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## GEOLOGY

# Vendian Metamorphism in the Accretionary–Collisional Structure of Central Asia

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The formation of the mosaic structure of Central Asia is related to the Early Caledonian tectogenesis. The fragments of continental massifs of Rodinia and its paleoceanic framework were juxtaposed 570–510 Ma ago with the Late Riphean and Vendian–Early Paleozoic oceanic basins and island arcs. Regional metamorphism and amalgamation of these structures by granitoids in the Early–Middle Ordovician resulted in the formation of the early Caledonian superterrane [1]. In the Vendian–Early Cambrian, shelf (platform) carbonate sediments accumulated in the Tuva–Mongolian and Dzabhan continental massifs of the superterrane; i.e., the continental massifs represent carbonate platforms. This was a reason to deny accretionary–collisional processes and regional metamorphism during this age in the southern framework of the Siberian Craton [2]. At the same time, the low-gradient metamorphism that predated the Early Paleozoic high-gradient metamorphism (~500 Ma ago) and terminated at the Vendian–Cambrian boundary (536 ± 5 Ma) has been established in the Moron Complex of the Tuva–Mongolian Massif, and the Vendian accretionary–collisional stage of its evolution was outlined on this basis [3]. However, it remained unknown how widely developed this stage was in the Early Caledonian superterrane of Central Asia.

In this communication, we present new geological and geochronological evidence (U–Pb dating of single zircon grains and microcharges) for identification of the Vendian stage in the formation of metamorphic complexes of the Baydrag block of the Dzabhan microcontinent. This, in turn, allows us to recognize the Vendian accretionary–collisional stage in the evolution of the Early Caledonian domain of Central Asia as a whole.

The metamorphic complexes of the Baydrag block make up the basement of the Dzabhan microcontinent and are regarded as elements of the South Hangay group of terranes in the present-day structure of Mongolia. This group also comprises the Bayanhongor ophiolitic zone, which marks a large suture between the Baydrag and Hangay blocks [4]. The pre-Riphean rocks metamorphosed in the Archean and Paleoproterozoic approximately 2650, 2365, and 1850 Ma ago [5] occur in the northwestern Baydrag block in the Bumbuger district (Fig. 1). The high-grade metamorphic rocks related to the Paleoproterozoic tectogenesis (1.9–1.7 Ga) are localized at the Tuyin-Gol and Tatsaiin-Gol interfluve to the east and southeast of Bayanhongor. The basement rocks are nonconformably overlain by mainly greenschist sequences [7] combined into the Olziit-Gol Complex (Burd-Gol Formation [6]), the age of which is presumably Middle Riphean or the lowermost Upper Riphean [4]. The age of greenschist metamorphism of the black shales of the Olziit-Gol Complex has been estimated at ~840 Ma (K–Ar age of actinolite). The younger metamorphism of the amphibolite facies is superimposed on greenschists along the contact with ophiolites of the Bayanhongor zone. The kyanite–staurolite–garnet–biotite–plagioclase–quartz and garnet–kyanite–muscovite–biotite–plagioclase–quartz schists are developed in this area. The hornblende–garnet assemblage appears in calcic rocks. The formation of these rocks was previously assigned to the Early Caledonian tectogenesis, when the obduction accompanied by zonal metamorphism and emplacement of granitic plutons took place in the Late Cambrian–Early Ordovician. However, the rocks with kyanite-bearing mineral assemblages are found along the entire zone of the ophiolitic melange irrespective of contact with granites, including its southeastern segment in the Tsatsaiin-Gol basin (Fig. 1). We suppose that these rocks represent a fragment of the metamorphic belt belonging to the kyanite–sillimanite facies series. The development of the younger andalusite-bearing mineral assemblage in the Olziit-Gol Complex is related to contact effects of postkinematic granites (Fig. 1). On the basis

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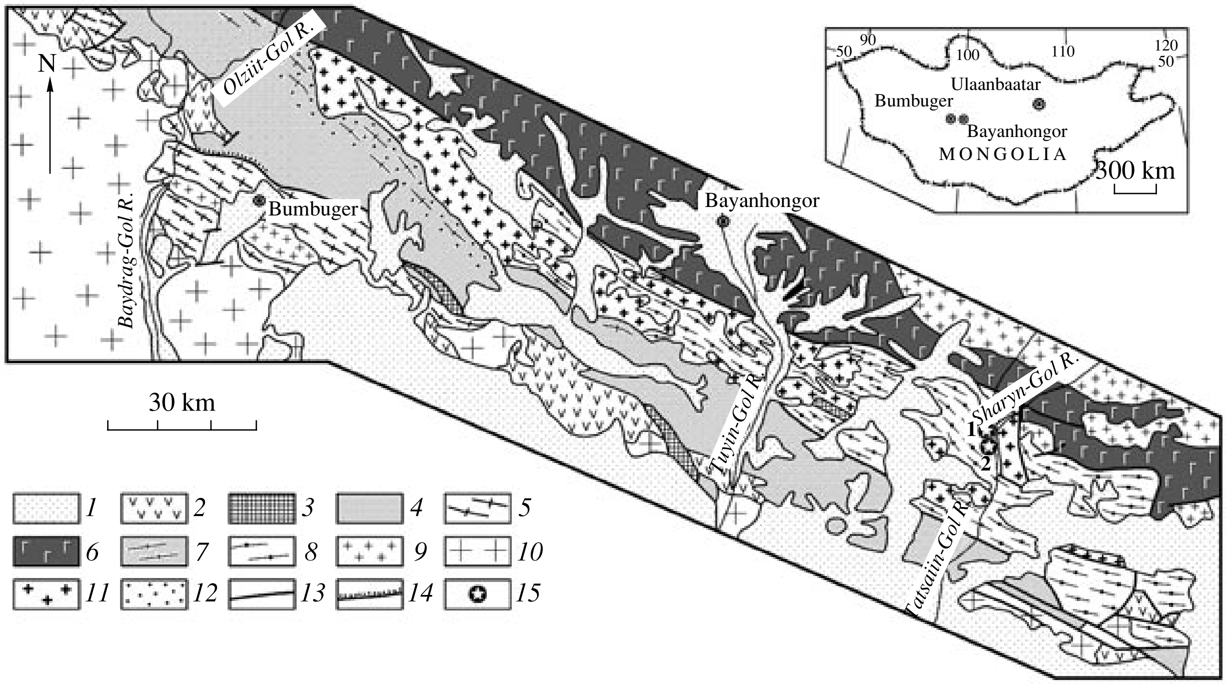


**Fig. 1.** Location of metamorphic complexes at the junction of the Baydrag block and ophiolites of the Bayanhongor zone (modified after [4–6]). (*1*) Mesozoic and Cenozoic rocks; (*2*) Upper Paleozoic rocks; (*3*) Lower and Middle Paleozoic rocks; (*4, 5*) Baydrag block: (*4*) Middle and Upper Riphean Olziit-Gol cover complex, (*5*) Neoarchean Baydrag and Paleoproterozoic Bumbuger basement complexes, unspecified; (*6*) Bayanhongor zone of ophiolitic melange; (*7, 8*) rocks of the Vendian low-gradient metamorphic belt superimposed on (*7*) rocks of the Olziit-Gol Complex and (*8*) crystalline rocks of unknown age; (*9*) Late Paleozoic and Early Mesozoic granitoids, unspecified; (*10*) Paleozoic granitoids, unspecified; (*11*) Vendian post-late collisional diorite–granodiorite–granite complex; (*12*) zone of high-gradient contact metamorphism; (*13*) fault zones; (*14*) proved stratigraphic unconformity at the base of the Olziit-Gol Complex; (*15*) sample location: (**1**) Sample 5975, (**2**) Sample 5967.

of K–Ar, Ar–Ar and Pb–Pb estimates, it was suggested that collision, metamorphism, and igneous activity were developed contemporaneously 540–450 Ma ago during the closure of the vast Vendian oceanic basin, the fragments of which have been retained in the Bayanhongor and Lake zones [8]. It should be noted that the Vendian age of ophiolites in the Lake zone (568 ± 4 and 573 ± 6 Ma) was determined with the U–Pb zircon method [9], whereas the age of plagiogranites from the ophiolitic complex of the Bayanhongor zone was estimated at 569 ± 21 Ma with the Sm–Nd isochron method for rock-forming minerals [10]. Therefore, the age correlation is rather unreliable.

The hornblende quartz diorite (sample 5975) and the pegmatoid kyanite-bearing granite (sample 5967) that crosscut the metamorphic rocks in the middle reaches of the Tatsaiin-Gol have been chosen as objects for geochronological study. The sample location is shown in Fig. 1.

The quartz diorite belongs to the older intrusive phase of the diorite–granodiorite–granite complex, which postdated low-gradient metamorphism and related folding. However, low-angle blastomylonite zones with newly formed fine-flaky biotite and chlorite are traced in diorite of the older phase. The kyanite-bearing two-mica pegmatoid granites are typically synkinematic rocks. They occur as thin (15–30 cm) and extended (>30 m) veins conformable with the main foliation of amphibolite facies and affected by folding. Boudinage is noted occasionally. The results of U–Pb isotopic study of zircons from the aforementioned granitoids are given in the table and in Figs. 2 and 3.

Zircons from the quartz diorite (sample 5975) are represented by crystals of two types. Zircon 1 comprises colorless or pale yellow euhedral transparent grains (primarily, their fragments) of long-prismatic and prismatic habit (Fig. 2). The grains have a high birefringence and zonal structure (Fig. 2). Zircon 2 crystals are subhedral, translucent, and fractured. They are characterized by prismatic, rarely long-prismatic habit and yellow color. They also reveal magmatic zoning, which completely disappears at crystal margins that have a lower birefringence (Fig. 2). The size of zircon grains varies from 50 to 250 μm ($K_{elong}$ = 2–4). Three charges of the most transparent zircon crystals of zircon 1 were used for the U–Pb study. They were picked out from size fractions of –85+60, –100+85, and >100 μm, and preliminarily subjected to air abrasion. As can be seen from the table and Fig. 3, the data points of isotopic compositions of the studied zircons virtually lie on concordia (the degree of discordance is 1.5–2.2%). Their average $^{207}Pb/^{206}Pb$ age is 547 ± 4 Ma, Results of U–Pb isotopic study of zircons

| No. | Size fraction, μm and its characteristics | Charge, mg | Content, μg/g | | Isotope ratios | | |
|---|---|---|---|---|---|---|---|
| | | | Pb | U | $^{206}Pb/^{204}Pb$ | $^{207}Pb/^{206}Pb$(a) | $^{208}Pb/^{206}Pb$(a) |
| Sample 5975; coordinates: 45°46′32″ N, 101°23′19″ E | | | | | | | |
| 1 | –100 + 85, A 40% | 0.30 | 61.3 | 640 | 4110 | 0.0584 ± 1 | 0.2220 ± 1 |
| 2 | >100, A 15% | 0.25 | 54.1 | 577 | 5903 | 0.0584 ± 1 | 0.1929 ± 1 |
| 3 | –85 + 60, A 10% | 0.06 | 38.4 | 396 | 667 | 0.0587 ± 3 | 0.1916 ± 1 |
| Sample 5967; coordinates: 45°45′14″ N, 101°23′28″ E | | | | | | | |
| 4 | >100, CLC, 2 grain | – | U/Pb* | = 11.9 | 424 | 0.0584 ± 5 | 0.1122 ± 1 |
| 5 | >150, 1 grain, prismatic | – | U/Pb* | = 11.2 | 2074 | 0.0588 ± 2 | 0.1217 ± 1 |
| 6 | –100 + 85, 25 grains, apices, acid treatment | – | U/Pb* | = 10.1 | 342 | 0.0590 ± 3 | 0.1179 ± 1 |
| 7 | >100, 40 grains, acid treatment | – | U/Pb* | = 10.9 | 2468 | 0.0612 ± 1 | 0.1138 ± 1 |

| No. | Size fraction, μm and its characteristics | Isotope ratios | | *Rho* | Age, Ma | | |
|---|---|---|---|---|---|---|---|
| | | $^{207}Pb/^{235}U$ | $^{206}Pb/^{238}U$ | | $^{207}Pb/^{235}U$ | $^{206}Pb/^{238}U$ | $^{207}Pb/^{206}Pb$ |
| Sample 5975; coordinates: 45°46′32″ N, 101°23′19″ E | | | | | | | |
| 1 | –100 + 85, A 40% | 0.6958 ± 14 | 0.0864 ± 2 | 0.88 | 536 ± 1 | 534 ± 1 | 546 ± 2 |
| 2 | >100, A 15% | 0.7015 ± 14 | 0.0871 ± 2 | 0.83 | 540 ± 1 | 538 ± 1 | 546 ± 2 |
| 3 | –85 + 60, A 10% | 0.7084 ± 39 | 0.0876 ± 3 | 0.54 | 544 ± 3 | 541 ± 2 | 554 ± 10 |
| Sample 5967; coordinates: 45°45′14″ N, 101°23′28″ E | | | | | | | |
| 4 | >100, CLC, 2 grain | 0.6548 ± 83 | 0.0813 ± 7 | 0.67 | 511 ± 6 | 504 ± 4 | 545 ± 20 |
| 5 | >150, 1 grain, prismatic | 0.7073 ± 35 | 0.0873 ± 4 | 0.83 | 543 ± 3 | 540 ± 2 | 558 ± 6 |
| 6 | –100 + 85, 25 grains, apices, acid treatment | 0.7406 ± 45 | 0.0910 ± 4 | 0.62 | 563 ± 3 | 562 ± 2 | 568 ± 10 |
| 7 | >100, 40 grains, acid treatment | 0.7584 ± 24 | 0.0898 ± 3 | 0.83 | 573 ± 2 | 554 ± 2 | 648 ± 4 |

Note: U–Pb study has been carried out for both large charges (~0.3 mg) and single grains of zircon. Control of the internal structure of individual zircon grains was performed with an optic microscope and cathode luminescence control (CLC). In the latter case, approximately half of a zircon grain directly extracted from the specimen was used for the CLC. The selected zircon grain underwent multistep removal of surficial contaminants with alcohol, acetone, and 1M $HNO_3$. After each step, the zircon grain or its fragment was washed with especially pure water. The decomposition and chemical separation of Pb and U was conducted by a modified Krogh technique [11]. The total blank over the period of study was not higher than 30 pg for Pb. Pb and U isotopic compositions were determined on a Finnigan MAT 261 mass spectrometer in a static regime or with an electronic multiplier (coefficient of discrimination for Pb is 0.32 ± 0.11 amu). The experimental data were processed with PbDAT [12] and ISOPLOT [13] programs. The commonly adopted constants of U decay were used in age calculations. Corrections for common Pb were introduced in compliance with model values. All uncertainties are given at 2σ level. (a) Isotope ratios corrected for procedure blank and common lead; (A10%) the amount of material removed by air abrasion of zircon; (*)zircon charge weight was not determined; (CLC) grain chosen with the CLC method. Uncertainties correspond to the last significant decimal digits.

which may be accepted as the timing of quartz diorite formation.

Zircons from the kyanite-bearing granite (sample 5967) are represented by light pink and light brown, subhedral and less frequent euhedral, translucent, and transparent crystals, and their short-prismatic and prismatic fragments (Fig. 2). The overwhelming majority of zircon crystals have partly metamict irregular cores and transparent zonal outer shells with a high birefringence (Fig. 2). The size of zircon grains varies from 50 to 200 μm ($K_{elong}$ = 2–3). The U–Pb study has been carried out for single zircon grains, including those chosen with cathode-luminescence control (CLC) (table, nos. 4, 5; Fig. 2), and for microcharges of zircons preliminarily subjected to acid treatment (table, nos. 6, 7). The data point of isotopic composition of single zircon grains and 25 apices of the crystals that underwent acid treatment make up a regression line. Its upper intercept with concordia corresponds to 570 ± 17 Ma (Fig. 2). The lower intercept corresponds to 224 ± 190 Ma (MSWD = 0.047) (Fig. 2). At the same time, the data points of 25 zircon grains are located on concordia, with a concordant age of 562 ± 2.4 Ma (MSWD = 1.02; probability = 0.31). This is consistent with the age corresponding to the upper intercept. A slightly older $^{207}Pb/^{206}Pb$ age was obtained for 40 zircon grains from the >100 μm fraction also affected by acid treatment (table, no. 7), indicating the presence of an older inher-

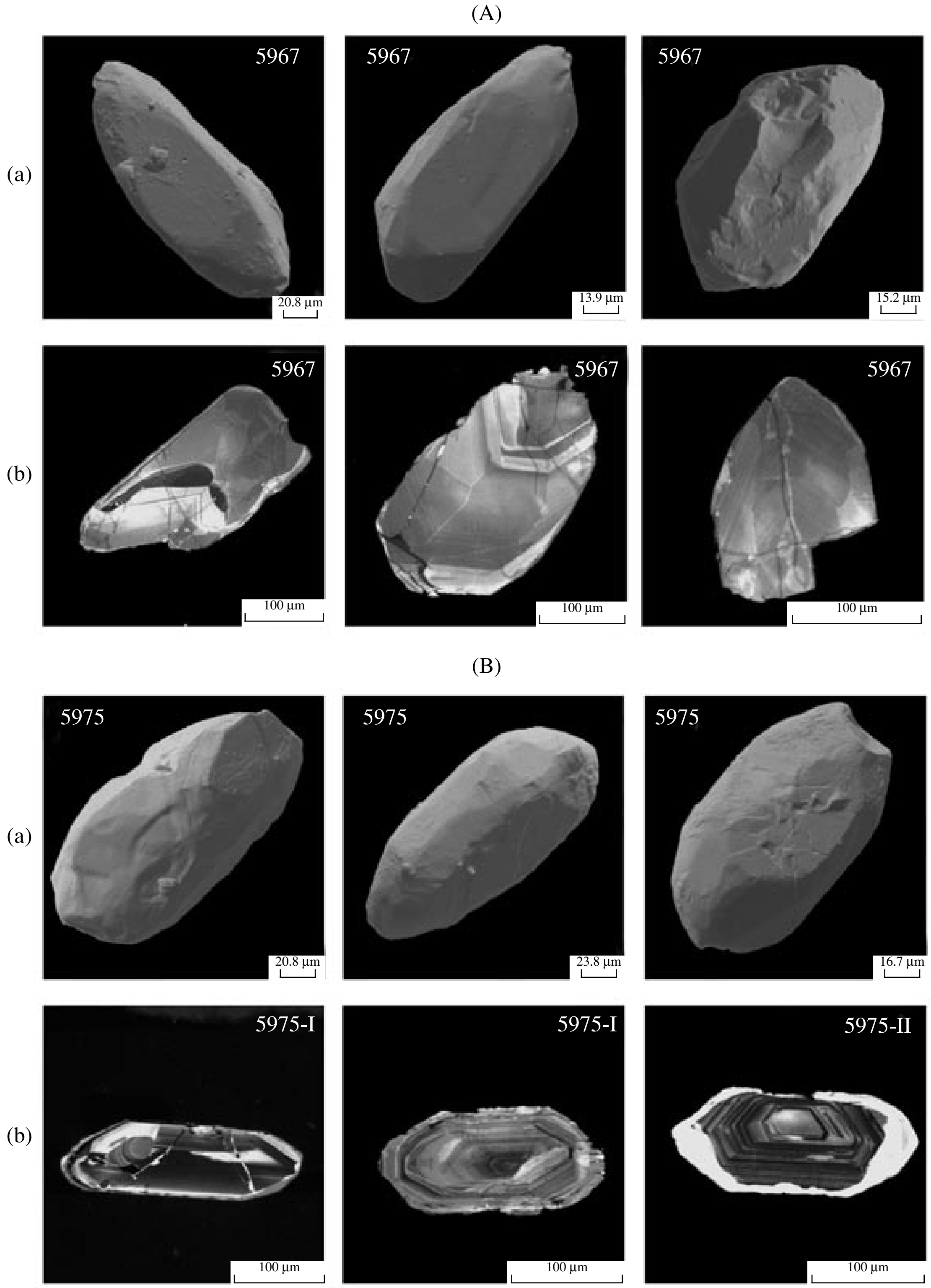


**Fig. 2.** Photomicrographs of zircon from (A) Sample 5967 and (B) Sample 5975, taken with (a) ABT-55 SEM (accelerating voltage 20 kV) and (b) cathode-luminescence detector on CamScan SEM (accelerating voltage 15 kV).

ited component of radiogenic Pb. The concordant zircon age of 562 ± 2.4 Ma is accepted as the most reliable estimate that characterizes the age of kyanite-bearing granite crystallization.

The data obtained make it possible to outline the Vendian (562 ± 2.4 to 547 ± 4 Ma) metamorphic belt of kyanite–sillimanite facies within the Baydrag block. The metamorphic belt was produced by the collision of

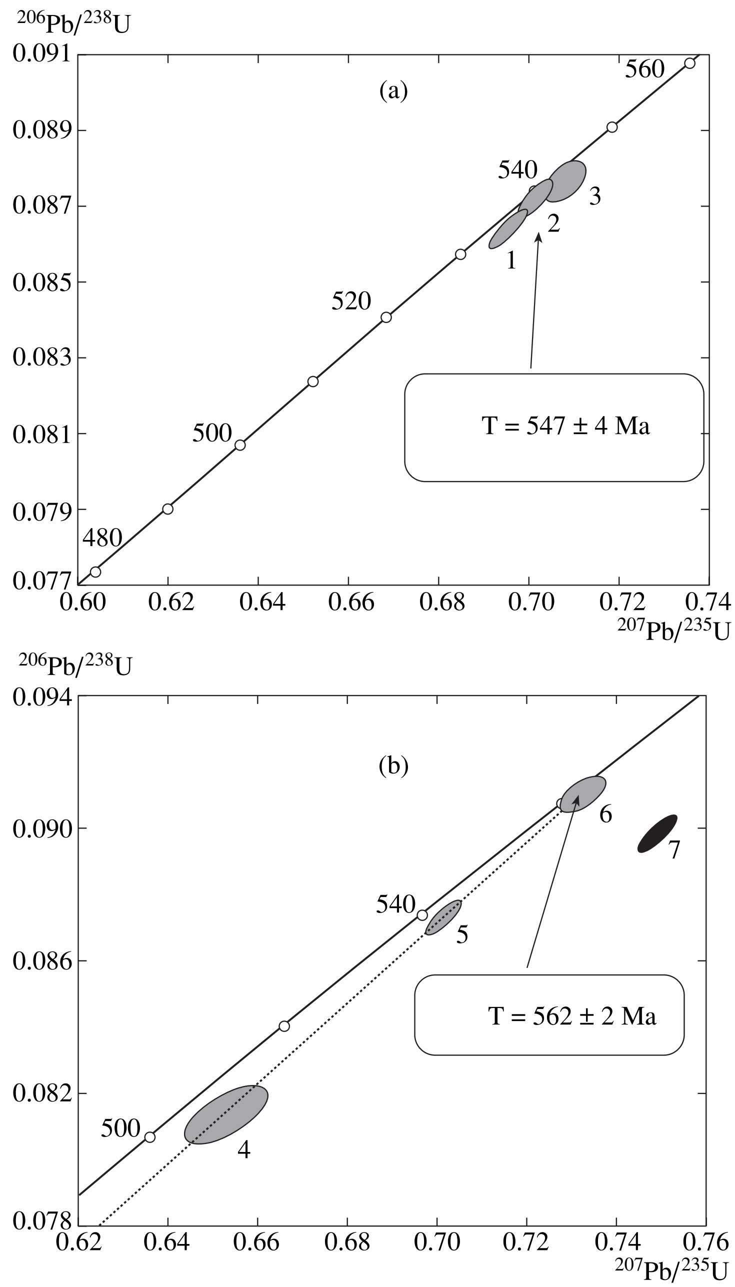


**Fig. 3.** Data points of (a) Sample 5975 and (b) Sample 5967 plotted on a diagram with concordia. Data point numbers correspond to those in the table.

continental massifs during the closure of the Bayanhongor paleoceanic basin (fragments of the massifs are presented in the Baydrag and Hangay blocks). The obtained estimates testify to the virtually synchronous (560–570 Ma) origination of short-lived basins of the Lake zone and development of the low-gradient metamorphic belt along the Bayanhongor suture. In [8], the simultaneous opening of the basins pertaining to the Lake and Bayanhongor zones and the formation of the metamorphic belt between them was suggested on the basis of the Sm–Nd age of ophiolites in the Bayanhongor zone estimated at 569 ± 21 Ma [10]. However, the above value reflects the time of closure of the Sm–Nd system of the respective minerals at the final stage of metamorphism, rather than a true age of the rocks belonging to the ophiolitic association. This statement has been confirmed recently by the data on the Late Riphean age (665 ± 15 Ma) of zircon contained in anorthosite from the layered complex of the Bayanhongor ophiolites [14]. Thus, we can suggest that the formation of the Vendian oceanic crust in the Lake zone and the subsequent evolution of island arcs occurred synchronously with the closure of the Late Riphean Bayanhongor basin. In this process, the subduction zone plunged beneath the northern (in present-day coordinates) part of the continental massif represented by the Baydrag block in the present-day structure. The manifestation of high-pressure metamorphism indicates that structures with the thick Earth's crust were formed in the Vendian during the evolution of island arcs or a single arc.

The results obtained provide evidence for recognition of the autonomous Baikalian accretionary–collisional stage in the evolution of the Bayanhongor zone marked by the formation of a single zone amalgamated by granitoids in the Late Vendian. The newly formed Late Baikalian composite massif was repeatedly involved in collision during the closure of basins pertaining to the Lake zone. This is confirmed by younger deformations and supported by K–Ar and Ar–Ar isotope data [8].

As has been mentioned above, the Vendian low-gradient metamorphism was established in the Moron Complex of the Tuva–Mongolian Massif [3], the South Chu inlier of Caledonides in Gorny Altai (555 ± 6 Ma) [15], and the Kan block of the eastern Sayan region (555 ± 5 Ma) [16]. In general, the geological and geochronological information obtained to date gives grounds to recognize the Vendian accretionary–collisional stage in the evolution of Paleoasian ocean. This stage predated the origin of the Early Caledonian superterrane in the mosaic structure of Central Asia in the Late Cambrian–Early Ordovician [4].

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