages of the Carboniferous. Metamorphism was subsynchronous with granite formation in the Main Range zone, and genetic link between two processes can be postulated: granites could represent final products of sialic material melting in infrastructure by metamorphism. Substratum that experienced regional metamorphism included the Devonian rocks. With due regard for recent data reported by Somin et al. (2004), it is possible to assume that protoliths of metamorphic rocks from suprastructure of the El'brus subzone mostly corresponded in age to the Ordovician. If this is correct, then there is a certain age inversion in the subzone basement: older but less metamorphosed rocks are at the higher structural level than younger rocks. This paradoxical situation can be explained by allochthonous position of the Makera Complex (Gamkrelidze et al., 1996) that was thrust over younger rocks at the very end of the Late Paleozoic, as we think. The alternative explanation is that metamorphism provoked migmatization and granite formation in infrastructure situated deeper, and the youngest processes can be dated by U–Pb method, the most precise at present, at this level exactly. Rocks of suprastructure, where migmatites are unknown, are lacking metamorphic zircons related in origin with anatexis, thus being inappropriate for precise dating of metamorphic events. According to recent measurements (IGEM RAS), K–Ar age of large high-K muscovite crystals from synmetamorphic andalusite-bearing vein of the Makera Complex (Mt. Cheget locality) is 300 Ma, suggesting synchronism of regional metamorphic events in supra- and infrastructures. It is natural to expect, however, a later closure of isotopic systems in rocks of the infrastructure situated deeper.

As is established by recent research of crystalline cores in the mid-European Variscan belt, manifestations of metamorphic processes in infrastructure are younger (or seemingly younger) than in suprastructure and even in the cover (*DECORP*..., 1999). Thus, we probably meet a widespread phenomenon deserving a further comprehensive study.

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