

Late Cretaceous–Early Cenozoic Volcanism of Southern Mongolia: A Trace of the South Khangai Mantle Hot Spot

V. V. Yarmolyuk^a, E. A. Kudryashova^a, A. M. Kozlovsky^a, and V. M. Savatenkov^b

^a *Institute of the Geology of Ore Deposits, Petrography, Mineralogy, and Geochemistry (IGEM),
Russian Academy of Sciences, Staromonetnyi per. 35, Moscow, 119017 Russia*

^b *Institute of Precambrian Geology and Geochronology, Russian Academy of Sciences,
nab. Makarova 2, St. Petersburg, 199034 Russia*

Received March 13, 2006

Abstract—The Gobi–Tien Shan volcanic area (in Southern Mongolia) is part of the South Khangai volcanic region (SKVR). The formation of its lava fields was related to three stages of volcanic activity: the Late Cretaceous (88–71 Myr), Paleocene–Early Eocene (62–47 Myr), and Early Oligocene (37–30 Myr). Volcanic occurrences of different age are represented by trachybasalt, trachyandesitobasalt, basanite, and melanephelinite with similar geochemical characteristics, which are also close to the geochemical characteristics of OIB basalt. The isotope composition (Sr, Nd) of the rocks indicates that the magma sources were formed as a result of mixing of a moderately depleted PREMA mantle and an EM-I mantle enriched in neodymium. The patterns of migration of volcanic centers of different ages over the area of interest have been studied. The earliest (Late Cretaceous) volcanic occurrences were concentrated mainly in the south of the area, the Paleocene–Early Eocene eruptions took place at the center of the area, and the Early Oligocene volcanism occurred in the northern area. The observed migration of the volcanic activity centers is related to lithospheric plate motions relative to a localized source of hot mantle (the South Khangai mantle hot spot), which controlled volcanic activity within SKVR. In the lithospheric structure of this region, local asthenospheric high, reaching a depth of ~50 km, correspond to this hot spot.

DOI: 10.1134/S0742046307010010

INTRODUCTION

The convergence of the northern Asian and Chinese continents, which was accompanied by a successive closure of the Paleoasian Ocean, took place at the end of the Carboniferous. The sea basin, identified as the Mongolia–Okhotsk trough, remained only in the eastern part of the paleo-ocean, but this basin also closed by the beginning of the Jurassic [3]. From that time, the largest part of Central Asia appeared deep within the continent. Therefore, the magmatic process, which proceeded in the Late Mesozoic and Cenozoic in Central Asia far from the boundaries of the lithospheric plates, produced a striking example of the effects of intraplate activity. This magmatic activity had many stages and, at the same time, was localized in a number of areas in Central Asia, which made it possible to distinguish in this region several volcanic provinces that developed for a long time, namely the Eastern Mongolian, Western Transbaikalian, and South Khangai provinces [18]. A prolonged (up to 150 Myr) development of these provinces was related to the formation of regions above the mantle hot spots, which affected a relatively stiff lithosphere. Evolutionary development of the regions, governed by periods of intensification of volcanic activity which alternated with intervals of calm, continued for up to 10 Myr without a pronounced change in the structural conditions of volcanic occurrence [15, 18].

During the Late Cretaceous and Paleogene, volcanic activity within the regions was sharply decreased, which became a general feature of these regions. Therefore, volcanic products of a limited distribution, corresponding to these stages of the volcanic evolution of the region, remained almost unstudied for a long time and the geochronological and isotope–geochemical data, and geological position of these products was unknown. As a result, the concepts that a relationship between the early (Early Cretaceous) and late (Late Cenozoic) stages in the regional formation was absent, and that the corresponding volcanic products belonged to independent volcanic areas of different age, predominated in the literature [9, 10]. As an example, we present the SKVR, whose history of formation includes the Early Cretaceous and Late Cenozoic epochs of increased activity which were characterized by relatively large volume volcanic eruptions (up to 1000 km³ and larger). The time interval between 90 and 30 Ma is considered as a period of quiescence. This made it possible to distinguish the Gobi–Altai and Late Cenozoic Khangai volcanic regions without historical relations [9, 10].

Based on new paleovolcanological, geochronological, and isotope–geochemical data, we here consider the specific features of volcanism in the Late Cretaceous–Early Cenozoic interval of SKVR magmatic his-

tory and present our determination of the time patterns of volcanic occurrence. Moreover, we show that the character of the volcanism, including structural patterns in volcanic occurrence and eruption types, remained almost unchanged during this period, and that the composition of volcanic products and, correspondingly, that of volcanic sources, insignificantly varied over time.

GEOLOGY

Fragments of slightly dislocated lava plateaus in wide areas between the Gobi Tien Shan ranges and central parts of the Khangai upland [15] belong to SKVR. The historical evolution of the region is traced from the late Jurassic. During the Early Cretaceous, volcanic activity was related to the formation of grabens in the Gobi Altai rift zone, within which volcanic sequences of up to 1000 m and more in thickness were accumulated [9, 13, 21]. Volcanism occurred here in a series of stages and gradually abated by 100–90 Ma ago, when the large group of laccolites, stocks, and small lava fields composed of basanite originated. Between 90 and 30 Ma ago, volcanic activity was mainly localized south of the Gobi Altai ranges in intermontane depressions of the Gobi Tien Shan [14, 19]. Beginning from 30 Ma, the zone of active volcanism gradually shifted northward in the region of the Lake Valley in the southern and central Khangai, where most of the centers of the Miocene–Pleistocene volcanism are concentrated [2, 9, 20]. During the entire historical evolution of SKVR, subalkaline basalts and alkaline basaltoids predominated in the composition of volcanic products. The formation of trachyrhyolite (in the middle of the Early Cretaceous), melanephelinite, phonolite, and trachyte in association with carbonatite (late Jurassic–early Cretaceous) became episodic [12, 13]. Only basic rocks of the subalkaline and alkaline series were formed from the second half of the Early Cretaceous to the Holocene inclusive.

The present work studied the volcanic fields formed 90–30 Ma ago [14, 19]. The area of their distribution (hereafter, the Gobi Tien Shan area) occupies a wide territory in the region of the eastern termination of the Gobi Tien Shan ranges (Fig. 1). Volcanic occurrences predominantly occurred in intermontane depressions filled with Late Cretaceous–Cenozoic sediments and are more rarely encountered along the slopes and the axial parts of ranges. The distribution of lava fields and associated dikes, shocks, and other subvolcanic bodies is controlled by sublatitudinal normal faults and grabens, which were formed simultaneously with the volcanism, judging by the absence of vulcanites on the elevated shoulders of these faults.

The first data on the age of the volcanic formations in this part of southern Mongolia and, specifically, showing that they belong to the Late Cretaceous and Early Cenozoic stages of volcanism, were obtained by Nikolaeva and Shuvalov [14]. Later, we performed systematic paleovolcanological, geochronological, and

isotope–geochemical studies of these fields [19]. The results of the K–Ar dating of igneous rocks, completed with new data, are given in Table 1 and are generalized in Fig. 2. Taking into account the available geochronological data and the patterns of distribution of volcanic fields with similar ages, we distinguish the following stages of volcanism in the history of the area: **Late Cretaceous** (between 71 and 88 Ma ago), **Paleocene–Early Eocene** (combining lava fields formed between 47 and 62 Ma ago), and **Early Oligocene** (30–37 Ma ago).

The Late Cretaceous stage. Volcanic fields of this age occur mainly in the southern area. They are localized along the sides of a large Bordzongiin–Gobi depression and within the Khurmen latitudinal graben, which cuts the Gurban–Saikhan range structures from the south (Fig. 1). Individual small stocks and rock extrusions are encountered east and north of these structures; therefore, the total distribution area of the rocks of this age is more than 30 thousand km².

The K–Ar ages of the rocks in this group of volcanic fields are mainly found in the interval of 71–75 (±5) Ma (Table 1). Trachybasalt stocks of the Durveldzhin–Ula mount at the southern slope of the Noen range and the rocks from the Khurmen volcanic field are slightly older (88 ± 4 and 84 ± 5 Ma, respectively). The latter rocks possibly correspond to the initial phase of the Late Cretaceous volcanism. Thus, during the considered stage, the rocks were formed from 88 to 71 Ma ago.

The magmatic products of this stage are represented by fragments of thin (several tens of meters) lava plateaus with local relics of strongly dislocated slag cones, as well as by volcanic domes, dikes, stocks, and laccolites. Igneous rocks and subvolcanic formations are, as a rule, separated in space, and subvolcanic bodies are usually controlled by faults and are observed within the Late Cretaceous sedimentary rocks underlying lava sheets. This makes it possible to consider the complex of the Late Cretaceous magmatic rocks as a set of manifestations of the sheet and root facies of volcanic eruptions. The lava plateaus are composed of a series of thin lava flows, which are characterized by a high porosity of rocks similar in composition to olivine and olivine–pyroxene trachybasalts. In contrast to these plateaus, subvolcanic bodies are composed of massive rocks: melanephelinites, basanites, and trachybasalts. Large (to 20 cm) xenoliths of lherzolites and pyroxenites are sometimes observed in the composition of these rocks.

Figure 3 demonstrates the schematic structure of the Khurmen lava field and its paleoneck. This volcanic neck is, first of all, distinguished due to debris of a slag–bomb cone, the area of which exceeds 5 km². Slightly stratified accumulations of bombs (up to 1.5 m across), lapilli, and slag of trachybasalt participate in the neck structure. The thickness of these explosive formations is more than 50 m judging by the depth of a quarry excavated here. These formations are cut by dikes of fluidal basanites (the thickness of which reaches 3 m, and sometimes exceeds this value) and by a stock

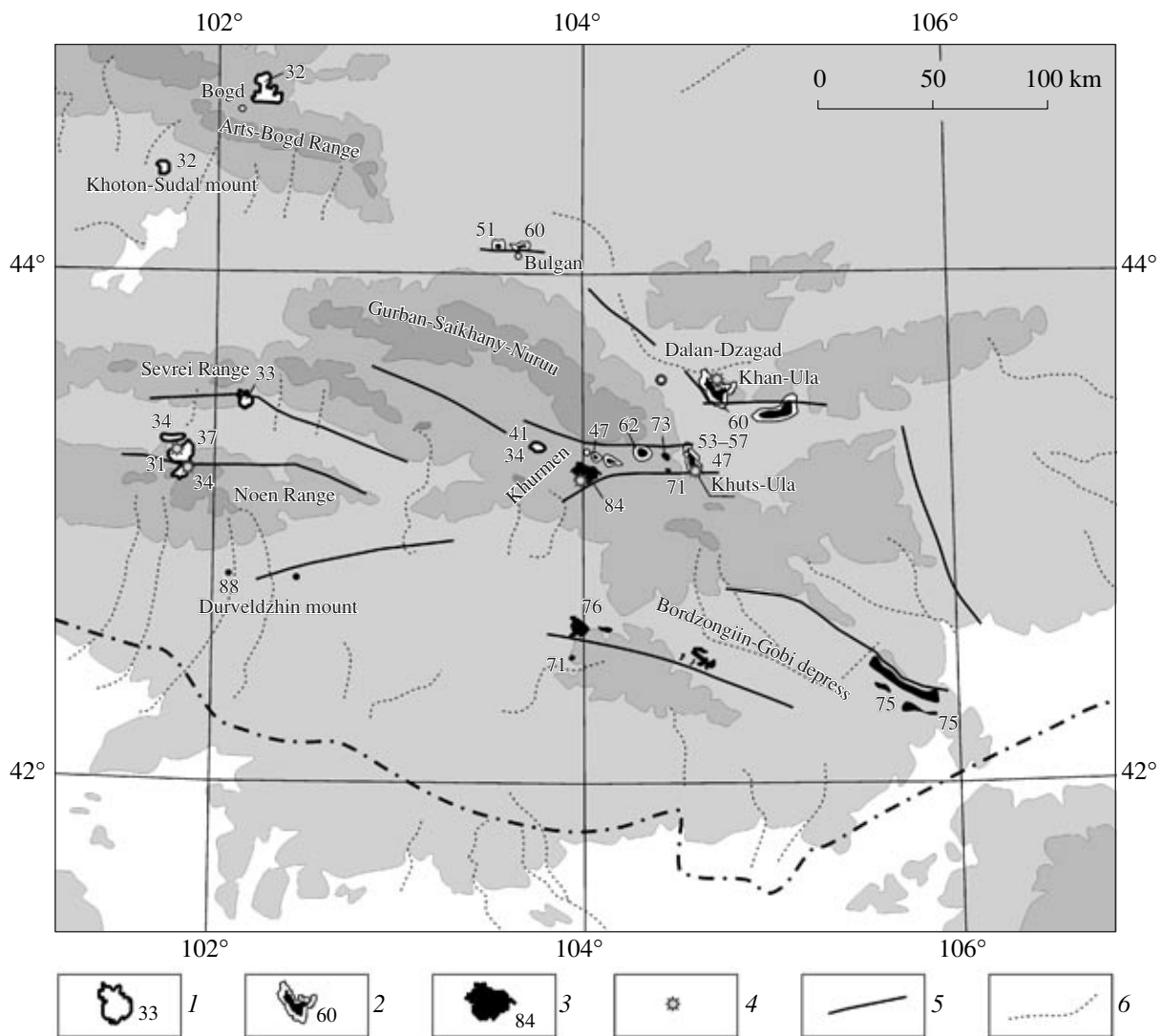


Fig. 1. Schematic location of the Late Cretaceous–Early Cenozoic lava fields within the Gobi Tien Shan volcanic area. (1)–(3) Lava fields of volcanic stages and their age in Ma: (1) Early Oligocene, (2) Paleocene–Early Eocene, (3) Late Cretaceous; (4) volcanic centers (slag cone debris); (5) faults; (6) ephemeral stream channels.

(50 × 70 m) of massive trachybasalts. The stock structure includes a linear zone of crushed rocks, which cuts the stock and is composed of semirounded fragments of trachybasalt (up to 10 cm across) cemented by the same lava. This zone possibly originated as a result of the protrusion of a cooled portion of trachybasalts, with the formation of breccia (fragments of which were rounded and cemented by its own lava) in the contact zone.

To elucidate the character of the volcanic activity within the lava field, it seems important to note that explosive breccias are overlain by thin (to 10 m) flows of trachybasalts, which are observed in the horizontal lava sheet on the neck flanks. Bombs, slag, and other explosive products are not encountered between lava flows. Such relationships were established not only in the region of the Khurmen lava field but also in other areas. These relationships indicate that volcanic activ-

ity at this stage began with a phase of explosive activity, with formation of slag cones, and then subsequently changed into a phase of calm lava outflows.

The total volume of volcanic products in the resulting lava fields is about 10 km³ at this stage.

The Paleocene–Early Eocene stage is represented by isolated dikes, lava domes, and the fragments of lava plateaus of small width (up to 50 m) composed of several (not more than seven) lava flows of olivine basalt, the outcrops of which are mainly concentrated around the Gurban-Saikhany range (Fig. 1). The K–Ar age of the lavas is 62–47 Ma (Table 1). The largest fragments of the lava fields of this age are located in the eastern part of the Khurmen graben. Shuvalov [14] established that, in the region of the Khuts-Ula and Tevsh-Ula summits, the lavas lie gently on the Early Eocene rocks but, additionally, they overlie the lava sequence dated as the

Table 1. Results of K–Ar dating of lava sequences in southern Mongolia

Volcanic stages	Sample no.	Rock	Sampling region	Sampling site		K, mass %	⁴⁰ Ar _{rad} , ppb	Age (± 2σ), Ma
				latitude, N	longitude, E			
Early Oligocene	1720/1	Trachyandesitobasalt	Noen Range	43°12.463'	101°50.645'	3.01	7.1	34 ± 3
	1720/2		Northern part of Noen Range	43°17.190'	101°44.991'	3.02	6.4	31 ± 3
	NT 4/1a			43°17.359'	101°47.888'	3.32	8.6	37 ± 3
	NT 4/2a			43°20.445'	101°46.908'	2.82	6.7	34 ± 3
	TV95-12.2*		Southern slope of Sevrei Range	43°30.65'	102°10.63'	–	–	33 ± 0.1
	TV95-12.7.2*			43°30.21'	102°09.92'	–	–	32.7 ± 0.2
	1727		Khurmen settlement	43°18.700'	103°45.720'	2.71	6.5	34 ± 3
	KhRM 2/1a		–	–	–	2.77	7.9	41 ± 4
	ATs 10/1		Arts-Bogd Range	44°42.926'	102°13.729'	–	–	32.5 ± 2
	TV95-2.10*		–	–	–	–	–	30.4 ± 0.1
	IKh 3/1		Khoton-Sudal	44°30.322'	102°20.509'	–	–	33.2 ± 1
	IKh 3/4			–	–	–	–	32 ± 2
Paleocene–Early Eocene	1743 Sh**	Trachybasalt	Khuts-Ula	43°15.936'	104°36.458'	1.10	3.5	47 ± 3
				–	–	–	–	53 ± 8
				–	–	–	–	54 ± 4
				–	–	–	–	57 ± 5
	1743/1		Khurmen settlement	43°17.165'	104°04.560'	1.85	6.1	47 ± 3
	1726		Bulgan settlement	43°17.489'	104°04.280'	1.68	7.3	62 ± 5
	1739			44°04.254'	103°40.536'	1.12	4.6	60 ± 4
	Sh**							51 ± 2
	YuKh-4/4	Trachyandesitobasalt	Khan-Ula	43°32.200'	104°44.387'	1.89	8.1	60.4 ± 1.5
Late Cretaceous	ULNZ3/1a	Trachybasalt	Khuts-Ula	43°13.726'	104°28.072'	1.49	7.4	71 ± 5
	ULNZ4/1a	Trachyandesitobasalt	Southern slope of Bordzongiin-Gobi depression	43°16.231'	104°27.738'	1.85	9.6	73 ± 5
	1723			42°29.997'	103°56.818'	2.18	10.8	71 ± 4
	GNKh1/3a			–	–	2.10	11.2	75 ± 5
	GNKh1/4a			42°35.578'	104°02.427'	1.98	10.6	76 ± 5
	GNKh1/5a			42°36.099'	104°01.238'	2.25	12.1	76 ± 5
	GNKh1/6a			–	–	2.27	12.0	75 ± 5
	1740/1	Melanephelinite	Northern slope of Bordzongiin-Gobi depression	42°16.989'	105°43.531'	–	–	75 ± 7**
	1741			42°15.780'	105°52.918'	1.11	5.8	75 ± 4
	KhRM1/1a	Trachyandesitobasalt	Khurmen settlement	43°12.120'	104°02.229'	1.55	9.2	84 ± 5
	1721/1	Basanite	Durveldzhin	42°48.178'	102°05.444'	1.40	8.7	88 ± 4

* Dating according to [20];

** dating according to [14].

end of the Early Cretaceous, and are separated from this sequence by a horizon of sedimentary rocks. The formation of the lava sequence was related to a volcanic center between the relics of a bomb–slag cone in the region of the Khuts-Khula mount. The relationships between the explosive products and the lava products

are the same here as within the Khurmen lava field considered above; i.e., volcanic activity also began here with explosions, which were subsequently replaced by calm outflows. Outside the Khurmen graben, individual volcanic occurrences were established east of Dalan-Dzagad in the regions of the Khan-Ula rock mass and

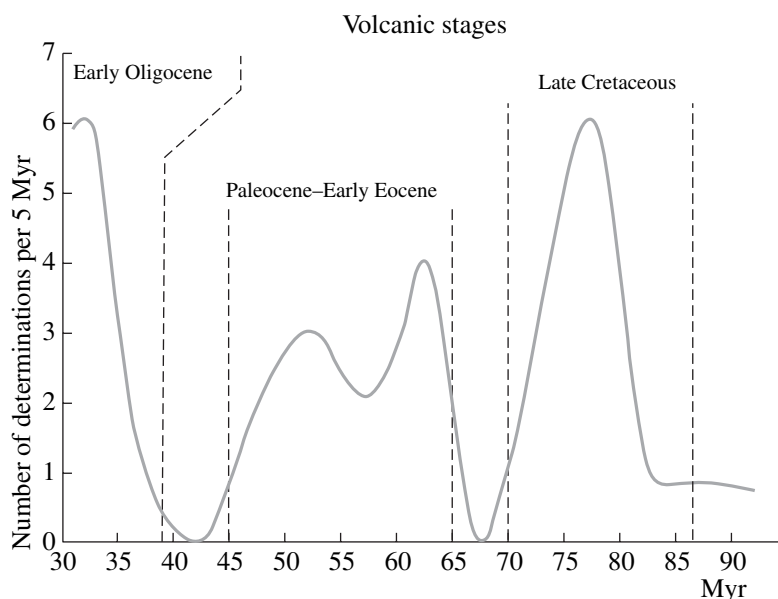


Fig. 2. Distribution of K–Ar dating of igneous rocks in the Gobi Tien Shan volcanic area (compiled according to data of Table 1).

Bulgan settlement (Fig. 1). The volume of volcanic products resulting from this stage is very small (not more than several cubic kilometers).

The Early Oligocene stage is represented by lava fields mostly concentrated in the intermontane depression between the Sevrei and Noen ranges as well as by isolated extrusions and lava fields scattered along the southern and northern periphery of the Arts-Bogd range (Fig. 1). These formations are mostly composed of trachyandesitobasalts with rare lherzolite inclusions. The rocks are dated as 37–31 Ma old. The vulcanites are structurally controlled by sublatitudinal normal faults. A volcanic graben with a unified volcanic cover apparently existed between the Noen and Sevrei ranges. The fragments of this graben are exposed at the edges of the intermontane depression as lava plateau relics, which armor depression slopes, but are also encountered in the axial part of the Noen range, thereby indicating that this range was formed later (after the Oligocene). The lava sheet thickness within this lava plateau reaches 50–70 m. Flows of flaggy or massive trachyandesitobasalts, separated by zones of lava with a large amount of slag, participate in the lava sheet structure. The volcanic field area was not less than 1000 km²; therefore, the volcanic occurrence volume possibly reaches 50 km³.

General regularities of evolution. The data presented here indicate that volcanic activity within the Gobi Tien Shan area repeatedly intensified during the Late Cretaceous and Early Cenozoic. Stages of volcanic activity and quiescence alternated and had approximately equal duration. Volcanic activity was relatively similar and manifested itself as outflows in the form of basaltic plateaus accompanied by the formation of a large group of subvolcanic intrusions: dikes, stocks, domes, laccolites, etc., which fixed effluent magma

channels. Eruptions apparently began with gas explosions, accompanied by ejections of bombs and slag, and subsequently changed into a phase of calm outflows of highly mobile lavas, which flowed over large distances and formed wide lava plateaus.

COMPOSITION OF VOLCANIC PRODUCTS

Methods of analytic studies. The percentages of major elements in rocks were determined at the Institute of Geochemistry, Siberian Division, Russian Academy of Sciences (Irkutsk) using X-ray fluorescent spectroscopy and a CRM-25 quantometer, according to the technique presented in [1].

Multi-elemental analysis of rock samples was performed at the Institute of Analytic Instrument Engineering, at the Russian Academy of Sciences (St. Petersburg) using a PlasmaQuad 3 VG elemental mass spectrometer with inductively coupled plasma. The shooting conditions were published in [7]. The drift of the relative device sensitivity in the series of specimens (not more than 5–10 specimens) was controlled by measuring standard solutions of heavy metals (Ti, Cr, Ni, Cu, and Pb) and a BCR-1 standard sample. To perform the REE analysis, the device was calibrated using a multi-elemental standard solution of REEs manufactured by Matthew Johnson. The relative error in determining elements was not higher than 5–10%.

An isotope analysis of Nd and Sr was performed at the Institute of Precambrian Geology and Geochronology (St. Petersburg). Preparation of specimens for isotope analyses, including chemical decomposition and extraction of elements using the method of ion-exchange chromatography, is described in detail in [11]. A Triton multiple-collector solid-phase mass

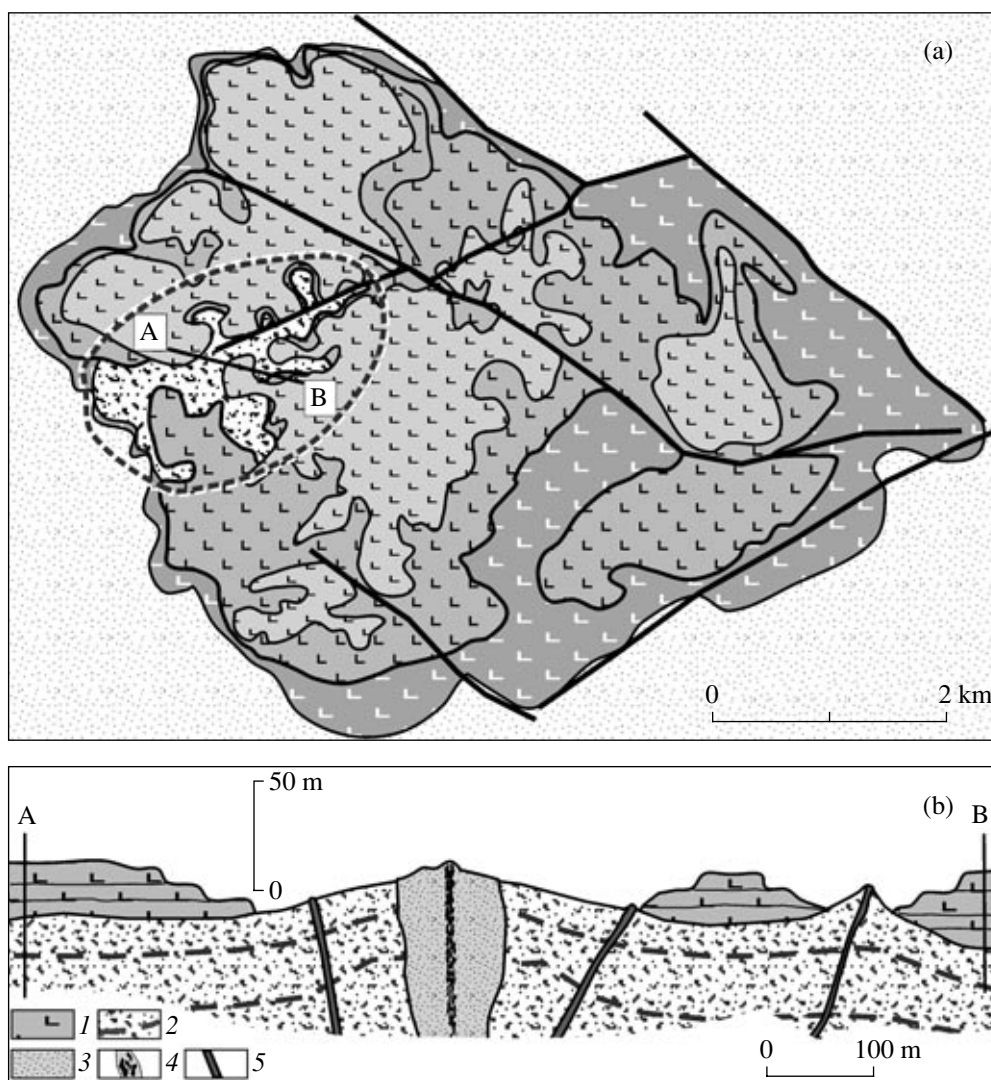


Fig. 3. Schematic structure of the lava field (a) at the southwestern termination of the Khurmen graben and (b) of the volcanic neck within the graben: (1) lava fields (in plan, the size and cast of symbols are changed); (2) bombs, tuff, and slag with stratification elements; (3) volcanic stock basanites; (4) rock crushing zone; (5) dikes.

spectrometer was used to determine the Nd and Sr isotope composition. The repeatability of determining Rb, Sr, Sm, and Nd concentrations, calculated based on multiple analyses of a BCR-1 standard, was $\pm 0.5\%$. The value of a blank experiment was 0.05, 0.2, 0.3, and 0.8 ng for Rb, Sr, Sm, and Nd, respectively. The results of the analysis of a standard BCR-1 sample (six measurements) are as follows: $[\text{Sr}] = 336.7$ ppm, $[\text{Rb}] = 47.46$ ppm, $[\text{Sm}] = 6.47$ ppm, $[\text{Nd}] = 28.13$ ppm, $^{87}\text{Rb}/^{86}\text{Sr} = 0.4062$, $^{87}\text{Sr}/^{86}\text{Sr} = 0.705036 \pm 22$, $^{147}\text{Sm}/^{144}\text{Nd} = 0.1380$, and $^{143}\text{Nd}/^{144}\text{Nd} = 0.512642 \pm 14$. The repeatability of isotope analyses was controlled by determining the composition of La Jolla and SRM-987 standards. During the period of Sr measurements, the obtained $^{87}\text{Sr}/^{86}\text{Sr}$ value in an SRM-987 standard was 0.710241 ± 15 (2σ , ten measurements), and the $^{143}\text{Nd}/^{144}\text{Nd}$ value in a La Jolla standard was 0.511847 ± 8

(2σ , 12 measurements). The isotope composition of Sr was normalized for $^{88}\text{Sr}/^{86}\text{Sr} = 8.37521$ and the Nd composition for $^{146}\text{Nd}/^{144}\text{Nd} = 0.7219$. The Nd isotope composition was reduced to the tabulated value of a La Jolla standard ($^{143}\text{Nd}/^{144}\text{Nd} = 0.511860$).

Petrography. Lavas of flows, domes, necks, and subvolcanic intrusions predominate among volcanic formations of southern Mongolia. Slag and bombs, which are encountered near centers of eruptive activity, are less widespread. Lavas are as a rule represented by dense dark gray to black olivine, more rarely, olivine-pyroxene or substantially pyroxene (which is typical of melanephelinites) rocks. The petrographic appearance of these rocks is rather uniform. These are mainly porphyric and, more rarely, aphyric rocks. Rocks from flow centers are usually represented by porphyric crystallized varieties including phenocrysts of olivine (to

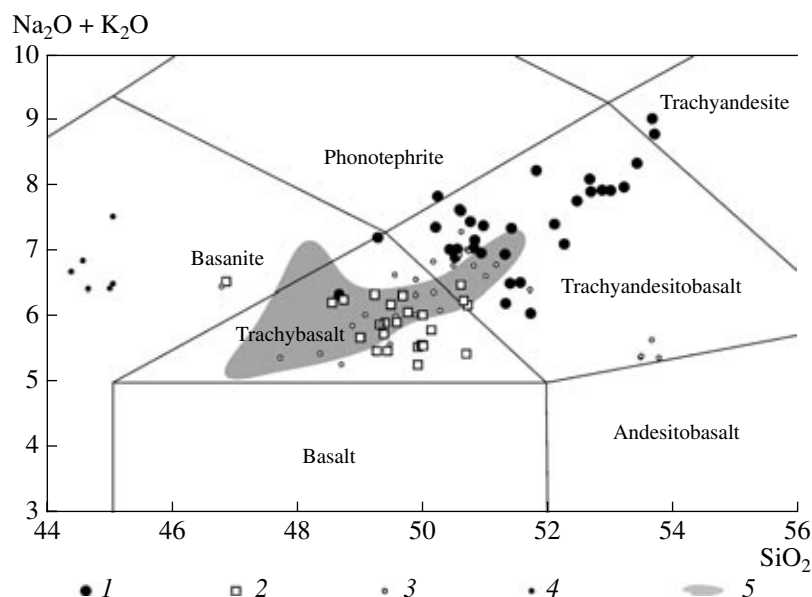


Fig. 4. The SiO_2 – $\text{Na}_2\text{O} + \text{K}_2\text{O}$ diagram for the rocks in the Gobi Tien Shan volcanic area. (1)–(4) Compositions of rocks of different-age volcanic stages: (1) Early Oligocene, (2) Paleocene–Early Eocene, (3)–(4) Late Cretaceous ((3) basaltoids, (4) melanephelinites); (5) field of compositions of the Late Oligocene–Miocene rocks from the Lake Valley Khangai area [19].

20%), clinopyroxene (to 10%), and, more rarely, plagioclase. Black vitreous varieties are typical of the lower (bottom) and upper (slag) flow zones.

Olivine phenocrysts are usually euhedral and reach 0.5–1.5 (and sometimes more than 2) mm across. Iddingsite develops differently over olivine. More rarely, iddingsite is replaced by a bowlingite aggregate, which sometimes develops over glass, forming specific tear-shaped yellowish and greenish separations. Pyroxene is represented by small prismatic brown or sometimes, greenish, augite crystals. The size of phenocrysts reaches 0.4–0.8 mm, much more rarely, 1.5 mm. Plagioclase is rarely encountered in the composition of phenocrysts and is observed in the form of blades of 0.5–1 mm in length. The main mass is usually crystallized (mainly fine, more rarely, medium-grained) and is composed of plagioclase microlites, small grains of olivine, clinopyroxene, and grains or elongated crystals of apatite and ore mineral of up to 1 mm across. The texture of the rocks mostly varies between microdoleritic and microdiabasic. Secondary changes in these rocks are atypical, and carbonate is the most ordinary secondary mineral.

Petrochemistry. According to the data on chemistry (Table 2), the rocks of the Gobi Tien Shan volcanic area correspond to melanephelinites, basanites, trachybasalts, and trachyandesitobasalts (Fig. 4). The composition varies within 44–54 mass % of SiO_2 and 5–9 mass % of $\text{Na}_2\text{O} + \text{K}_2\text{O}$. On the whole, these rocks are highly enriched in titanium ($\text{TiO}_2 > 1.8$ mass %) and are characterized by increased contents of P_2O_5 (from 0.3 to 1.1 mass %) and moderate contents of Al_2O_3 (<16 mass %) (Fig. 5). The rock composition

indicates that melts are rather highly differentiated, which is specifically confirmed by a relatively small Mg number value (<0.6; on average, 0.50–0.52).

The composition is most variable in the Late Cretaceous rocks, including melanephelinites, basanites, trachybasalts, and trachyandesitobasalts, the SiO_2 content of which varies from 44.3 to 53.3 mass %. In this group, differences in the compositions are regional. Lower silicate melanephelinites and basanites ($\text{SiO}_2 < 48$ mass %) gravitate to the southeaster margin of the Bordzongiin-Gobi depression, where they form chains of subvolcanic intrusions. On graphs, these rocks are characterized by an inverse linear dependence or a constant content of major elements with a change in the Mg number value (except MgO). This makes it possible to relate these rocks to olivine fractioning from an initial melanephelinite melt. Igneous rocks of the same age from the western part of the Bordzongiin-Gobi depression and the Khurmen graben region correspond to the composition of trachybasalts and trachyandesitobasalts. Judging by the absence of regularities in the distribution of figurative points on variation diagrams, the compositions of initial melts of rocks from this age group slightly differed at different sites of the volcanic area.

The Paleocene–Early Eocene rocks form a group of a rather stable composition. With only one exception, the percentage of SiO_2 varies within 46.8–50.7 mass %. On graphs, these rocks form a rather compact group of figurative points, which can probably be related to the composition of the initial melts.

The Early Oligocene rocks are apparently closest to the products of differentiation of initial magmas with

Table 2. Chemical, standard (mass %), and rare element (ppm) composition of volcanic rocks in the Gobi Tien Shan area

Volcanic stages	Early Oligocene							
Sample no.	NT-24/19	NT-24/20	NT-24/21	NT-24/22	NT-24/23	NT-24/24	YuKh-1/1	YuKh-1/2
N E	43°12.451' 101°50.657'			43°17.141' 101°44.923'	43°17.161' 101°44.948'		43°30.161' 102°10.464'	43°30.278' 102°10.321'
SiO ₂	52.57	52.80	52.04	52.16	51.77	51.25	49.94	49.16
TiO ₂	2.10	2.12	2.18	2.19	2.17	2.30	2.34	2.27
Al ₂ O ₃	14.88	14.78	14.66	14.71	14.57	14.86	14.20	14.22
Fe ₂ O ₃	9.29	9.33	9.90	9.93	10.08	9.63	10.44	10.06
MnO	0.10	0.10	0.11	0.11	0.11	0.11	0.12	0.12
MgO	4.01	3.69	4.45	4.02	4.43	4.82	5.10	5.35
CaO	5.55	5.91	5.86	5.58	6.23	5.60	7.84	7.83
Na ₂ O	4.68	4.82	4.13	4.39	3.99	3.89	4.27	4.59
K ₂ O	3.94	4.08	3.69	3.77	3.79	4.01	2.98	2.82
P ₂ O ₅	0.68	0.68	0.70	0.70	0.72	0.78	0.72	0.71
LOI	2.32	1.66	2.22	2.57	2.29	2.82	1.70	2.52
Total	100.13	99.97	99.94	100.13	100.15	100.07	99.65	99.64
CIPW*								
Albite	33.0	31.1	33.6	34.1	31.4	31.1	25.9	24.3
Diopside	12.8	15.4	11.8	11.7	13.1	9.7	19.7	20.6
Nepheline	4.1	5.6	1.2	2.1	1.7	1.5	6.0	8.5
Olivine	10.4	8.8	12.2	11.5	11.9	13.3	10.9	10.9
Ti	10977	10886	11799	12074	11880	12590	12118	12481
Cr	69	78	101	88	102	75	127	107
Co	27	28	29	29	29	29	33	31
Ni	65	69	77	77	80	73	72	64
Rb	45	64	55	53	48	67	40	55
Sr	944	915	1142	998	977	982	1277	1082
Y	16	16	16	16	16	17	18	17
Zr	201	199	176	180	176	210	200	202
Nb	46	46	42	43	42	52	52	54
Ba	656	661	683	759	723	647	761	628
La	24	24	24	24	24	25	28	29
Ce	58	56	55	59	63	57	60	63
Pr	7.83	7.85	7.82	8.00	8.05	8.15	7.98	8.22
Nd	35	34	35	35	36	35	37	36
Sm	7.22	7.29	7.41	7.36	7.54	7.46	8.06	8.28
Eu	2.21	2.25	2.23	2.11	2.37	2.26	2.61	2.55
Gd	5.94	5.90	5.87	5.87	6.04	6.22	6.76	7.00
Tb	0.87	0.83	0.76	0.78	0.76	0.78	0.88	0.84
Dy	3.50	3.71	3.57	3.42	3.52	3.64	4.11	4.19
Ho	0.63	0.64	0.60	0.63	0.66	0.68	0.75	0.66
Er	1.52	1.28	1.41	1.49	1.59	1.47	1.52	1.71
Tm	0.17	0.17	0.18	0.18	0.19	0.18	0.25	0.19
Yb	1.04	1.06	1.22	1.04	0.98	1.11	1.04	0.98
Lu	0.13	0.16	0.13	0.15	0.12	0.12	0.18	0.15
Hf	4.78	4.67	4.25	4.23	4.01	4.66	5.24	5.22
Ta	2.71	2.74	2.40	2.50	2.52	2.94	3.37	3.38
Pb	2.54	2.11	2.16	3.24	2.32	1.54	1.96	2.01
Th	2.09	2.06	1.61	1.61	1.58	1.55	2.51	2.53
U	0.52	0.60	0.28	0.31	0.25	0.27	0.95	1.17

Table 2. (Contd.)

Volcanic stages	Early Oligocene			Paleocene–Early Eocene				
Sample no.	YuKh-1/3	KhRM 2/1a	1727	YuKh-2/5	YuKh-2/6	YuKh-2/7	YuKh-2/8	YuKh-2/9
N E	43°30.173' 102°10.239'	43°18.700' 103°45.720'		43°14.669' 104°35.412'	43°14.367' 104°36.058'	43°14.591' 104°36.213'	43°15.815' 104°36.387'	43°16.766' 104°35.133'
SiO ₂	49.39	51.37	51.25	48.60	48.21	48.95	49.11	48.48
TiO ₂	2.39	2.53	2.76	2.04	2.18	2.14	2.02	2.13
Al ₂ O ₃	14.24	14.46	13.42	14.92	15.44	15.40	15.12	15.33
Fe ₂ O ₃	10.56	9.81	10.69	10.58	10.95	11.17	10.99	10.99
MnO	0.13	0.11	0.12	0.14	0.14	0.15	0.14	0.16
MgO	5.31	5.31	6.43	5.61	5.33	6.85	5.68	5.56
CaO	7.34	6.66	6.79	9.63	8.60	8.07	9.26	8.83
Na ₂ O	4.48	4.08	3.97	4.04	4.40	4.32	3.99	4.16
K ₂ O	2.99	3.24	3.37	1.36	1.64	1.55	1.49	1.64
P ₂ O ₅	0.79	0.97	0.84	0.43	0.52	0.51	0.44	0.50
LOI	2.32	1.30	0.37	2.40	2.57	0.90	1.79	2.25
Total	99.93	99.84	100.01	99.75	99.98	100.00	100.04	100.02
CIPW*								
Albite	25.7	33.0	29.0	27.0	26.9	28.1	27.9	26.9
Diopside	18.0	12.6	15.9	22.4	18.4	15.4	20.2	18.7
Nepheline	7.1	1.1	2.5	4.4	6.2	4.7	3.5	4.9
Olivine	12.0	12.8	14.1	11.6	12.7	16.5	12.7	13.1
Ti	12625	15525	14045	12257	12650	12283	12310	13023
Cr	105	92	133	125	110	115	122	131
Co	32	33	34	42	40	42	41	42
Ni	69	64	76	106	97	109	108	109
Rb	36	51	30	25	29	45	26	26
Sr	899	1239	1027	616	736	689	595	695
Y	17	19	17	21	20	21	22	21
Zr	208	209	193	176	205	196	176	193
Nb	57	83	59	37	46	43	37	44
Ba	605	748	655	356	412	401	359	466
La	31	28	23	22	29	28	23	27
Ce	69	68	57	48	58	59	50	57
Pr	8.68	9.73	8.15	5.70	7.01	7.15	5.96	6.65
Nd	39	45	39	24	29	29	26	29
Sm	8.39	9.02	8.19	5.71	6.21	5.87	5.86	6.54
Eu	2.66	2.66	2.55	1.88	2.13	2.08	1.96	2.11
Gd	6.94	6.89	6.66	5.74	6.19	6.37	5.79	6.23
Tb	0.87	0.88	0.92	0.85	0.95	0.87	0.88	0.92
Dy	4.23	4.27	3.89	4.37	4.48	4.96	4.91	4.56
Ho	0.69	0.71	0.65	0.86	0.84	0.96	0.94	0.95
Er	1.53	1.54	1.56	2.13	1.95	2.27	2.34	2.44
Tm	0.19	0.19	0.19	0.28	0.28	0.33	0.36	0.29
Yb	1.01	1.08	0.96	1.51	1.54	2.00	1.90	1.79
Lu	0.16	0.14	0.13	0.23	0.21	0.23	0.24	0.23
Hf	5.42	4.73	4.48	4.42	5.05	5.09	4.49	4.95
Ta	3.77	5.03	2.94	2.32	2.93	2.89	2.39	2.78
Pb	1.80		2.33	3.15	4.58	3.29	2.07	1.94
Th	2.52	1.75	1.22	2.90	3.79	4.06	3.69	4.04
U	0.74	0.44	0.30	0.93	1.05	1.27	1.09	1.37

Table 2. (Contd.)

Volcanic stages	Paleocene–Early Eocene							
Sample no.	YuKh-2/10	ULN2/1	ULN2/2	ULN2/3a	1726	1739	YuKh-4/1	YuKh-4/2
N E	43°13.733' 104°28.015'		43°15.936' 104°36.458'		43°17.489' 104°04.280'	44°04.254' 103°40.536'	43°33.547' 104°44.159'	43°33.502' 104°44.626'
SiO ₂	48.65	47.44	48.28	47.85	48.01	48.66	48.13	46.19
TiO ₂	2.09	2.01	2.08	2.07	2.16	2.17	2.30	2.66
Al ₂ O ₃	15.78	15.05	15.42	15.31	15.12	14.39	15.24	14.86
Fe ₂ O ₃	11.20	11.03	10.90	10.98	10.40	11.01	11.32	12.58
MnO	0.15	0.14	0.14	0.14	0.14	0.14	0.14	0.15
MgO	6.84	5.71	5.98	6.29	4.90	6.58	5.36	6.90
CaO	8.33	9.76	8.81	8.95	9.42	8.97	8.51	8.21
Na ₂ O	4.46	4.50	4.14	3.97	3.91	3.71	4.46	4.34
K ₂ O	1.81	1.59	1.63	1.59	1.96	1.43	1.75	2.11
P ₂ O ₅	0.53	0.50	0.52	0.51	0.45	0.41	0.57	0.63
LOI	0.37	2.48	2.14	2.44	3.64	2.47	2.00	1.22
Total	100.21	100.21	100.04	100.10	100.11	99.94	99.77	99.85
CIPW*								
Albite	23.8	20.3	25.9	24.8	23.8	28.5	26.0	17.5
Diopside	16.6	24.5	18.2	18.3	22.2	19.7	18.7	18.2
Nepheline	7.6	10.1	5.3	5.2	5.7	2.0	6.8	10.7
Olivine	16.0	11.4	13.9	14.6	10.1	14.6	12.8	16.6
Ti	12544	11976	12330	12343	11956	12732	14131	14906
Cr	156	104	110	107	167	144	103	125
Co	40	41	40	43	37	43	42	44
Ni	107	41	61	34	42	90	98	98
Rb	39	26	26	26	31	13	37	44
Sr	696	689	725	731	661	482	769	781
Y	22	22	23	24	19	20	21	21
Zr	196	191	200	197	166	170	229	255
Nb	45	40	41	41	39	33	52	59
Ba	422	440	421	437	421	304	572	470
La	26	25	26	26	20	18	30	36
Ce	57	50	53	53	42	39	64	76
Pr	6.86	6.16	6.41	6.46	5.22	4.81	7.61	9.04
Nd	29	26	27	27	23	21	32	37
Sm	6.81	5.80	5.97	5.98	5.48	4.97	7.86	7.96
Eu	2.09	1.93	2.02	1.90	1.77	1.79	2.29	2.62
Gd	6.36	5.47	5.70	5.72	5.17	4.78	6.60	7.44
Tb	0.93	0.81	0.81	0.83	0.83	0.78	1.00	1.10
Dy	4.91	4.40	4.51	4.58	4.09	3.84	5.09	5.01
Ho	0.92	0.78	0.82	0.82	0.72	0.75	0.92	0.80
Er	2.36	1.94	2.01	2.03	1.87	2.00	2.16	1.86
Tm	0.36	0.27	0.26	0.27	0.28	0.27	0.28	0.27
Yb	1.89	1.55	1.57	1.57	1.47	1.54	1.64	1.40
Lu	0.21	0.22	0.21	0.22	0.20	0.20	0.23	0.19
Hf	5.05	4.36	4.53	4.57	4.01	3.77	5.87	6.11
Ta	2.92	2.50	2.60	2.62	2.10	1.84	3.29	3.77
Pb	3.20				2.67	3.63	2.60	2.84
Th	3.79	3.18	3.32	3.27	2.28	1.90	4.17	4.82
U	1.24	0.92	0.85	0.86	0.41	0.55	1.19	1.42

Table 2. (Contd.)

Volcanic stages	Paleocene–Early Eocene			Late Cretaceous				
Sample no.	YuKh-4/3	YuKh-4/4	ULN1/2a	ULN3/1a	YuKh-3/1	YuKh-3/2	YuKh-3/3	1723
N E	43°32.320' 104°44.357'	43°32.200' 104°44.387'		43°13.726' 104°28.072'	42°36.048' 104°01.644'	42°36.149' 104°01.501'	42°36.215' 104°01.033'	42°29.997' 103°56.818'
SiO ₂	49.41	50.06	50.10	48.73	49.61	49.12	49.56	50.13
TiO ₂	2.15	2.27	2.11	2.05	2.07	2.08	2.07	2.13
Al ₂ O ₃	14.75	15.01	14.97	15.77	15.17	15.30	15.25	15.19
Fe ₂ O ₃	11.06	10.78	10.86	11.16	10.24	10.26	10.18	10.42
MnO	0.14	0.14	0.14	0.15	0.13	0.13	0.13	0.14
MgO	5.70	5.43	6.33	7.48	4.85	5.15	4.74	6.80
CaO	9.18	8.40	8.48	7.99	8.37	8.65	8.70	6.42
Na ₂ O	3.96	3.96	3.67	4.11	4.31	4.21	4.23	4.53
K ₂ O	1.77	2.15	1.71	1.75	2.58	2.50	2.58	2.71
P ₂ O ₅	0.43	0.47	0.43	0.53	0.51	0.51	0.51	0.55
LOI	1.59	1.31	1.22	0.42	2.29	2.10	2.13	0.99
Total	100.13	99.99	100.02	100.14	100.11	100.00	100.08	100.01
CIPW*								
Albite	26.9	28.8	31.1	25.7	25.0	23.3	24.1	27.9
Diopside	21.3	18.2	16.6	13.7	20.1	20.3	20.9	12.5
Nepheline	3.8	2.8	0.2	5.0	6.6	7.1	6.7	5.8
Olivine	12.3	12.4	14.9	18.2	10.5	10.9	9.8	16.7
Ti	12178	13699	12930	11601	12962	12491	12647	12448
Cr	138	148	118	114	123	123	104	98
Co	40	42	39	39	38	37	37	40
Ni	99	110	39	37	81	78	85	44
Rb	33	43	33	37	54	39	47	54
Sr	567	663	629	736	783	760	749	764
Y	19	21	23	25	21	20	20	25
Zr	165	167	176	199	201	196	194	236
Nb	35	40	36	42	38	37	37	41
Ba	355	500	407	431	460	445	433	466
La	18	21	19	25	25	24	23	27
Ce	40	46	42	52	56	54	55	60
Pr	5.14	5.96	5.38	6.42	7.28	6.87	7.10	7.78
Nd	24	27	24	27	30	29	29	33
Sm	5.46	6.74	5.69	6.05	6.61	6.76	6.94	6.98
Eu	1.88	2.13	1.96	1.98	2.13	2.15	2.21	2.26
Gd	5.73	6.03	5.54	5.74	6.09	5.95	5.73	6.31
Tb	0.84	1.00	0.81	0.84	0.99	0.89	0.89	0.89
Dy	4.35	4.94	4.43	4.53	4.62	4.61	4.52	4.73
Ho	0.75	0.83	0.79	0.85	0.87	0.84	0.77	0.84
Er	1.87	1.82	1.92	2.12	1.73	2.05	2.06	2.02
Tm	0.24	0.28	0.25	0.29	0.27	0.26	0.26	0.26
Yb	1.52	1.50	1.49	1.65	1.56	1.58	1.54	1.49
Lu	0.21	0.20	0.20	0.24	0.19	0.20	0.22	0.21
Hf	4.27	4.45	4.27	4.46	5.26	5.12	4.99	5.49
Ta	2.16	2.70	2.29	2.53	2.33	2.37	2.34	2.64
Pb	2.91	2.04			3.22	2.12	1.85	
Th	2.37	2.12	2.23	3.10	2.66	2.50	2.42	2.64
U	0.82	0.57	0.68	0.86	0.97	1.35	0.81	0.81

Table 2. (Contd.)

Volcanic stages	Late Cretaceous							
Sample no.	GKhN1/3a	GKhN1/4a	GKhN1/5a	GKhN1/6a	1740/1	1740/3	1741	1742
N E		42°35.578' 104°02.427'	42°36.009' 104°01.238'		42°16.989' 105°43.531'	42°17.866' 105°45.251'	42°15.780' 105°52.918'	42°20.495' 105°52.812'
SiO ₂	50.79	51.48	49.74	48.78	43.09	43.90	44.17	44.13
TiO ₂	2.15	2.12	2.13	2.11	2.22	2.09	1.97	2.12
Al ₂ O ₃	15.86	15.68	15.33	15.09	13.61	14.21	13.91	13.95
Fe ₂ O ₃	10.53	10.76	10.03	9.82	12.68	12.43	12.09	12.63
MnO	0.13	0.13	0.12	0.13	0.21	0.22	0.22	0.24
MgO	4.93	5.16	5.12	5.25	7.04	7.63	8.24	6.54
CaO	7.55	7.25	8.00	8.30	10.91	10.43	10.30	9.96
Na ₂ O	4.10	3.90	3.82	3.93	5.03	4.83	5.10	5.47
K ₂ O	2.65	2.49	2.64	2.63	1.48	1.94	1.29	1.93
P ₂ O ₅	0.54	0.53	0.56	0.56	0.93	0.93	0.86	1.09
LOI	0.74	0.58	2.56	3.13	2.52	1.15	1.74	1.87
Total	99.97	100.08	100.05	99.73	99.72	99.76	99.89	99.93
CIPW*								
Albite	29.7	33.2	27.7	24.5	5.3	4.5	8.8	7.3
Diopside	14.1	12.1	16.2	18.5	32.1	28.5	28.5	29.0
Nepheline	2.9		2.9	5.4	20.9	20.0	19.1	21.6
Olivine	12.8	13.6	12.1	11.5	12.9	14.8	15.9	13.0
Ti	11933	13295	12219	12907	12131	11626	10771	11951
Cr	95	94	132	84	106	126	175	124
Co	37	35	33	33	37	37	39	34
Ni	38	20	78	35	34	33	95	49
Rb	46	44	39	48	28	49	17	56
Sr	778	747	766	818	1428	1431	1477	1547
Y	22	22	19	23	37	37	36	41
Zr	202	205	199	226	445	425	434	519
Nb	35	37	40	40	99	98	105	116
Ba	536	440	459	490	746	843	849	989
La	23	22	25	27	80	83	84	91
Ce	51	49	55	59	156	160	161	176
Pr	6.66	6.42	6.96	7.47	18.06	18.21	18.25	20.45
Nd	29	28	30	32	68	69	69	78
Sm	6.45	6.16	6.41	6.75	12.36	12.31	12.03	13.99
Eu	2.05	1.93	2.16	2.32	3.79	3.74	3.60	4.14
Gd	5.67	5.62	5.86	6.04	10.27	10.11	10.06	11.24
Tb	0.80	0.78	0.86	0.85	1.43	1.39	1.38	1.59
Dy	4.29	4.26	4.25	4.42	7.28	7.40	7.18	8.25
Ho	0.76	0.77	0.78	0.77	1.26	1.34	1.27	1.48
Er	1.87	1.86	1.77	1.83	2.92	3.18	2.94	3.45
Tm	0.24	0.25	0.25	0.24	0.38	0.41	0.38	0.46
Yb	1.49	1.46	1.13	1.36	2.12	2.29	2.13	2.68
Lu	0.20	0.20	0.15	0.19	0.28	0.32	0.29	0.36
Hf	4.77	4.73	4.89	5.15	9.19	8.69	8.86	10.71
Ta	2.25	2.21	2.29	2.48	6.79	6.46	6.81	7.81
Pb			1.69					
Th	2.33	2.29	2.09	2.55	10.20	10.16	10.45	13.43
U	1.16	1.00	0.94	1.24	2.58	2.62	2.72	2.91

Table 2. (Contd.)

Volcanic stages	Late Cretaceous					
Sample no.	YuKh-2/1	YuKh-2/2	YuKh-2/3	YuKh-2/4	1721/1	1721/2
N E	43°13.570' 103°59.843'	43°13.611' 103°59.758'	43°13.279' 103°59.109'	43°13.216' 103°59.042'	42°48.049' 102°05.643'	42°47.551' 102°27.384'
SiO ₂	49.04	48.84	48.94	47.75	46.12	47.14
TiO ₂	2.07	2.54	2.04	2.20	2.01	1.79
Al ₂ O ₃	15.20	15.25	15.01	15.36	15.06	15.68
Fe ₂ O ₃	10.72	10.29	11.25	11.70	11.78	11.05
MnO	0.15	0.14	0.15	0.16	0.21	0.19
MgO	5.36	4.33	6.26	7.11	7.12	8.49
CaO	9.37	9.51	9.31	8.62	8.96	8.60
Na ₂ O	4.17	3.92	4.00	4.31	4.62	4.58
K ₂ O	1.77	2.52	1.53	1.07	1.76	0.74
P ₂ O ₅	0.47	0.56	0.42	0.49	1.00	0.56
LOI	1.72	2.08	1.19	1.18	1.38	1.56
Total	100.03	99.97	100.10	99.94	100.02	100.38
CIPW*						
Albite	25.3	22.4	25.6	26.6	17.7	24.9
Diopside	21.8	22.8	20.7	16.7	18.9	15.6
Nepheline	5.7	6.2	4.7	5.6	11.9	7.8
Olivine	11.2	8.0	13.8	17.1	16.7	19.8
Ti	10882	14324	10412	11617	11288	10474
Cr	146	122	150	133	85	126
Co	37	33	40	41	35	42
Ni	78	40	93	100	29	33
Rb	27	37	25	39	54	21
Sr	574	663	557	611	1038	661
Y	20	22	19	21	38	28
Zr	164	203	164	185	304	182
Nb	39	48	36	41	93	47
Ba	407	472	354	399	762	710
La	22	22	20	25	60	23
Ce	43	45	39	49	114	48
Pr	5.39	5.91	4.92	6.14	13.20	5.95
Nd	23	27	22	27	52	26
Sm	5.56	6.74	5.06	6.52	9.90	5.72
Eu	1.80	2.09	1.72	2.16	2.99	1.87
Gd	5.51	6.68	5.08	5.60	8.48	5.50
Tb	0.78	0.88	0.76	0.92	1.21	0.84
Dy	4.06	4.53	4.21	4.69	6.66	4.95
Ho	0.79	0.88	0.77	0.81	1.25	0.96
Er	1.79	2.12	1.68	2.01	3.21	2.47
Tm	0.28	0.31	0.25	0.24	0.46	0.35
Yb	1.41	1.44	1.50	1.42	2.72	2.10
Lu	0.23	0.23	0.20	0.23	0.41	0.32
Hf	4.62	4.76	4.23	4.87	6.47	4.25
Ta	2.53	2.98	2.32	2.88	5.66	2.94
Pb	2.90	3.64	2.42	2.07		
Th	2.62	2.71	2.54	3.06	7.58	2.86
U	0.53	0.74	0.99	0.96	1.82	0.72

* Content of standard minerals is given in mass %.

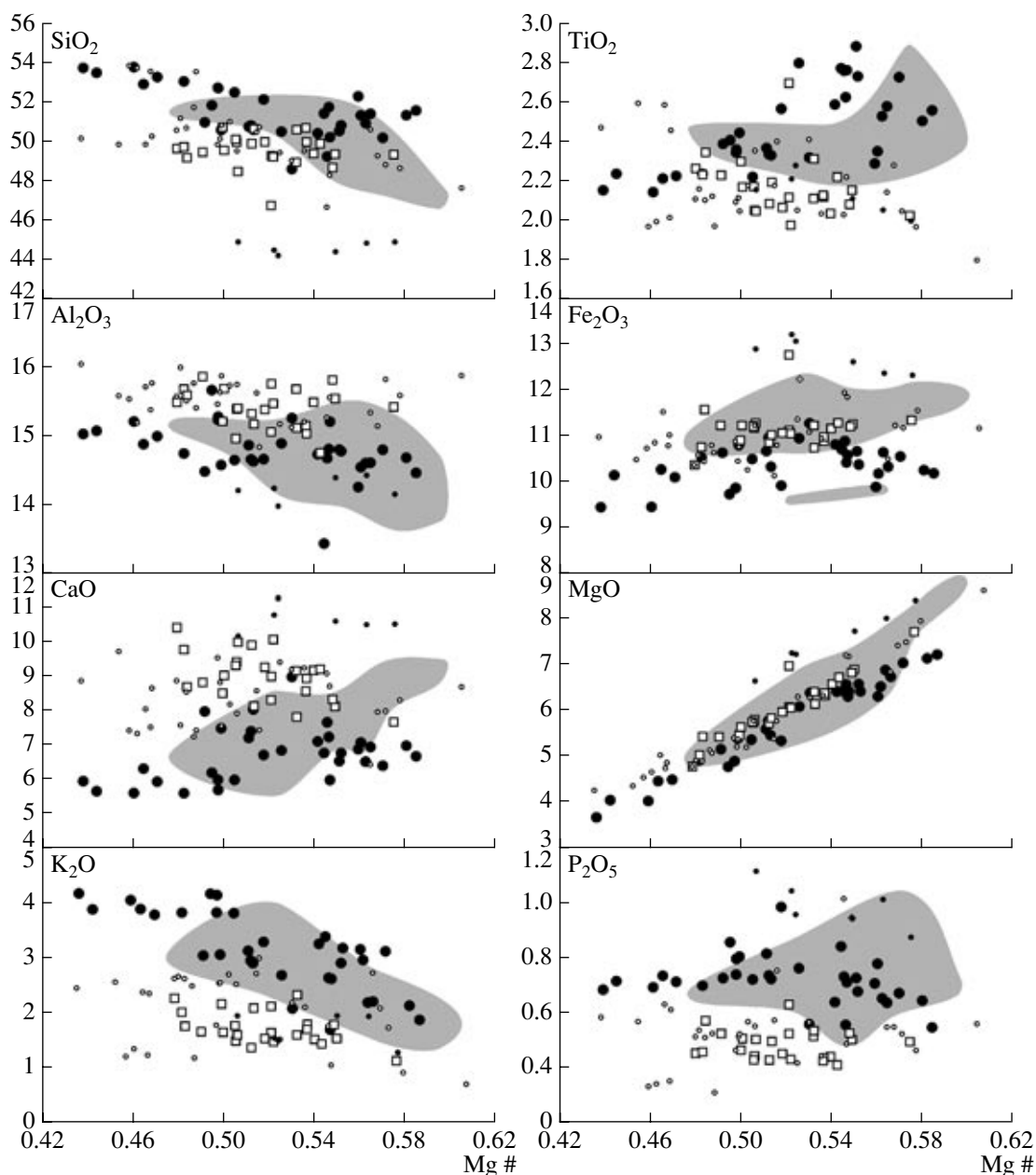


Fig. 5. Major element distribution with respect to the Mg number value in different-age rocks of the Gobi Tien Shan volcanic area. The denotations are the same as in Fig. 4.

similar composition. Positive linear relationships between the Mg number value and the TiO_2 , Fe_2O_3 , MgO , CaO , and P_2O_5 content, and inverse correlation with Al_2O_3 , Na_2O , and K_2O , which can be related to fractioning of olivine, pyroxene, ore minerals, and apatite, are typical of these rocks.

The presence of nepheline, the percentage of which varies from 0 to 10% and reaches 20% in melanephelinites including standard albite (<10%), is typical of the standard rock composition in the volcanic area (Fig. 6a). The standard content of olivine and diopside is approximately identical. Variations in nepheline with

respect to diopside and olivine are related to differentiation processes. The relative amount of standard olivine and diopside decreases and the fraction of standard nepheline increases with increasing differentiation, which is registered based on a decrease in the Mg number value (Fig. 6b).

Rocks of different ages are generally characterized by similar contents of major elements, although certain systematic differences can be observed, which apparently indicate that the sources or conditions of melting and melt differentiation were different. These differences are most contrasting when the Late Cretaceous

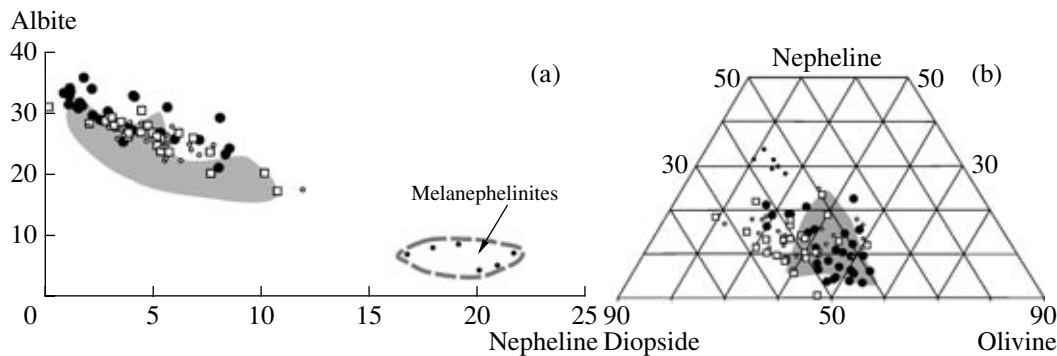


Fig. 6. Diagrams for the standard compositions of rocks in the Gobi Tien Shan volcanic area. The denotations are the same as in Fig. 4.

and Early Oligocene rocks are compared, whereas the compositions of the Paleocene–Early Eocene rocks are intermediate. Thus, the Late Cretaceous rocks are relatively less rich in titanium and potassium and are more enriched in alumina (Fig. 5). Lower contents of P_2O_5 are also typical of these rocks. The Early Oligocene rocks differ from other rocks in their high contents of K_2O (to 4 mass %) and in higher percentages of SiO_2 (50.2–53.8 mass %) at almost the same range of Mg number values.

For comparison, Fig. 5 demonstrates the field of compositions of the Late Oligocene–Miocene volcanites from the Lake Valley. These rocks are generally characterized by the same regularities of change in composition that are observed in the rocks of different age from the Gobi Tien Shan area but are closest to the composition of the Early Oligocene rocks.

Geochemistry. The concentrations of trace elements in the rocks from the Gobi Tien Shan area are given in Table 2. In spite of the above differences in the content of major oxides, the rocks of all age groups are characterized by similar contents of incompatible elements (see spidergrams in Fig. 7). The spectrum of distribution of these elements is close to the spectrum of OIB basalts and differs from the latter spectrum by a lower content of most elements except Rb, Ba, Nb, Ta, Sr, Gd, and Tb. The content of these elements is close to, or even higher than, their content in OIB basalts. The presence of a clearly defined Ta–Nb maximum, typical of intraplate magmatism sources, is of special importance. The Late Cretaceous and Paleocene–Early Eocene volcanites, the average compositions of which are almost the same, are most similar to one another in their contents of incompatible elements. The Early Oligocene lavas differ from these volcanites in a more fractionated distribution of REEs, a higher content of Rb, Ba, and Sr, and a lower concentration of Th, U, and Pb. The spidergrams compare the compositions of the Late Cretaceous rocks from the Gobi Altai zone rift zone and the Late Oligocene–Miocene from the Lake Valley. All compared rock groups have evidently similar compositions, which thereby indicates that the compositions of

the SKVR magmatism sources were similar in the Late Cretaceous–Late Cenozoic.

Figure 8 shows the patterns in the distribution of the trace elements with respect to Th, which was selected as a differentiation index. Inverse correlation between the contents of Th and compatible elements (Ni, Cr, Co) is observed, which indicates that the role of olivine and pyroxene fractioning during melt evolution was important. However, the graphs as a rule do not show rigorous linear dependences, and the composition points usually do not form extended swarms, indicating that the element distribution in a melting substrate was nonuniform. Such a nonuniformity is most pronounced in the Late Cretaceous rocks (Fig. 8; trends 1, 2) but mostly is due to geographic factors. Therefore, variations in the composition of rocks from a specific lava field are linear and correspond to differentiation of a homogeneous melt. The presented graphs (Fig. 8) also evidently demonstrate schematic differences between the compositions of rocks from different age groups. The Early Oligocene rocks are most different. They differ in their higher contents of Ba and Sr and lower contents of Cr, Th, U, and Yb. Probable negative correlation between Th and Sr, Ba, and Pb may indicate that feldspars participated in the fractioning processes. We also note that, on the diagrams (Fig. 8), the data points for different-age basalts are grouped in the series of subparallel trends responsible for melt evolution lines. These trends evidently demonstrate a certain difference in the geochemical characteristics of magma sources of different age groups.

Table 3 presents the variations in a number of indicator relationships. These variations mostly show differences which are not larger than two- or threefold between the extreme values. In the coordinates of correlation between the pairs of relationships between incompatible elements (Fig. 9), the compositions of the Gobi Tien Shan area form compact groups of data points with insignificant variations in the element relationships and considerable overlaps of the compositions of different-age rocks. Certain systematic differences between the Early Oligocene and other rocks,

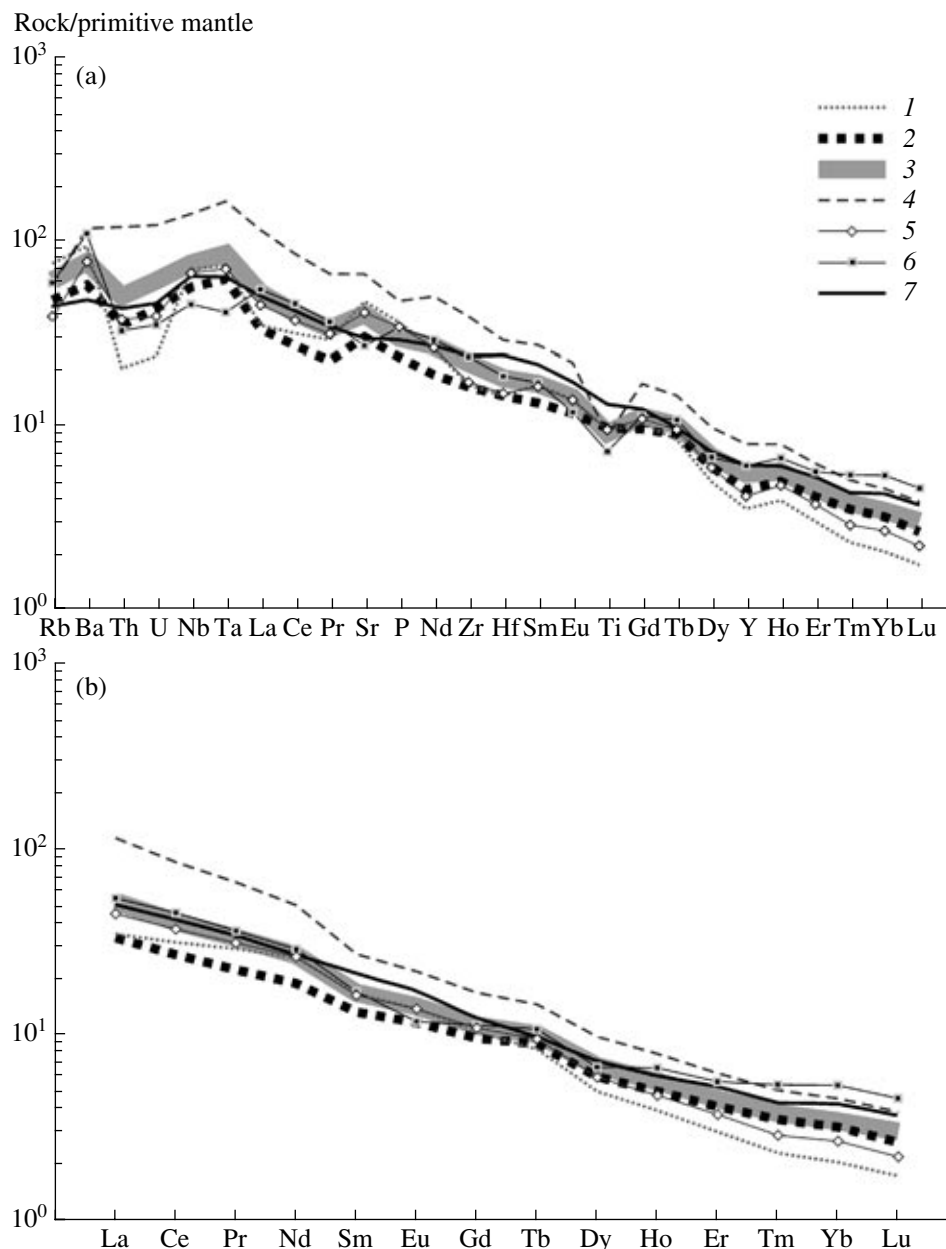


Fig. 7. Spidergrams of distribution of (a) incompatible and (b) rare earth elements in the mean compositions of different-age rocks in the Gobi Tien Shan volcanic area. (1)–(6) Rock age groups: (1) Early Oligocene; (2) Paleocene–Early Eocene, (3)–(4) Late Cretaceous ((3) basaltoids, (4) melaneophelinites), (5) Late Oligocene–Miocene in the Lake Valley (Khangai area), (6) Late Cretaceous of the Gobi Altai rift zone, (7) OIB [27].

shown by the Zr/Nb, Th/Ta, and Nb/U ratios, are, first of all, related to the fact that these rocks are relatively depleted in Th and U and enriched in Nb. In spite of the above-mentioned generally insignificant differences in the composition of the rocks of different age, all these data indicate that the sources of melts in the volcanic area are similar and, specifically, the composition of these sources is close to that of ocean island basalts (see the graphs).

The conditions of melting were also rather similar. According to the graphs (Figs. 9g, 9h), magma was

formed in mantle containing garnet. Lines in Fig. 9g indicate the variations in the melt composition at a given change in the degree of melting and content of garnet in the mantle source according to [25, 26]. The data points for the composition of rocks in the Gobi Tien Shan area indicate that the sources of these rocks were garnet peridotites, in which the content of garnet reached 6% and the degree of melting was generally about 1–1.5%. In this case the melting conditions were rather stable and insignificantly varied only for the Late Cretaceous rocks. Trachybasalts predominated in the

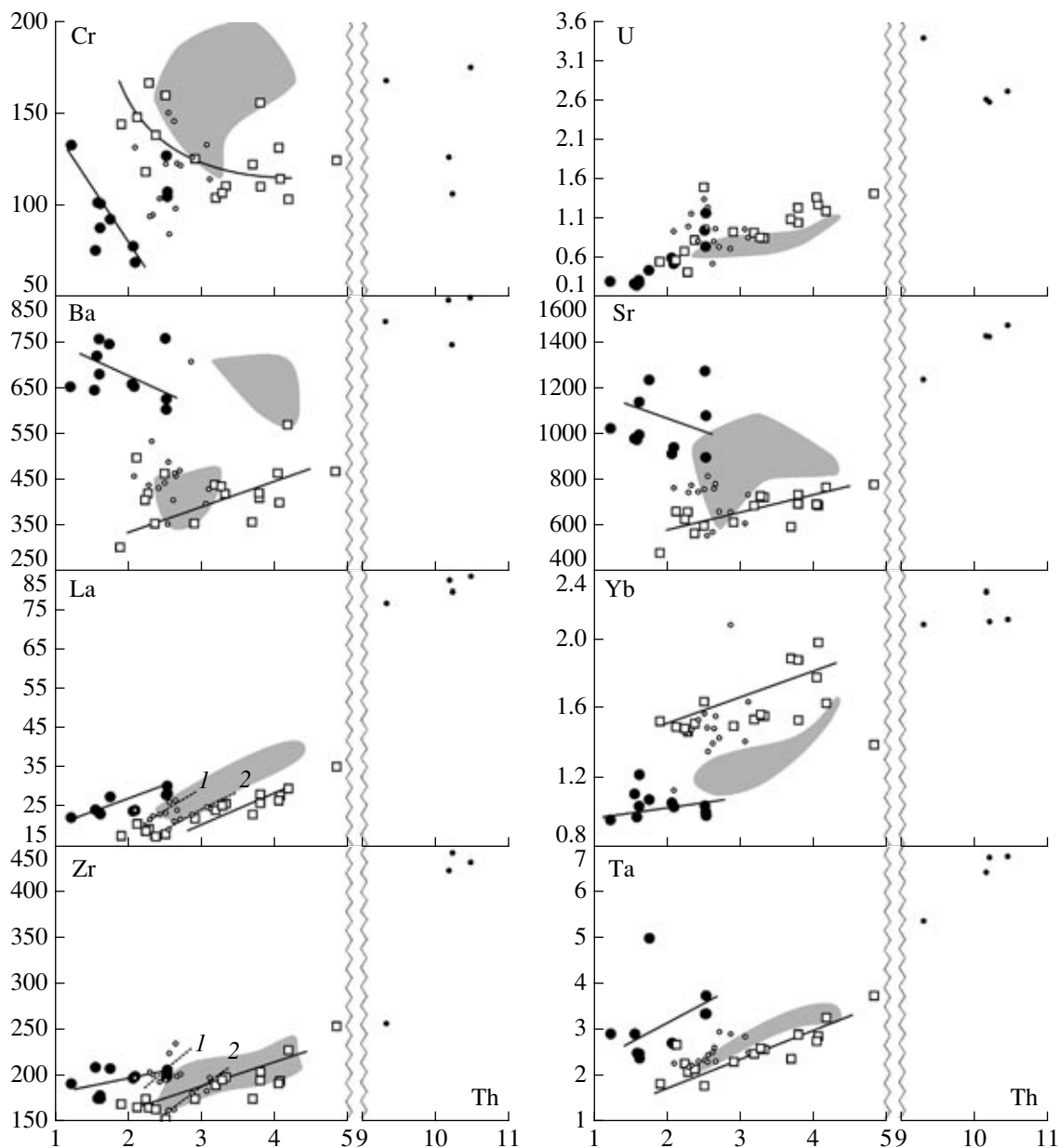


Fig. 8. Trace element distribution with respect to Th in different-age rocks of the Gobi Tien Shan volcanic area. The denotations are the same as in Fig. 4. Trends 1 and 2 reflect regional variations in the lava composition: (1) Bordzongiin-Gobi depression, (2) Khurmen graben. Solid lines correspond to trends of melt composition evolution for the Early Oligocene and Paleocene–Early Eocene rock groups.

composition of these rocks and were formed from magmas that were melted from the garnet-containing mantle (about 2%) at a degree of melting of about 1.5%. Magmas of melaneophelinites were apparently formed at a larger depth (the garnet content is ~5.5–6%) and at a lower degree of melting (about 1%). The conditions of melt formation during the Paleocene–Early Eocene stage of volcanism were close to those during the formation of the Late Cretaceous trachybasalt magmas, whereas at the Early Oligocene magmas were formed under the conditions of formation of melaneophelinite melts.

Isotope composition. The data on the isotope composition of the rocks in the Gobi Tien Shan volcanic area are given in Table 4 and are reflected in Fig. 10. The data points in Fig. 10 indicate that the isotope composition is substantially variable. For the rocks in this area, the observed composition variations generally correspond to the trend, which is traced from the quadrant of depleted mantle sources to a region slightly enriched in radiogenic Sr and depleted in radiogenic Nd, and can be considered to be the result of the mixing of a moderately depleted PREMA mantle and a source

Table 3. Variations in the ratio of incompatible elements in rocks of different age groups in the Gobi Tien Shan volcanic area

Ratio of elements	Age groups, intervals (Myr)		
	30–37	47–62	71–88
Zr/Nb	2.5–4.4	4.1–5.2	3.3–5.8
Zr/Hf	38–45	37–45	36–47
Nb/Ta	15–20	15–19	14–18
Nb/U	46–198	32–95	28–73
Ce/Y	3.3–3.9	1.8–3.7	1.7–4.5
Ce/Pb	16–38	11–29	12–32
Ce/U	54–255	42–101	40–82
Ba/Ce	8.8–12.9	6.2–10.9	5.3–14.7
Ba/Sr	0.58–0.76	0.56–0.75	0.52–1.07
Ba/Nb	9.0–17.6	8.0–12.4	7.5–15.4
Th/Ta	0.35–0.77	0.78–1.54	0.91–1.72
U/Th	0.16–0.46	0.18–0.35	0.20–0.54
U/Pb	0.10–0.58	0.15–0.71	0.18–0.64
La/Yb	19–31	12–25	11–39
(Tb/Yb) _n *	3.1–4.8	2.2–3.9	2.0–3.8

* The ratio is normalized to primitive mantle.

enriched in Nd. Such a source could be either the crust or the EM-I mantle enriched in Nd. The effect of crustal contamination on the formation of melt composition should generally be ruled out, since the Gobi Tien Shan area originated within the Hercynian block of Central Asia, the crust of which is of a juvenile origin, and, therefore, is characterized by mostly positive $\epsilon_{Nd}(T)$ values. In particular, for the last 100 Myr (the Late Cretaceous–Cenozoic), the isotope composition of this crust has changed to the lowest values of $\epsilon_{Nd}(T)$, which were nevertheless larger than -1 [5, 6, 24]. Therefore, the source enriched in Nd should, most probably, be compared with the enriched EM-I mantle.

Within the field of isotope compositions of the igneous rocks in the Gobi Tien Shan area, relatively homogeneous characteristics are typical of the Late Cretaceous and Paleocene–Early Eocene rocks, the figurative points of which are mainly grouped in the quadrant of depleted mantle sources in the region of moderately depleted PREMA mantle. In contrast to these rocks, the Early Oligocene rocks have mostly low values of $\epsilon_{Nd}(T)$, which correspond to the leading role of the enriched mantle source in the formation of the latter rocks. Two data points of this set deviate from the general trend of compositions toward high $\epsilon_{Sr}(T)$ values with a simultaneous growth of the $\epsilon_{Sr}(T)$ value, i.e., toward a source enriched in radiogenic Sr and relatively depleted in Nd. To elucidate the causes of a similar deviation, we consider specific features of correlation between the Sr and Nd isotope composition and the

concentration of these elements in rocks (Figs. 10b, 10c). The graphs indicate that the variations in the Nd isotope composition are almost independent of the Nd content, especially if these variations are considered within one age group. At the same time, a difference is traced in the isotope composition and Nd content between different age groups of basalts. Thus, the Early Oligocene basalts are characterized by smaller ϵ_{Nd} values and larger Nd content than the Paleocene–Early Eocene basalts. These data are in good agreement with data on rare elements (Fig. 8). The latter data indicate that the concentration of light REEs in the source of the Early Oligocene basalts is higher, which most likely corresponds to the enriched mantle parameters. On the other hand, a positive correlation is established between the Sr isotope composition and Sr content of the Early Oligocene rocks. The observed dependence is described by a broken line, the gentler and steeper branches of which correspond to the compositions of the mixing trends of the PREMA and EM-I sources and to the above compositions relatively enriched in radiogenic Sr, respectively. Such a behavior of these lines could indicate that an additional component appeared in the source of melts. The characteristics of this source are very specific since, according to the graph, the source should be enriched in radiogenic ($\epsilon_{Sr}(T) > +25$) and show high Sr contents (>1400 ppm) and increased Nd (~ 40 ppm). Such characteristics indicate that the crust cannot be a possible contaminant. Among the regional mantle sources, the corresponding parameters were typical of the sources of the Late Jurassic–Early Cretaceous carbonatite magmatism, which had high contents of Sr and REEs [12], as well as large $\epsilon_{Sr}(T)$ values (larger than $+20$) and almost zero $\epsilon_{Nd}(T)$ values [19]. These sources occupy the rightmost position in the field of compositions of the Late Jurassic–Early Cretaceous SKVR rocks (see Fig. 10a). Therefore, we can assume that, in addition to the EM-I mantle, EM-II mantle sources participated in the formation of the Early Oligocene magmatism. The latter sources produced carbonatite melts in the Early Cretaceous and apparently remained in the structure of the lithospheric mantle up to the Early Oligocene.

DISCUSSION

Geological Regularities of Intraplate Magmatism

The above data indicate that intraplate magmatism episodically manifested itself in the Gobi Tien Shan volcanic area from the end of the Late Cretaceous to the Early Cenozoic inclusive, i.e., during more than 50 Myr. In the geological history of Central Asia, this period was characterized by a platform regime [10, 21]; however, within the volcanic area, this regime was complicated by the processes of intraplate tectonics. As a result, the system of small grabens and normal fault belts, which are generally irregularly distributed over the area and form the pattern of a broken plate (Fig. 1), originated here. The distribution of these faults is con-

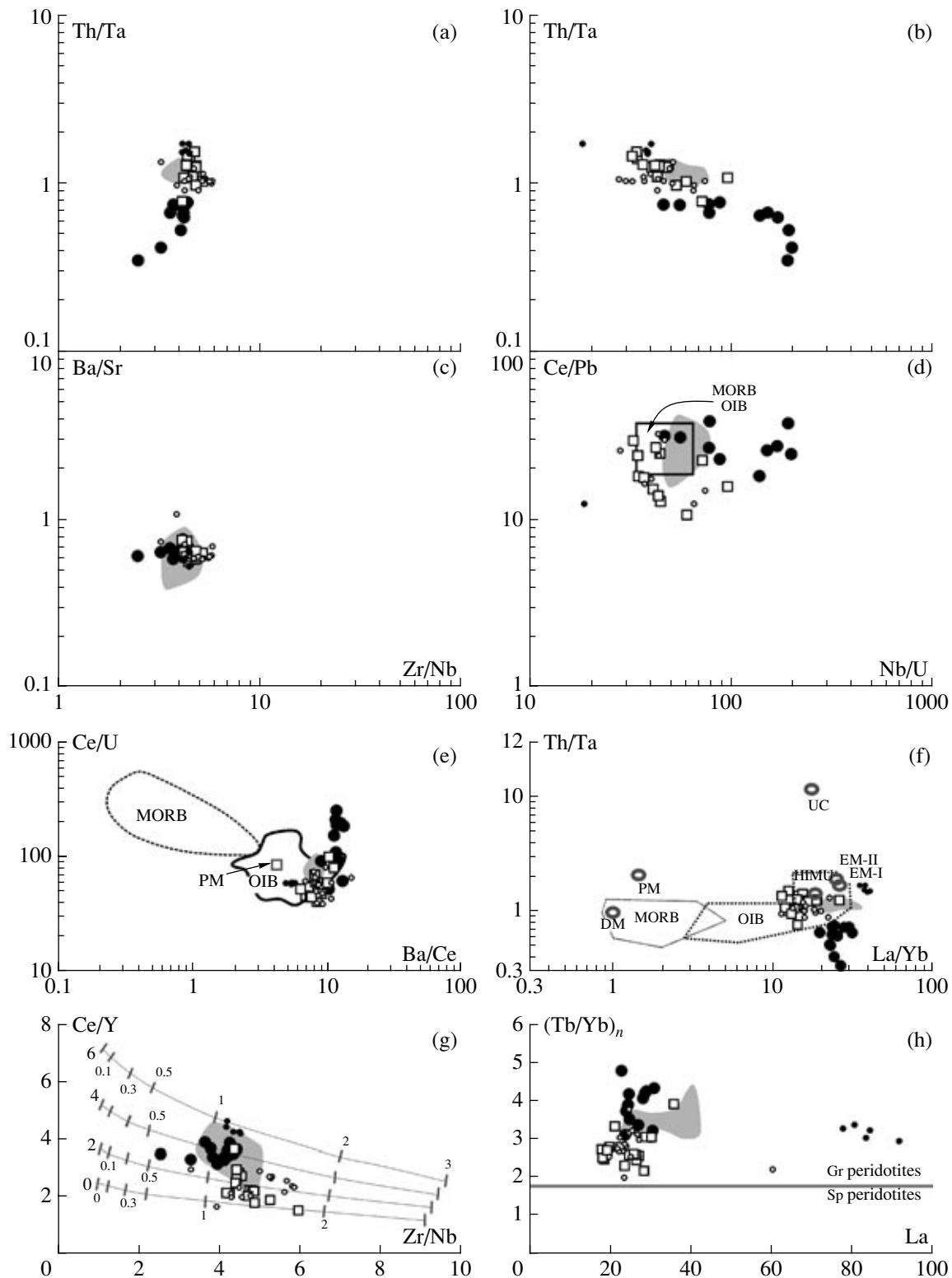


Fig. 9. Distribution of pair ratios of incompatible elements in the different-age rocks of the Gobi Tien Shan volcanic area. The denotations are the same as in Fig. 4. The fields of reference compositions and boundary values are given: according to [23], [28], and [25, 26] for panels (e), (f), and (g) and (h), respectively. Gray lines in panel (g) correspond to melting fractions; numerals from the left of trends show the percentage of garnet in the initial substratum. Fields of different-type basalts and certain mantle sources: (OIB) ocean islands, (MORB) mid-ocean ridges, (HIMU) source with high $^{238}\text{U}/^{204}\text{Pb}$, (EM-I) and (EM-II) enriched mantle, (DM) depleted mantle, (PM) primitive mantle, (UC) upper crust.

Table 4. Isotope composition (Sr, Nd) of igneous rocks in the Gobi Tien Shan volcanic area

Sample no.	Age, Ma	Rb, ppm	Sr, ppm	$^{87}\text{Rb}/^{86}\text{Sr}$	$^{87}\text{Sr}/^{86}\text{Sr}$	$(^{87}\text{Sr}/^{86}\text{Sr})_0$
YuKh-1/1	33	43.1	1370	0.0910	0.706317 ± 10	0.7063
1727	34	49.6	1153	0.1243	0.704756 ± 12	0.7047
1720/1	34	51.0	1031	0.1431	0.704977 ± 16	0.7049
NT-24/21	34	44.8	1238	0.1046	0.705539 ± 11	0.7055
NT-4/1	37	53.3	1114	0.1384	0.705047 ± 9	0.7049
YuKh-2/9	47	25.5	685	0.1078	0.704133 ± 8	0.7041
ULN-2/2	53	25.1	611	0.1187	0.704539 ± 14	0.7045
1743/1	47	23.7	548	0.1251	0.704302 ± 21	0.7042
1726	62	33.5	665	0.1459	0.704586 ± 12	0.7045
1739	60	22.6	516	0.1265	0.704097 ± 14	0.7040
YuKh-4/4	60	37.7	639	0.1706	0.704417 ± 3	0.7043
1741	75	15.5	1400	0.0320	0.703632 ± 17	0.7036
YuKh-3/1	75	39.6	759	0.1510	0.704602 ± 12	0.7044
YuKh-2/1	84	29.3	615	0.1377	0.704172 ± 12	0.7040
Sample no.	$\epsilon_{\text{Sr}}(\text{T})$	Sm, ppm	Nd, ppm	$^{147}\text{Sm}/^{144}\text{Nd}$	$^{143}\text{Nd}/^{144}\text{Nd}$	$\epsilon_{\text{Nd}}(\text{T})$
YuKh-1/1	25.7	7.49	37.5	0.1256	0.512337 ± 8	-5.5
1727	3.3	8.98	42.2	0.1286	0.512142 ± 8	-9.2
1720/1	6.4	7.73	38.1	0.1227	0.512073 ± 10	-10.7
NT-24/21	14.6	7.67	38.9	0.1238	0.512042 ± 11	-11.3
NT-4/1	7.3	7.81	37.7	0.1254	0.512165 ± 7	-8.8
YuKh-2/9	-5.4	5.71	26.5	0.1354	0.512798 ± 10	3.5
ULN-2/2	0.2	5.72	25.8	0.1339	0.512843 ± 7	4.4
1743/1	-3.2	5.11	21.8	0.1411	0.512856 ± 10	4.6
1726	0.4	5.38	23.4	0.1393	0.512730 ± 7	2.2
1739	-6.3	5.26	22.3	0.1426	0.512848 ± 8	4.4
YuKh-4/4	-2.2	5.73	25.0	0.1436	0.512686 ± 12	1.1
1741	-11.6	11.64	65.8	0.1070	0.512844 ± 10	4.9
YuKh-3/1	0.4	6.26	29.7	0.1323	0.512475 ± 11	-2.6
YuKh-2/1	-5.6	5.44	24.6	0.1387	0.512799 ± 13	3.7

trolled by fragments of large basement faults, which were intensified in the region of volcanism manifestation.

We especially emphasize that the period of formation of the Gobi Tien Shan volcanic area straddles the time interval between the development of the Gobi Altai rift zone (the end of the Late Jurassic–the beginning of the Late Cretaceous, or 150–90 Ma ago) and the formation of the Khangai (the Late Oligocene–Pleistocene–Holocene, <30 Ma ago) volcanic area. Taking into account the spatial conjugacy of these volcanic areas, we can speak about their relation to the formation of the same volcanic region. The latter was distinguished by us as SKVR [15, 16], the development of which has been traced for the last 150 Myr. The region

was formed as a series of bursts (or stages) of volcanic activity, which can be generally classified in periods of high (150–100 and 30–0 Ma ago) and low 100–30 Ma ago) productivity of volcanic processes.

The Gobi Tien Shan volcanic area developed during the period of low volcanic productivity, which combined three activity stages separated by intervals of volcanic quiescence that lasted about 10 Myr each. In spite of considerable time intervals between the stages of high volcanic activity, the character of volcanism remained almost unchanged in time. Volcanism manifested itself in the formation of basaltic plateaus as a result of outflows that were accompanied by explosive activity and formation of slag–bomb structures at the initial stages. Several (five–seven) flows are observed in

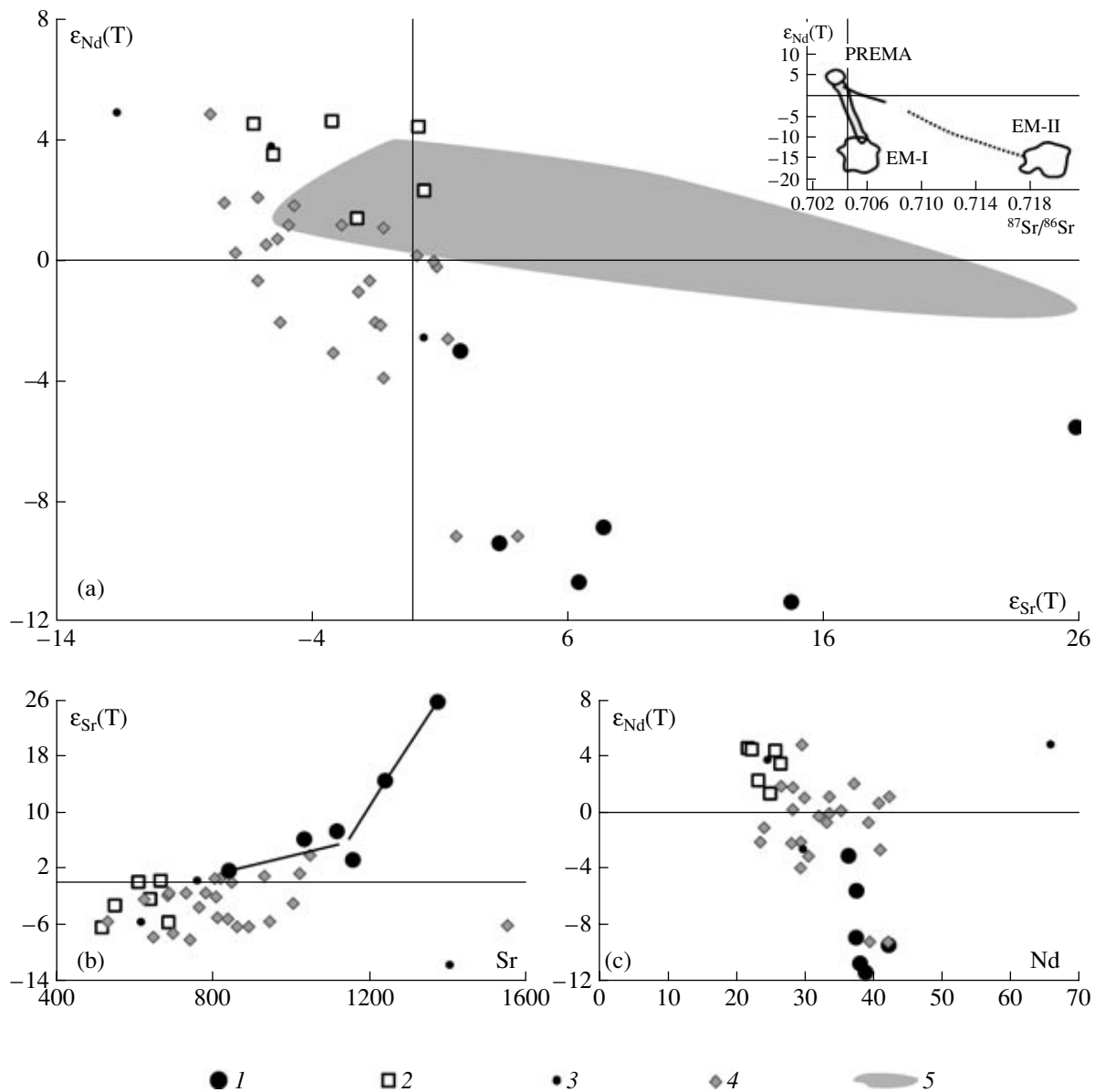


Fig. 10. Location of the composition of rocks in the Gobi Tien Shan volcanic area on the (a) $\epsilon_{\text{Nd}}-\epsilon_{\text{Sr}}$, (b) $\epsilon_{\text{Sr}}-\text{Sr}$, and (c) $\epsilon_{\text{Nd}}-\text{Nd}$ diagrams. (1)–(5) Compositions of rocks at different-age volcanic stages: (1) Early Oligocene, (2) Paleocene–Early Eocene, (3) Late Cretaceous, (4) Late Cenozoic rocks of the Khangai area, (5) field of rock composition of the Gobi Altai rift zone older than 90 Ma. The inset indicates the position of rocks of the South Khangai Late Mesozoic–Cenozoic volcanic region relative to the compositions of a moderately depleted PREMA mantle and of enriched EM-I and EM-II mantles [30].

the structure of every studied volcanic field, which most probably indicates that these fields and volcanic centers were formed during short periods. In addition to lava outflows, subvolcanic activity resulted in the formation of numerous dikes, volcanic domes, sills, laccolites, and stocks. Under the conditions of almost horizontal occurrence of the Late Cretaceous rocks and their slight present-day dissection, these formations are encountered in areas that were subjected to considerable denudation. Assuming that these formations correspond to the root zones of volcanic outflows, we can conclude

that the scales of initial distribution of volcanic fields, subsequently destroyed by erosion, were larger.

One of the most interesting features of magmatism in the Gobi Tien Shan area was migration of the centers of volcanism of different ages over this area (Fig. 11). The history of this migration can be traced at least from the beginning of the Late Cretaceous, i.e., from ~100 Ma ago, when volcanic activity within SKVR was localized in the eastern part of the Gobi Altai rift zone in the region of the Arts-Bogd range immediately north of the area. In the interval of 90–80 Ma ago, the volcanic centers gradually shifted

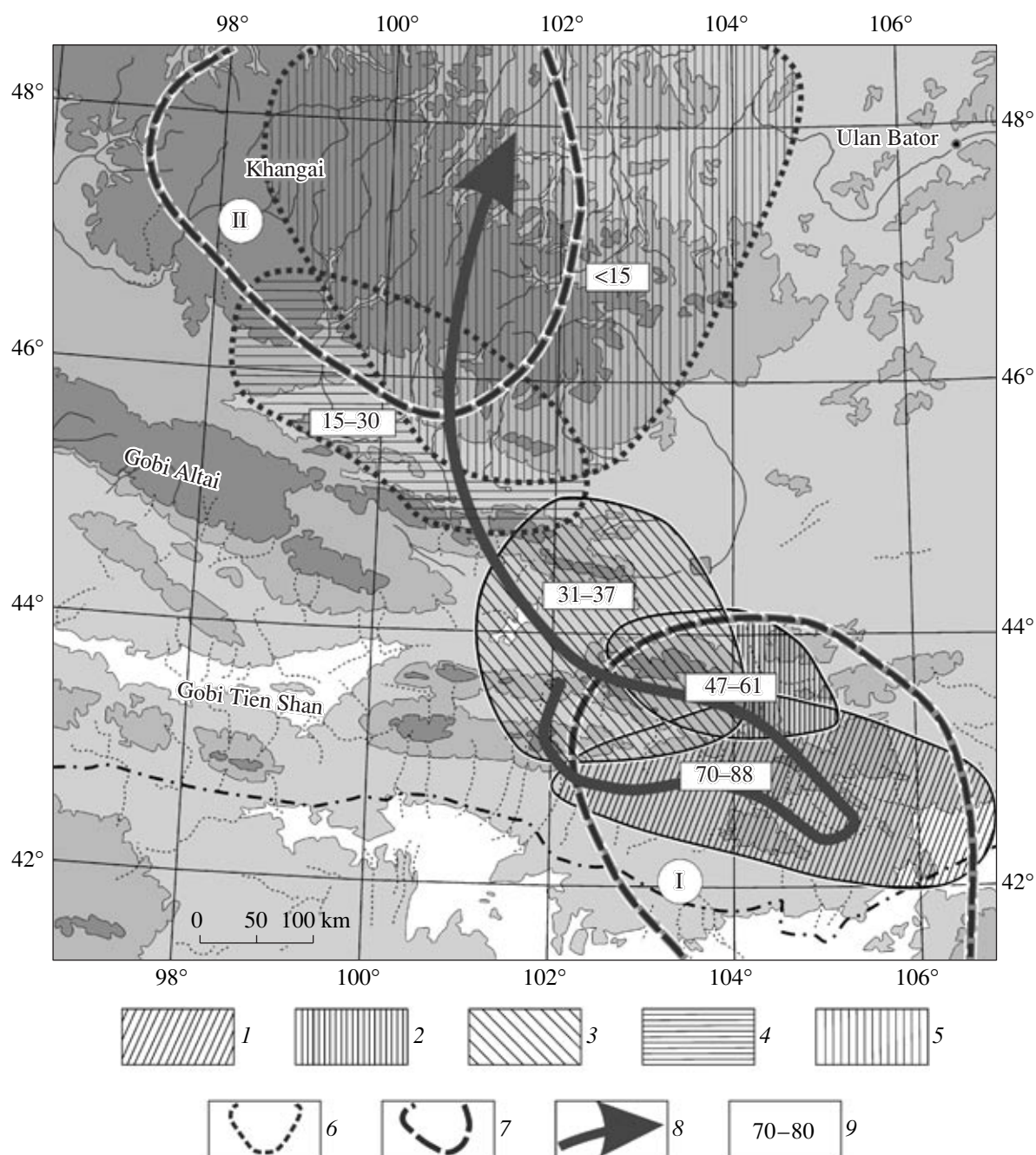


Fig. 11. Schematic migration of volcanic centers within SKVR for the last 90 Myr. (1)–(5) Areas of distribution of even-aged igneous rocks: (1) Late Cretaceous, (2) Paleocene–Early Eocene, (3) Early Oligocene, (4) Late Oligocene–Early Miocene, (5) Middle Miocene–Holocene; (6) boundaries of distribution of different-age lava field groups within the Khangai area; (7) boundaries of asthenospheric high projections associating with (I) Gobi Tien Shan and (II) Khangai volcanic areas; (8) trajectory of displacement of the distribution centers of different-age lava field groups; (9) age range of volcanic occurrences within isolated areas.

southward in the region of the Bordzongiin-Gobi desert, where the series of volcanic fields was formed about 75 Ma ago. In the interval from 62–47 Ma ago (the Paleocene–Early Eocene stage), the volcanic centers shifted northward, and the lava field fragments are distributed around the Gurban-Saikhan range. In the Early Oligocene, volcanic activity shifted northwest-

ward and occupied the Noen, Sevrei, and Arts-Bogd ranges. If we take into account the fact that the Lake Valley and southern slopes of Khangai were involved in the process of volcanism in the Late Oligocene and Early Miocene and the central parts of Khangai were involved in this process in the Late Miocene–Holocene, we can easily understand (Fig. 11) that the centers of

intraplate volcanism within SKVR have described a loop for the last 100 Myr. They first descended from the eastern termination of the Gobi Altai ranges southward and then gradually moved northward from the Gobi Tien Shan area to Khangai. We relate this migration to the lithospheric plate movements relative to the localized hot mantle source (mantle hot spot), which controlled intraplate volcanic activity.

Intraplate Magmatism Sources

Above, we presented the geochemical and isotope characteristics of rocks and indicated the geochemical similarity of different-age vulcanites, which indicates that the composition of magmatism sources at the stages of area formation was relatively constant. In addition to the considered rocks, the graphs (Figs. 4–9) show for comparison the characteristics of vulcanites of the Late Oligocene in the Khangai area and of the Late Cretaceous rocks corresponding to the final stage of formation of the Gobi Altai rift zone (Fig. 7), which are usually considered to be within independent volcanic regions. The Late Cretaceous rocks of the Gobi Altai rift zone and vulcanites of the Khangai area are completely similar to lavas of the Gobi Tien Shan and show that particular rock composition (trachybasalts and alkaline basaltoids) and the specific features of volcanic processes. Evidently, these rocks are also very similar in their geochemical characteristics. Insignificant differences (such as, e.g., lower contents of Ta and Nd in the Late Cretaceous vulcanites of the Gobi Altai rift zone or variations in the U, Th, and HREEs content of rocks of different age groups) are quite explainable from the standpoint that the composition of the intraplate magmatism sources insignificantly changed in time.

The isotope data additionally indicate that the compositions of the magmatism mantle sources are similar. Previously [19], we considered the general regularities of variations in the isotope composition of the magmatic rocks during the SKVR development based on relatively limited data. We indicated that the composition of magmatic products depended on three final components, which were compared with a moderately depleted mantle (PREMA) and enriched EM-I and EM-II mantles (inset in Fig. 10a). We also indicated that the composition of the sources regularly changed during the development of the volcanic region. The composition of the magmatic rocks was governed by the combinations of the PREMA with EM-II and PREMA with EM-I sources at the early (Late Jurassic–Early Cretaceous) stages and beginning from the Late Cretaceous, respectively. The new material does not contradict previously revealed patterns. In Fig. 10, the variations in the isotope compositions of the Late Cretaceous (Gobi Altai rift zone), Late Cretaceous–Early Cenozoic (Gobi Tien Shan area), and Late Cenozoic (Khangai area) igneous rocks show general patterns. These compositions are localized within a linearly extended field,

which is traced from the quadrant of depleted mantle sources into the region depleted in radiogenic Nd and moderately enriched in radiogenic Sr. As was mentioned above, the compositions of rocks in this field can be considered as the result of mixing of the PREMA source and the EM-I source enriched in Nd. The general relationships between the isotope composition and content of Sr and Nd are also typical of these rocks (Figs. 10b, 10c), which indicates that the isotope and REE compositions of the final components that participated in the formation of magmatism sources are uniform.

Thus, both isotope and geochemical data indicate that the compositions of the mantle sources, which controlled intraplate magmatic activity within southern and central Mongolia for ~100 Myr, are similar in spite of the fact that the spatial position of magmatism manifestation varied in time. This source is similar to the source of OIB basalts, which is evident in the La/Yb–Th/Ta, Ce/Pb–Nb/U, and Ba/Ce–Ce/U graphs (Fig. 9), where the compositions of rocks in the area fall in the field of OIB compositions or are grouped near this field.

We should note that the mantle, with similar isotope and geochemical characteristics, did not participate in magmatism at the earlier stages of formation and transformation of the Central Asian Hercynides of the lithosphere. Above, we stated that the regional crust is juvenile, which suggests that a depleted mantle, with an $\epsilon_{\text{Nd}}(420) \sim +8$ by the time of formation of the Hercynian lithosphere (~420 Ma ago), was present in the crust basement. The regional lithosphere was subsequently transformed with the participation of mantle magmatism. At first, this transformation proceeded under the conditions of an active continental margin (~330 Ma ago) with the participation of melting products of the metasomatized mantle wedge, where the $\epsilon_{\text{Nd}}(330) \sim +5$. Later, the crust was transformed under the conditions of continental rifting, which manifested itself (1) in the Late Paleozoic and (2) in the Late Jurassic–Early Cretaceous in connection with the formation of the Central Asian and Gobi Altai rift zones, respectively [17, 21]. In both cases the cause of melts was mantle with isotope and material parameters close to such parameters of an EM-II enriched mantle, $\epsilon_{\text{Sr}}(\text{T}) > +20$, $\epsilon_{\text{Nd}}(\text{T}) \sim$ from +1 to –2, or the combination of EM-II with a moderately depleted mantle. A deficit of the Ta and Nb content was typical of the mantle melting products, which possibly indicates that the concentration of water in the melting chambers increased. The PREMA mantle became predominant in the source of mantle magmas in the Gobi Altai rift zone only at the end of the Early Cretaceous–beginning of the Late Cretaceous. The source of magmatism, which has manifested itself for the last 100 Myr and formed the Gobi Tien Shan and Khangai areas, evidently differs in composition from both the lithospheric mantle and Hercynides on the whole. This allows us to assume that this source originated because a region of melting of the mantle substratum appeared

which was never previously registered in the structure of the regional upper mantle.

Magmatism and Regional Depth Structure

The conclusion that a hot mantle high existed for a long time in the basement of the lithosphere suggests that this high is reflected in the lithospheric structure. Based on the data of the gravimetric and seismological studies, the authors of [4, 22, 31] indicated that a large region of hot mantle, within which the asthenosphere reaches depths of less than 100 km, is located in the basement of the lithosphere in Central Asia. Local highs of the asthenosphere, which reach the crust bottom at depths of about 50 km, were distinguished in the structure of this region [4]. The location of these highs is in good agreement with the regions of recent volcanism (Fig. 12). These highs are observed beneath the South Khangai Late Cenozoic volcanic area and at the bottom of the southern Baikalian, Vitim, and Udokan volcanic regions, within which volcanic activity continued for the Late Cenozoic. A similar relation suggests that precisely these highs of the hot mantle (asthenosphere) control recent volcanic activity in the region, and, consequently, they can be considered as mantle hot spots (or hot fingers by analogy with small plumes controlling the Late Cenozoic rifting in Western Europe [29]). On the other hand, the absence of older (the Early Cenozoic) volcanic occurrences above these hot spots allows us to conclude that they were formed approximately simultaneously with volcanism intensification in the region, i.e., about 20–30 Ma ago [18, 19].

The asthenospheric high, located outside the regions of recent volcanism near the termination of the Gobi Tien Shan ranges, is especially pronounced against this background. The position of this high is, nevertheless, quite natural since it is in good agreement with the territory of development of the Late Cretaceous–Early Cenozoic volcanism of the Gobi Tien Shan area. Within this territory, intraplate volcanic activity has taken place for not less than 60 Myr between 90 and 30 Ma ago. As was mentioned above, 30 Ma ago, volcanically active territories within SKVR were reconstructed, and volcanic centers shifted northward into the regions of southern slopes and central part of the Khangai upland, beneath which the Khangai asthenospheric high was formed. Therefore, we can assume that the SKVR volcanic occurrences during the last 90 Myr were related to a small plume of the mantle (hot-finger), which was stable relative to the lithosphere and injected the lithospheric mantle, forming local highs of the asthenosphere. Under such an assumption, the Gobi Tien Shan asthenospheric high can be considered as a residual trap, which remained in the structure of the lithosphere after its displacement relative to the mantle hot jet. The evidences of insignificant migrations of the Central Asian lithosphere during the Cretaceous and Cenozoic

were obtained previously based on the paleomagnetic studies [8].

Model of Magmatic Area Formation

The above data allow us to make the following conclusions.

(1) The Gobi Tien Shan volcanic area is part of the southern Khangai volcanic region and originated during a sharp weakening of volcanic activity in this region. Migration of volcanic centers over the region beginning ~100 Ma ago was related to motions of the lithospheric plate relative to the mantle source of magmatism.

(2) The composition of the magmatism sources remained more or less homogeneous over not less than 100 Myr. Insignificant changes in the composition were directed and were related to an increase in the degree of enrichment of sources in incompatible elements and to an increase in the contribution of the component with isotope characteristics of the EM-I enriched mantle.

(3) These sources were located in the Earth's interior not higher than the boundary of the upper and lower mantle and were carried to the bottom of the lithosphere by a mantle jet (mantle plume), which locally affected the bottom of the lithosphere forming here asthenospheric highs.

These boundary conditions make it possible to propose the following geodynamic model of formation of the Gobi Tien Shan volcanic area in the composition of SKVR. The Early Cretaceous history of the volcanic region was related to the formation of an extensive (to 500 km) Gobi Altai rift zone, which supposedly originated [15, 18, 19] above the mantle plume, and had enriched EM-II mantle in the composition of its sources. The effect of this plume on the lithosphere gradually weakened over time and took place only in an area less than 90×90 km in the region of the rift zone eastern margin by the beginning of the late Cretaceous. A decrease in volcanic productivity was accompanied by not only a replacement of sources, the composition of which from ~100 Ma ago became dependent on a moderately depleted PREMA mantle and on an enriched mantle with EM-I characteristics, but also by a clearly defined migration of volcanic centers over the volcanic region. We relate this migration to displacements of the lithosphere relative to the mantle plume. By ~90 Ma ago, the volcanic centers shifted from the Gobi Altai southward to the eastern termination of the Gobi Tien Shan ranges. From 90 to 30 Ma ago, active volcanism took place in a small area beneath which the Gobi Tien Shan asthenosphere trap is now recognized (Fig. 12). Within this area, Late Cenozoic manifestations of volcanism are absent; therefore, we can assume that this trap originated as a result of the prolonged (~60 Myr) Late Cretaceous–Early Cenozoic impact of the hot spot on the basement of a stiff lithosphere. From the second half of the Oligocene, the northward migra-

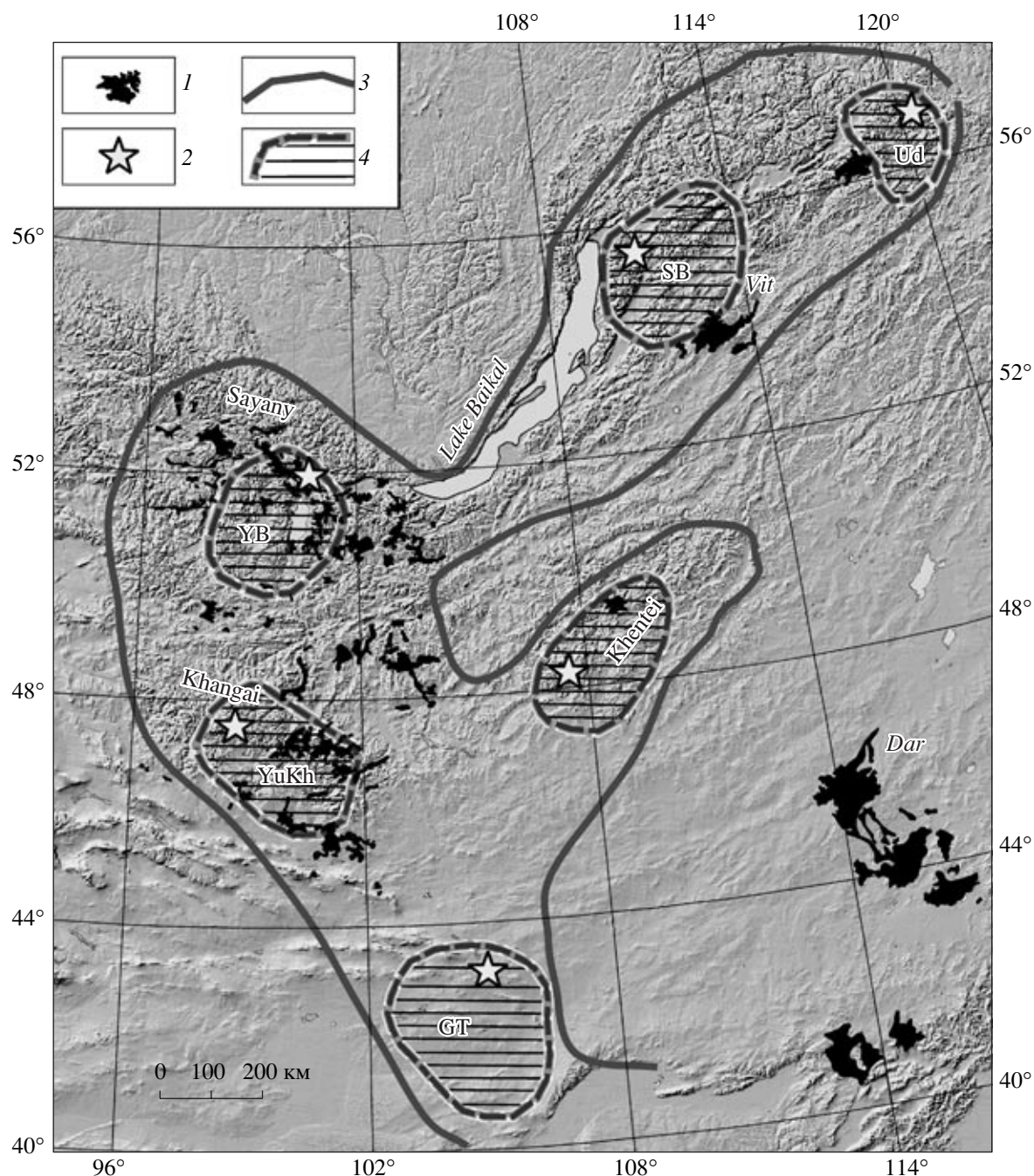


Fig. 12. Schematic distribution of the recent volcanism regions and mountain systems relative to the asthenospheric highs related to the mantle hot field in southern Eastern Siberia and central Mongolia [31]: (1) lava fields, (2) the highest summits of mountain systems, (3) region of asthenospheric rise to depths less than 100 km, (4) asthenospheric highs (mantle plumes) to depths less than 50 km. Letters mark highs of the asthenosphere (mantle hot spots): (YuKh) South Khangai, (GTA) Gobi Tien Shan, (YB) southern Baikalian, (SB) northern Baikalian, (UD) Udokan. Italic letters denote lava plateaus: (Vit) Vitim and (Dar) Dariganga.

tion of volcanism areas became evident, which possibly indicates that the motions of the lithosphere relative to the hot spot were more pronounced. From the middle of the Miocene, the region of volcanism became stable within the Khangai upland, beneath which the asthenosphere ascended to the bottom of the regional crust under the action of the mantle plume.

CONCLUSIONS

(1) The Gobi Tien Shan volcanic area is part of SKVR. This area originated after the termination of magmatic activity within the Gobi Altai rift zone and immediately before the formation of the Late Cenozoic Khangai Volcanic area. This makes it possible to trace

the history of SKVR development beginning from the Late Jurassic to the Holocene inclusive, i.e., in the interval of about 150 Myr.

(2) The Gobi Tien Shan volcanic area was formed in three stages of activity (the Late Cretaceous, Paleocene–Early Eocene, and Early Oligocene) against a background of a sharp decrease in volcanic productivity. This activity resulted in thin lava plateaus and associated different subvolcanic occurrences: stocks, dikes, laccolites, and domes. Volcanic activity was accompanied by the formation of small volcanic grabens and normal fault systems.

(3) The composition of magmatic products of the area is represented by basic products of the alkaline and subalkaline petrochemical series: melanephelinites, basanites, trachybasalts, and trachyandesitobasalts. These rocks are similar to OID basalts in the geochemical characteristics. The isotope characteristics of magmatic sources correspond to the combination of a moderately depleted PREMA mantle and enriched EM-I mantle. The isotope parameters of the latter indicate that this mantle was isolated in the Earth's interior before the Phanerozoic but did not participate in the formation and intraplate transformation of the regional crust up to 100 Ma ago and, consequently, in the processes proceeding in the upper mantle.

(4) The formation of the volcanic area, as well as of the entire SKVR, was related to the mantle plume (South Khangai mantle hot spot), which supplied matter from the lower mantle to the bottom of the lithosphere. The intensity of interaction between the mantle jet and the lithosphere substantially changed in time. Specifically, this intensity was sharply decreased during the formation of the area. Nevertheless, the hot spot trace in the processes of intraplate intensification is expressed rather distinctly. The trace was registered not only in the composition of magmatic products corresponding to the deep-seated mantle sources but also in the localized manifestation of volcanic activity within the territory that was in the state of platform development. This localization, expressed through a regular migration of volcanic centers within a limited area on the surface and through an asthenospheric high at the bottom of the region, was evidently caused by a point impact of the hot mantle on the lithosphere, which insignificantly shifted relative to this mantle.

ACKNOWLEDGMENTS

This work was supported by the Russian Foundation for Basic Research (project nos. 05-05-64000 and 05-05-64001); program no. 10, Department of Earth Sciences, Russian Academy of Sciences, and program no. 16 of the Presidium of the Russian Academy of Sciences.

REFERENCES

1. Afonin, V.P., Gunicheva, T.N., and Piskunova, L.F., *Rentgenofluorescentnyi silikatnyi analiz* (X-ray Fluorescence Silicate Analysis), Novosibirsk: Nauka, 1984.
2. Devyatkin, E.V., *Kainozoi Mongolii* (The Cenozoic of Mongolia), Moscow: Nauka, 1980.
3. Zonenshain, L.P., Kuz'min, M.I., and Natapov, L.M., *Tektonika litosfernykh plit territorii SSSR* (Tectonics of Lithospheric Plates in the USSR), Moscow: Nedra, 1990.
4. Zorin, Yu.A., Balk, E.V., Novoselova, M.R., and Turutanov, E.Kh., Lithospheric Thickness below the Mongolian–Siberian Highland and Adjacent Regions, *Fiz. Zemli*, 1988, no. 7, pp. 32–42.
5. Kovalenko, V.I., Yarmolyuk, V.V., Kovach, V.P., et al., Magmatism as Factor of Crust Evolution in the Central Asian Foldbelt: Sm–Nd Isotopic Data, *Geotektonika*, 1999, vol. 33, no. 3, pp. 21–41 [*Geotectonics* (Engl. Transl.), 1999, vol. 33, no. 3, pp. 191–208].
6. Kovalenko, V.I., Yarmolyuk, V.V., Kovach, V.P., et al., Sources of Phanerozoic Granitoids in Central Asia: Sm–Nd Isotope Data, *Geokhimiya*, 1996, no. 8, pp. 14–30 [*Geochem. Int.* (Engl. Transl.), 1996, no. 8].
7. Kovalenko, V.I., Yarmolyuk, V.V., Sal'nikova, E.B., et al., Sources of Igneous Rocks and Genesis of the Early Mesozoic Tectonomagmatic Area of the Mongolia–Transbaikalia Magmatic Region: 2. Petrology and Geochemistry, *Petrologiya*, 2003, vol. 11, no. 3, pp. 227–254 [*Petrology* (Engl. Transl.), 2003, vol. 11, no. 3, pp. 205–229].
8. Kovalenko, D.V., Yarmolyuk, V.V., and Solov'ev, A.V., Migration of Volcanic Centers at the South Khangai Hot Spot: Paleomagnetic Evidence, *Geotektonika*, 1997, vol. 31, no. 3, pp. 66–73 [*Geotectonics* (Engl. Transl.), 1997, vol. 31, no. 3, pp. 228–235].
9. *Kontinental'nyi vulkanizm Mongolii* (Continental Volcanism of Mongolia), Moscow: Nauka, 1983, p. 153.
10. *Mezozoiskaya, kainozoiskaya tektonika i magmatizm Mongolii* (Mesozoic and Cenozoic Tectonics and Magmatism of Mongolia), Moscow: Nauka, 1975, p. 305.
11. Savatkov, V.M., Morozova, I.M., and Levsky, L.K., Behavior of the Sm–Nd, Rb–Sr, K–Ar, and U–Rb Isotopic Systems during Alkaline Metasomatism: Fenites in the Outer-Contact Zone of an Ultramaphic–Alkaline Intrusion, *Geokhimiya*, 2004, no. 10, pp. 1027–1049 [*Geochem. Int.* (Engl. Transl.), 2004, no. 10, pp. 899–920].
12. Samoilov, V.S. and Kovalenko, V.I., *Kompleksy shchelochnykh porod i karbonatitov Mongolii* (Complexes of Alkaline Rocks and Carbonates of Mongolia), Moscow: Nedra, 1983.
13. Frikh-Khar, D.I. and Luchitskaya, A.I., *Pozdnemezozoiskie vulkanity i svyazannye s nimi gipabissal'nye intruzivny Mongolii* (Late Mesozoic Vulcanites and Related Hypabyssal Intrusions of Mongolia), Moscow: Nauka, 1978.
14. Shuvalov, V.F. and Nikolaeva, T.V., On the Age and Spatial Distribution of Cenozoic Basalts in Southern Mongolia, *Vestn. Leningr. Univ.*, 1985, no. 14, pp. 52–59.
15. Yarmolyuk, V.V., Ivanov, V.G., and Kovalenko, V.I., Dynamics of Formation and Magmatism of the Late Mesozoic–Cenozoic Southern Khangai Mantle Hot Spot (Mongolia), *Geotektonika*, 1994, no. 5, pp. 28–45.

16. Yarmolyuk, V.V., Ivanov, V.G., and Kovalenko, V.I., Sources of Intraplate Magmatism of Western Transbaikalia in the Late Mesozoic–Cenozoic: Trace Element and Isotope Data), *Petrologiya*, 1998, vol. 6, no. 2, pp. 115–139 [*Petrology* (Engl. Transl.), 1998, vol. 6, no. 2, pp. 101–123].
17. Yarmolyuk, V.V. and Kovalenko, V.I., Deep Geodynamics and Mantle Plumes: Their Role in the Formation of the Central Asian Fold Belt, *Petrologiya*, 2003, vol. 11, no. 6, pp. 556–586 [*Petrology* (Engl. Transl.), 2003, vol. 11, no. 6, pp. 504–531].
18. Yarmolyuk, V.V., Kovalenko, V.I., and Ivanov, V.G., Intraplate Late Mesozoic–Cenozoic Volcanic Province of the Central–Eastern Asia Is Projection of the Mantle Hot Spot, *Geotektonika*, 1995, no. 5, pp. 41–67.
19. Yarmolyuk, V.V., Kovalenko, V.I., Ivanov, V.G., and Zhuravlev, D.Z., Isotopic Composition of Sr and Nd Basic Vulcanites of the Southern Khangai Hot Spot in Central Asia, *Dokl. Akad. Nauk*, 1995, vol. 342, no. 2, pp. 230–234.
20. Barry, T.L., Saunders, A.D., Kempton, P.D., et al., Petrogenesis of Cenozoic Basalts from Mongolia: Evidence for the Role of Asthenospheric versus Metasomatized Lithospheric Mantle Sources, *J. Petrol.*, 2003, vol. 44, pp. 55–91.
21. Dergunov, A.B., Kovalenko, V.I., Ruzhentsev, S.V., and Yarmolyuk, V.V., *Tectonics, Magmatism, and Metallogeny of Mongolia*, London: Routledge. Taylor Francis Group, 2001.
22. Gao, S.S., Liu, K.H., Davis, P.M., et al., Evidence for Small-Scale Mantle Convection in the Upper Mantle Beneath the Baikal Rift Zone, *J. Geophys. Res.*, 2003, vol. 108B, no. 4, pp. 1–12.
23. Halliday, A.N., Lee, D.-Ch., Tommasini, S., et al., Incompatible Trace Elements in OIB and MORB and Source Enrichment in Sub-Oceanic Mantle, *EPSL*, 1995, vol. 133, pp. 379–395.
24. Kovalenko, V.I., Yarmolyuk, V.V., Kovach, V.P., et al., Isotope Provinces, Mechanism of Generation and Sources of the Continental Crust in the Central Asian Mobile Belt: Geological and Isotopic Evidence, *J. Asian Earth Sci.*, 2004, vol. 23, pp. 605–627.
25. Macdonald, R., Rogers, N.W., Fitton, J.G., et al., Plume–Lithosphere Interactions in the Generation of the Basalts of the Kenya Rift, East Africa, *J. Petrol.*, 2001, vol. 42, no. 3, pp. 877–900.
26. Rogers, N., Macdonald, R., Fitton, J.G., et al., Two Mantle Plumes beneath the East African Rift System: Sr, Nd and Pb Isotope Evidence from Kenya Rift Basalts, *EPSL*, 2000, vol. 176, pp. 387–400.
27. Sun, S.S. and Donough, W.F., Chemical and Isotopic Systematizes of Oceanic Basalts: Implications for Mantle Composition and Processes, Magmatism in the Ocean Basins, Saunders, A.D. and Norry, M.J., Eds., *Geol. Soc. London, Spec. Publ.*, 1989, vol. 42, pp. 313–346.
28. Tomlinson, K.Y. and Condie, K.C., Archean Mantle Plumes: Evidence from Greenstone Belt Geochemistry, in *Mantle Plumes: Their Identification through Time. Spec. Paper*, 2001, pp. 341–358.
29. Wilson, M. and Patterson, R., Intraplate Magmatism Related to Short-Wavelength Convective Instabilities in the Upper Mantle: Evidence from the Tertiary–Quaternary Volcanic Province of Western and Central Europe, *Geol. Soc. Am. Spec. Paper*, 2001, vol. 352, pp. 37–58.
30. Zindler, A., Hart, Ä. Chemical Geodynamics, *Ann. Rev. Earth Planet. Sci.*, 1986, vol. 14, pp. 493–571.
31. Zorin, Yu.A., Turutanov, E.Kh., Mordvinova, V.V., et al., The Baikal Rift Zone: the Effect of Mantle Plumes on Older Structure, *Tectonophysics*, 2003, vol. 371, pp. 153–173.