

# **DEEP SEATED MAGMATISM**, Its sources and plumes

Глубинный магматизм, его источники и плюмы



MIRNY IRKUTSK 2006 Russian Academy of Sciences Siberrian Branch Vinogradov Institute of Geochemistry Central Science and Research Geology And Prospecting Institute Of The Stock Company "ALROSA" Russian Foundation of Basic Research





# Deep-seated magmatism, its sources and plumes

(Глубинный магматизм, его источники и плюмы)

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#### Deep-seated magmatism, its sources and plumes

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The given book presents invited reports of the VI-th International Workshop « Deep-seated magmatism, its sources and plumes », 2006, Mirny.

The invited reports discuss the features of alkaline, carbonatite and kimberlite magmatism. These papers view the features of alkaline, carbonatite and kimberlite magmatism. They also disclose the influence of the fluid on the thermal regime of mantle plumes. The present issue discusses mass-balance of global components of the silicate Earth, Enriched Mantle Reservoirs as the source of alkaline magmatism. The papers included into this issue also give new data on geology, geodynamics and age of the largest and the most rare-metal massif of alkaline granites Khan-Bogdo, Mongolia. and consider a number of aspects of origin and ore potential of kimberlites. Geochemistry of basalts and komatiites and their association with plume processes are discussed in papers by I.Yu. Safonova and Ya. B. Bychkova with co-authors while the problems of alkaline magmatism are considered. All works assume the plume mechanism of fusion of primary alkaline magmas in some areas of the Earth and in different age periods.

The book is of great importance for petrologists, geochemists, and specialists studying deep alkaline and kimberlite magmatism, students and teaching staff of universities.

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The book cover depicts the sample of layered benstonite cabonatite from the Murun massif. The yellow color designates benstonite, grey is quartz, black is pyroxene, brown is tinaxite and green shows microcline.

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#### FOREWORD

The fact is well accepted that alkaline rocks represent unique formations on the Earth. They have been long attractive for research because large Nb, Ta, Zr, Y, TR, Cu and P deposits, gemstones of charoite, Cr-diopside, dianite are associated with them. For instance, in Australia diamonds are recovered in lamproites. The complicated processes of their formation provoked scientific disputes still going on. The newly developed analytical methods and techniques provided abundant information on the composition of alkaline rocks. The data on geochemistry of isotopes confirm the evidence on the mantle sources of the substance of alkaline rocks. The new concepts of plume tectonics are applied by scientists when studying alkaline rocks as the deep-seated geodynamics of the Earth is interpreted based on these data.

These problems were discussed at the international workshops held in 2001 at the Institute of Geochemistry in Irkutsk; in 2002 at the Far-East Geological Institute, Vladivostok; in 2003 at the Institute of Tectonics and Geophysics in Khabarovsk, in 2004 at Geological Institute in Ulan-Ude, in 2005 at the Institute of Volcanology and Seismology in Petropavlovsk-Kamchatsky.

This book offers invited papers of the international workshop organized in Mirny city under the title **"Deep-seated magmatism, its sources and plumes**".

These papers view the features of alkaline, carbonatite and kimberlite magmatism. They also disclose the influence of the fluid on the thermal regime of mantle plumes. The present issue discusses mass-balance of global components of the silicate Earth (Academician I.D. Ryabchikov), Enriched Mantle Reservoirs as the source of alkaline magmatism (Academician L. N. Kogarko). The papers included into this issue also give new data on geology, geodynamics and age of the largest and the most rare-metal massif of alkaline granites, Khan-Bogdo, Mongolia, (Academician V. I. Kovalenko et al.,) and consider a number of aspects of origin and ore potential of kimberlites (articles by Profs. Z.V. Spetsius, I.V., Aschepkova, S.I. Sablukova, L.V. Solov'eva et al.,). Geochemistry of basalts and komatiites and their association with plume processes are discussed in papers by I.Yu. Safonova and Ya. B. Bychkova with co-authors while the problems of alkaline magmatism are considered in papers by L.N. Kogarko, G.S. Ripp, etc.

All works assume the plume mechanism of fusion of primary alkaline magmas in some areas of the Earth and in different age periods.

The book might present interest to specialists involved in petrological and geochemical investigations as well as those studying deep alkaline and kimberlite magmatism.

Chairman of Organizing Committee, Chief Editor

Dr. N.V. Vladykin

# UDK 550.42 +552 Mass-Balance of Global Components in the Silicate Earth

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Mass balance of incompatible elements was analyzed for different inferred reservoirs of the bulk silicate Earth. It was shown that the lower mantle, as well as the primitive mantle, includes an MORB-source depleted mantle component and material with a high content of incompatible elements. Contribution of the continental crust was found to be subordinate. The predominant part of the enriched mantle formed through differentiation of the mantle itself. Enriched material was supplied into the deep-seated zones of silicate shell through delamination of subcontinental lithospheric blocks affected by mantle metasomatism and their subsequent involvement in the mantle convective system. The osmium isotope composition of the plunged lithospheric material is modified in the lower mantle by the infiltration of sulfide melts.

# **INTRODUCTION**

The chemical composition of the Earth is a cosmo-chemical and geochemical problem. The concept of the chondritic Earth was formulated by Vinogradov [1]. The pyrolite model of Ringwood used this concept in considering petrogenic components, suggesting that the primitive mantle is a mixture of depleted mantle peridotites and supposedly complementary magmas in proportions giving the chondritic Ca/Al ratio [2]. The chondritic model of the silicate Earth was extended later to trace elements. This approach is based on the fact that elements produced by high-temperature nebular condensation (for which calculated temperatures of 50% gas-to-solid transition are higher than 1400 K) occur in different types of chondrites approximately in equal proportions. Since the Earth was accreted from the same material as the parental bodies of chondritic meteorites, the bulk silicate Earth should show almost chondritic ratios of lithophile elements of high-temperature condensation. This principle, together with geochemical data on mantle rocks and magmas, was used to estimate the bulk composition of the Earth's primitive mantle [3, 4].

Some researchers accept more extreme versions of the chondritic model, assuming that the Earth is composition-ally identical to a certain chondrite type [5, 6]. Covariations of indicator elements in meteorites and the Earth led them to conclude that the Earth's composition corresponds to carbonaceous chondrites, in particular, to their hypothetical type poor in volatile contents [5, 6].

The chondritic proportions of components in the Earth are not exactly proved yet. For example, they could have been shifted at early stages by the collision of a partially differentiated Earth with a large cosmic body and the consequent loss of a part of the outer shell [4]. The chondritic model is tested by the comparison of deduced conclusions with data on the abundances of non-volatile lithophile elements and their isotopes in mantle rocks and magmas. The most important criterion is provided by the Sm-Nd and Lu-Hf isotope systems, since all these elements belong to the lithophile elements of high-temperature condensation. Recent isotope data on different chondrites show a close positive correlation between <sup>143</sup>Nd/<sup>144</sup>Nd and <sup>176</sup>Hf/<sup>177</sup>Hf in the mantle rocks. At the same time, <sup>176</sup>Lu/<sup>177</sup>Hf and <sup>176</sup>Hf/<sup>177</sup>Hf ratios are somewhat more widely scattered than is seen in the Sm-Nd system. The average <sup>176</sup>Hf/<sup>177</sup>Hf value for carbonaceous chondrites lies within the mantle correlation trend in the Hf-Nd isotope diagram [7]. In general, data on these isotope systems and many elemental ratios in the mantle rocks indicate that the chondritic model is a good approximation for the bulk Earth composition.

Composition of the bulk silicate Earth with chondritic ratios of nonvolatile lithophile elements was estimated in [3, 4]. The global differentiation of the Earth produced geochemical reservoirs depleted (depleted mantle as a MORB source) and enriched (for example, continental crust) in incompatible elements. As a result, most mantle rocks and magmas lost the geochemical characteristics of the chondrite Earth.

#### THREE-COMPONENT MODEL OF THE SILICATE EARTH

The composition of the depleted mantle was estimated by modeling of the generation of primary magmas and their differentiation, which produced typical mid-ocean ridge basalts [8]. The average composition of the continental crust was obtained from data on various sedimentary and metamorphic rocks and lower crustal xenoliths in alkali basalts and kimberlites [9].



Fig. 1. Primitive pyrolite-normalized [4] calculated contents of incompatible elements in a mixture of continental crust and depleted mantle with weight proportions of the continental crust and mantle above 660-km seismic boundary.

The simplest model of the differentiated Earth suggests that continental crust and depleted mantle compose the outer shells of the Earth to the boundary with the lower mantle, while the latter is identical to the undifferentiated bulk silicate Earth. However, mixing between continental crust (according to [4]) and depleted mantle (according to [8]) yields a composition significantly different from the primitive mantle in the chondritic model [4] (Fig. 1). The two components were mixed in weight proportions of continental crust and upper mantle. Thus, this simplest model failed the quantitative test.



Fig. 2. Values  $X_c = X_{cc} / (X_{cc} + X_{dm})$ , where  $X_{cc}$  and  $X_{dm}$  are fractions of continental crust and depleted mantle in the silicate Earth, as calculated for incompatible elements.

In the considered three-component model of the silicate Earth (continental crust + depleted mantle + primitive mantle), the mixing proportions of continental crust and depleted mantle can be changed to obtain primitive mantle with chondritic ratios of nonvolatile lithophile elements. This corresponds to the case when the depleted mantle and continental crustal components are present not only in the outer part of the Earth, but also in the lower mantle, where they can be mixed with primitive mantle. This is a more realistic model, which suggests a mixture of depleted and primitive mantles with isotope characteristics of most intraplate oceanic island basalts, whose source is depleted in incompatible elements (PREMA [10], FOZO [11], or PHEM [12]), but to a lesser extent than the MORB source. This model implies that:

$$(1-X_c) \cdot C_{dm} + X_c \cdot C_{cc} = C_{pm}$$

where  $X_c = X_c c/(X_{cc} + X_{dm})$ ,  $X_{cc}$  and  $X_{dm}$  are respective fractions of continental crust and depleted mantle in the silicate Earth,  $C_{cc}$  and  $C_{dm}$  are respective concentrations of an element in these reservoirs, and  $C_{pm}$  is the concentration of the element in the primitive mantle. Hence,  $X_c = (C_{pm} - C_{dm})/(C_{cc} - C_{dm})$ . This formula was applied to calculate  $X_c$  on the basis of the contents of different incompatible elements in the continental crust and depleted and primitive mantles. Results of these calculations are shown in Fig. 2. The optimal  $X_c$  for a group of elements was calculated by the least square method as 0.021. Figure 2 shows that the  $X_c$  values for some elements significantly differ from this optimal value, which may indicate either imperfection of the applied model or uncertainty in the estimated average contents of these elements in components of the silicate Earth.

Figure 3b demonstrates a primitive mantle-normalized multielement diagram for a mixture of depleted mantle (97.9%) and continental crust (2.1%) with the optimal mixing proportions obtained by the least square method. There are significant deviations for the elements which behave differently in processes of basaltic mantle magmatism (magmatism of mid-ocean ridges and intraplate oceanic islands) and formation of the continental crust (average composition of the continental crust and calc-alkaline magmas). These are Nb and Ta, whose abundances in the continental crust are lower than those of their analogues in the basaltic systems (e.g., U for Nb), as well as fluid-mobile components (e.g., Pb, Sn, and others), which are abundant in the continental crust and calc-alkaline magmas [13]. Thus, the considered model (Fig. 3b) demonstrates a negative deviation for Nb and a positive deviation for Pb and Sn, from the accepted composition of the primitive mantle. Thus, calculated optimal proportions of the depleted mantle and continental crust show excess crustal geochemical signatures. Hence, the material enriched in incompatible elements without distinct negative anomalies of Nb and Ta and positive anomalies of fluid-mobile components is required, in addition to the continental crust, to solve the problem of global geochemical reservoirs.

# EARLY BASALTIC CRUST

Concepts of three-component composition of the silicate Earth (continental crust + depleted mantle + primitive mantle) are also not consistent with U and Th geochemistry. The Th/U ratio is practically the same in the primitive mantle (3.83) and the continental crust (3.85), and almost two times lower (2.1) in the depleted mantle (MORB source). At such Th/U ratios, the primitive mantle cannot be represented as a mixture of continental crust and depleted mantle. One more component with higher Th/U ratio as compared to that in the primitive mantle is required.

The oceanic crust, which formed at the earliest stages of the Earth's evolution, can be inferred as a possible enriched reservoir, in addition to the continental crust. The early basaltic crust enriched in incompatible elements was proposed to be at least temporarily isolated from the upper mantle to explain rapid changes in Nd and Hf isotope ratios in the depleted mantle. In the upper mantle depleted in Nd and Hf relative to Sm and Lu, the <sup>143</sup>Nd/<sup>144</sup>Nd and <sup>176</sup>Hf/<sup>177</sup>Hf ratios significantly increased during the first several million years of the Earth's history [14]. It was also proposed that the early basaltic crust might be a significant

component of the lowermost lower mantle (below 1600 km), which could be partially isolated from the overlaying convective system of the silicate Earth [15].

Table 1.

Element	PM	DM	CC	FMORB	EM
Rb	0.605	0.055	32	13.9	1.06
Ва	6.75	0.61	250	154	11.8
Th	0.083	0.008	3.5	1.9	0.11
W	0.016	0.002	1	0.29	0.021
U	0.022	0.004	0.91	0.4	0.027
Nb	0.6	0.19	11	10.7	1.06
Κ	260	49	9100	3492	364
La	0.686	0.29	16	8.6	1.03
Ce	1.785	0.96	33	22.4	2.2
Pb	0.185	0.039	8	2.3	0.25
Pr	0.27	0.17	3.9	3.2	0.3
Nd	1.327	0.93	16	9.4	1.8
Sr	20.3	16.05	260	217.6	24.4
Sm	0.431	0.35	3.5	5.3	0.5
Sn	0.138	0.113	2.5	1.8	0.17
Hf	0.3	0.23	3	3.8	0.37
Zr	10.8	8.14	100	89.5	13.2
Ti	1282	1168	5400	12225	1441
Eu	0.162	0.14	1.1	1.6	0.18
Gd	0.571	0.5	3.3	5.5	0.59
Tb	0.105	0.092	0.6	1.03	0.1
Dy	0.711	0.64	3.7	6.8	0.7
Ho	0.159	0.14	0.78	1.4	0.16
Y	4.37	4.06	20	43.1	4.4
Er	0.465	0.42	2.2	4.4	0.47
Tm	0.072	0.065	0.32	0.67	0.07
Yb	0.462	0.42	2.2	4.3	0.46
Lu	0.072	0.064	0.3	0.66	0.072

Contents of incompatible elements (in ppm) in different hypothetical reservoirs of the silicate Earth

**Note:** (PM) primitive mantle [4]; (DM) depleted mantle [8]; (CC) continental crust [9]; (FMORB) early basaltic crust (this paper); (EM) enriched peridotite mantle (this paper).

To estimate the composition of the early basaltic crust, the ancient magma generation was assumed to be similar to the magma generation in the modern midocean ridges; however, the source of the magmas was primitive pyrolite material rather than the depleted mantle of present-day composition. Consistently with this assumption, the average contents of incompatible elements in the N-MORB were multiplied by the ratios of their contents in the primitive mantle [4] to those in the depleted upper mantle [8] (table, Fig. 3a).

Assuming similar formation conditions for the early and modern basaltic crust, we accepted that the temperature of the convective upper mantle was mainly defined by parameters of the transition of peridotite material from rigid to plastic state and practically did not vary with time. Significantly higher global heat flow in the first several million years after accretion owing to the significantly higher

contents of heat-generating elements led to the lower average thickness of the lithosphere and significant increase of thermal convection. As a result, magma production was significantly more intense than in the present day. However, the thin lithosphere universally defined low partial melting pressures, which were similar to those in the shallow magma-generating systems beneath modern mid-ocean ridges.



Fig. 3. (a) PM-normalized [4] contents of incompatible elements in different hypothetical reservoirs of the silicate Earth. (CC) continental crust [9]; (FMORB) early basaltic crust; (DM) depleted mantle [8], (EM) enriched peridotite mantle; (DM + CC + FMORB + EM) is a mixture of depleted mantle (68.58%), continental crust (1.05%), early basaltic crust (0.19%), and enriched peridotite mantle (30.18%). (b) Simulation of incompatible element contents in primitive mantle by estimate of optimal proportions in a mixture of different depleted and enriched materials. A criterion of minimization of squared deviations for calculated and initial contents was used. The values were normalized to primitive mantle [4]. (DM + CC) mixture of depleted mantle (97.9%) and continental crust (2.1%); (DM + CC + FMORB) mixture of depleted mantle (97.46%), continental crust (1.60%), and early basaltic crust (0.94%); (DM + CC + FMORB + EM) is a mixture of depleted mantle (30.18%).

The least square estimate of the optimal proportions of depleted mantle, continental crust, and hypothetical early basaltic crust (FMORB) on the basis of incompatible elements yields 97.46, 1.60, and 0.94%, respectively. This three-component mixture is compared in Fig. 3b with generally accepted composition of primitive mantle. This model is somewhat better than the previous one; however, it still shows significant negative Nb and Ta anomalies, and positive, Pb and Sn. This indicates overestimated amount of the continental crust in the latter model and requires the involvement of an additional enriched geochemical reservoir.

# **ENRICHED PERIDOTITE MANTLE**

Estimates of EM sources of intraplate oceanic islands were used as an enriched geochemical reservoir [10]. Data on the chemical composition of lavas from Tristan da Cunha Island were taken for calculations.

Suggesting that the considered magmas were derived by the partial melting of mantle lherzolites, we calculated the parameters of this process taking into consideration that the contents of moderately incompatible elements, including HREE, in different early peridotites practically depend neither on the addition of a minor intergranular melt (mantle metasomatism) nor on the previous removal of such a melt (generation of depleted MORB-source mantle from primitive mantle). These contents of moderately incompatible elements are close to those in the primitive mantle and significantly differ from contents of strongly incompatible elements (LREE, U, Th, and others).

Assuming that HREE contents in the oceanic basalt source were similar to those in the primitive mantle prior to melting, we can subsequently calculate the parameters controlling the generation of primary magmas and the contents of strongly incompatible elements, including LREE, in the magma source. To solve this problem, we used the above calculation procedure suggesting the magma formation by the model of aggregate fractional melting.

Fractional melting caused by adiabatic decompression of the ascending mantle primarily depends on the initial mineral composition of source and the degree of melting at the end of the process. Depending on the depth of protolith solidus, the melting material initially contains certain amounts of garnet.

Garnet is generated from enstatite, diopside, and tschermakite components of pyroxenes by the following scheme:

$$(Mg,Fe)_2Si_2O_6 + (Mg,Fe)Al_2SiO_6 = (Mg,Fe)_3Al_2Si_3O_{12}$$

The simulation of the phase composition of primitive pyrolite with a PARMEL program [8] shows that garnet-free two-pyroxene pyrolite under nearliquidus conditions (at pressure of about 2.6 GPa) can be represented as a mixture of respective 0.26, 0.14, and 0.6 weight fractions of clinopyroxene, orthopyroxene, and olivine, the amount of which decreases after garnet appearance (weight fraction  $F_{Grt}$ ) according to the following equations:

$$\begin{split} F_{Cpx} &= 0.26 - 0.2 \cdot F_{Grt} \\ F_{Opx} &= 0.14 - 0.3333 \cdot F_{Grt} \\ F_{Ol} &= 0.6 - 0.4667 \cdot F_{Grt}, \end{split}$$

which define the initial mineralogic composition of pyrolite prior to melting.

Discretization of continuous fractional melting was attained by segmentation of the total (aggregate) fraction of generated melt into small intervals (taken as 0.001 of total weight of the system).

Within each interval, the proportions of crystalline phases vary owing to different extent of resorption (or crystallization) during melting, as well as pressure decrease during adiabatic decompression decreasing mainly the garnet fraction. Modeling of pyrolite melting during adiabatic decompression at 2.8 GPa with a PARMEL program shows that the appearance of a unit weight of the melt decreases garnet and clinopyroxene contents by respective factors of 0.97 and 1.08 and increases olivine and ortho-pyroxene contents by respective factors of 0.63 and 0.42. These proportions taken as constant for the considered process were used to calculate the mineralogic composition of pyrolite at all stages of fractional melting. The mineral composition is required to estimate combined (bulk) distribution coefficients. Distribution coefficients for individual minerals were taken from [16-18]:

$$C_{n}^{S} = \frac{\Delta F_{N-1}^{L} \cdot (1 - F_{n}^{L} + \Delta F^{L})}{\Delta F^{L}/K + 1 - F_{n}^{L}}$$

where  $(f_n$  is the content of trace element in the crystalline residue at the end of an interval of fractional melting,  $C_{n\_l}$  is the content of trace element in a crystalline residue at the end of the previous interval of fractional melting,  $F_n$  is the weight fraction of the melt from the total weight of the system for the interval of fractional melting, and Kis the combined (weighted average) distribution coefficient.

After completion of the entire cycle of fractional melting ( $F_nL = F^L$ , where  $F^L$  is the melt fraction after completion of partial melting), the content ( $C^L$ ) of a trace element in an aggregate melt (mixtures of all melts produced during individual intervals with  $F_n$ 

increment equal to  $AF^{L}$ ) was calculated from  $C^{S}$  using the following mass balance formula:

$$C^{L} = \frac{C^{0} - C^{S} \cdot (1 - F^{L})}{F^{L}}$$

where  $C^{\circ}$  is the element content in the source prior to the partial melting.

At given concentrations of trace elements in source and primary magma, deviations of the calculated  $C^L$  from the observed contents in magmatic rocks (with correction for olivine fractionation) were minimized by SIMPLEX algorithm

[19] with variations of  $F^L$  and  $F^{Grt}$ , thus providing optimal estimates of these parameters.

The estimated  $F^L$  and  $F^{Grt}$  were then used together with distribution coefficients to estimate the contents of strongly incompatible elements in an aggregate melt, assuming that their contents in the source were identical to those in pyrolite. Deviations of thus calculated contents from those observed in primary magma (in magmatic rocks with correction for olivine fractionation) could be explained by differences between real source and model pyrolite. Therefore, the contents of these elements in the initial protolith were calculated from the formula  $C^{PR} = Cpy^r C^{obs}/C^{calc}$ , where  $C^{PR}$  is the element content in protolith,  $C^{pyr}$  is its content in pyrolite,  $C^{obs}$  is the observed element content in the magmatic rock (with correction for olivine fractionation), and  $C^{calc}$  is the calculated element content in the aggregate melt assuming pyrolite composition of the source.

Calculations of the peridotite source for volcanic rocks from Tristan da Cunha Island using the described procedure yield a composition significantly enriched in strongly incompatible elements as compared to primitive pyrolite:  $(La)_N = 1.5$  (table and Fig. 3a). Least-square calculations showed that the primitive mantle can be represented as a mixture of 1.05% continental crust, 0.19% early basaltic crust, 68.58% depleted mantle, and 30.18% model enriched mantle.

The comparison of the calculated mixtures with modeled composition of the primitive mantle (Fig. 3b) shows that the addition of enriched mantle as a hypothetical reservoir, accompanied by a decrease in amount of the continental crust and hypothetical early basaltic crust, significantly improved the model: the Nb and Pb anomalies almost disappeared, while the positive Sn anomaly somewhat decreased.

The mass balance calculations for the chondritic model of the primitive mantle raise two questions: (1) which are the proportions of the proposed compo nents, in addition to primitive pyrolite, in the lower mantle and (2) which mechanism is responsible the for generation of enriched mantle?

Assuming that the material of the continental crust in the silicate Earth accounts for 0.65%, i.e., is equal to the fraction of the continental crust, the lower mantle should consist of 50.99% primitive mantle, 23.14% depleted mantle, and 24.85% enriched mantle. In this hypothetical case, the continental crust is absent in the lower mantle, while the proportion of primitive material is the highest. Early basaltic crust is estimated to account for only 0.15%.

On the other hand, assuming that the lower mantle does not contain primitive pyrolite, it should consist of 58.70% depleted mantle, 40.64% enriched mantle, 0.41% continental crust, and 0.25% hypothetical early basaltic crust. In this case, the contribution of continental crust to the lower mantle composition is the highest.

Actually, the lower mantle composition can be intermediate between the considered extreme cases. Thus, according to our estimate, the lower mantle consists of 0-51% primitive mantle, 23-59% depleted mantle, 25-41% enriched

mantle, 0-0.4% continental crust, and 0.15-0.25% early basaltic crust. As mentioned above, these components are not universally present in pure state in the lower mantle. Real lower-mantle rocks presumably present mixtures of different materials. As estimated by Allegre from isotope data, the lower mantle contains  $20 \pm 10\%$  depleted mantle, [20]. It is seen that our lower limit for this value coincides with the upper limit estimated by Allegre.

According to the data in [21], which were obtained based on Ar isotopic evidence, the contribution of continental material to the lower mantle does not exceed 0.24%, a value that also lies within our estimated range of the possible percentage of continental crustal material in the lower mantle (0-0.4%).

Several hypotheses have been advanced to explain the origin of the material enriched in incompatible elements, which is traced in the isotope signatures of the magmas of some intraplate ocean islands. A very popular viewpoint is that the geochemical features of these magmas are caused by subsidence of continental material enriched in incompatible elements, probably in the form of abyssal sediments subducted together with oceanic plates, into the deep-seated mantle horizons. For some volcanic complexes, this hypothesis is confirmed by the correlation between continental signatures, for example, between the size of Nb mining and the <sup>87</sup>Sr/<sup>86</sup>Sr ratio [22]. The mass balance calculations presented in this work are consistent with the possible contribution of a few tenths of a percent of continental material in the lower mantle processes. However, the predominant part of the enriched mantle bears no signatures of continental material, and presumably originated owing to differentiation of the mantle itself.

A very attractive idea is that the enriched material is transported into deepseated zones owing to break-off (delamination) of subcontinental lithospheric blocks, which were influenced by the low-degree melts enriched in incompatible elements, i.e., silicate or carbonate melts penetrated into the lithosphere from the underlying convective mantle [23]. Accumulation of Re-Os isotope data cast some doubts on this hypothesis, since the residual material that predominates in the lithosphere has very low Re/Os ratios, which resulted in time-integrated low radiogenic <sup>187</sup>Os/<sup>188</sup>Os ratio. At the same time, ocean-island magmas bearing Sr and Nd isotope signatures of enriched mantle typically show no anomalously low <sup>187</sup>Os/<sup>188</sup>Os. We suggest that this fact is not decisive. Actually, the subsidence of lithospheric blocks in the lower mantle can significantly affect their Os isotope characteristics owing to the percolation of sulfide melts. This explanation can be substantiated by the fact that temperatures in the upper mantle are higher than cotectic temperatures in the Fe-Ni-S system, i.e. sulfide melts can be universally present in the interstices between crystalline silicates and oxides. New geochemical data indicate that the infiltration of sulfide melts in the mantle rocks can significantly affect Os isotope composition [24].

The considered model of the mass balance of enriched and depleted reservoirs in the primitive mantle cannot be taken as final. The crystallization products of the global magmatic ocean, including predominant magnesian silicate perovskite with subordinate calcium silicate perovskite and ferripericlase have to be added to the enriched components [25]. Such a mineral assemblage can be significantly enriched in incompatible elements owing to their high distribution coefficients for calcium silicate perovskite. The presence of such a material in the lowermost third part of the lower mantle, which is supposedly isolated from the global convective system of the silicate Earth, was inferred from data on the planetary heat flow [15].

# CONCLUSIONS

1. Mass balance calculations of incompatible elements for different possible reservoirs of the bulk silicate Earth show that the lower mantle, in addition to primitive mantle material, contains a depleted MORB-source mantle component and domains enriched in incompatible elements.

2. Only an insignificant part of the enriched material in the lower mantle can be represented by the continental crustal component. The predominant part of the enriched mantle resulted from differentiation of themantle itself.

3. The enriched material was presumably supplied into the deep-seated zones of the silicate shell owing to delamination of subcontinental lithospheric blocks affected by mantle metasomatism and their subsequent involvement in mantle convective system.

4. Variations of Os isotope composition in the subsided lithospheric material are caused by the infiltration of sulfide melts.

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#### GEOLOGY AND AGE OF KHAN-BOGDINSKY MASSIF OF ALKALINE GRANITOIDS IN SOUTHERN MONGOLIA

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#### INTRODUCTION

The Khan-Bogdinsky massif (KHBM) of alkaline granitoids, one of the largest in the world, occurs in the south of the Gobi desert (Mongolia) and occupies the central part of the Late Paleozoic trough, in which the island arc calc-alkaline (varying alkalinity) differentiated volcanic sequences are replaced by rifting bimodal basaltcomendite alkaline granite associations. Position of KHBM is controlled by the intersection of sublatitudinal Gobi-Tien Shan rift zone with the oblique transverse fault, controlling bimodal magmatism. The massif is composed of Western and Eastern ring-like bodies; it has sharp crossing magmatic contacts with the island arc sequence and tectonic fault boundaries with bimodal sequence. The internal ring structure is characteristic of the western body and is specified by location of ring dykes and roof sagging of the hosting island arc sequence. Considering preliminary gravimetric data, the massif looks as a flattened body (laccolith), with the basement dipping in a step-like manner towards north-west.

Reliable geochronological data was acquired for the rocks of the massif and its setting: alkaline granites (main intrusive phase (MIP)), measured by U-Pb method after zircon (290±1 Ma), Ar-Ar method after amphibole and polylithionite (283±4 Ma and 285±7 Ma), Rb-Sr isochron method (287±3 Ma), about 290 Ma. It is close to the age, defined after zircon, of red non-alkaline biotite granites (290±1 Ma) and Rb-Sr age of bimodal sequence of the southern setting of the massif. The earliest magmatic formations enclose the rocks of the island arc complex within the massif outlines and in the roof sagging, which formed about 330 Ma ago. The geodynamic model of KHBM formation involves collision of the Hercynian continent with the hot spot of

paleo-ocean as two versions. In the first version, after collision of the edge of the North-Asian paleo-continent with the hot spot the mantle of plume processed the subducted lithosphere, having produced something like the asthenosphere window, which was the source of rifting magmatism and KHBM proper. In the second version of the model the hot and buoyant mantle of plume smoothed the subducted plate, stopped the island arc magmatism, and possibly created the local convective system in the asthenosphere of the mantle wedge, which contributed to formation of the sources of rifting magmatism. A huge volume of alkaline-granite plutonic and volcanic rocks of KHBM and its ring structure are formed due to the caldera mechanism of intrusion and evolution of its magmas.

Alkaline granitoids and their volcanic analogs (pantellerites and comendites) are comparatively rare magmatic rocks searched for by geologists, petrographers and geochemists. This is conditioned both by rarity of these rocks and frequent occurrence of rare-metal mineralization, relationship with certain geodynamic settings, uncommon mineral and chemical composition, causing many questions on the genesis. This is typical for huge massifs of alkaline granitoids, exemplified by KHBM in the southern Gobi of Mongolia (over 1500 km<sup>2</sup>). In contrast to many large massifs of alkaline granites, e.g. Cave on the Kola Peninsula, the KHBM is well outcropped, and it is feasible to observe relationships both with host rocks, and between different granitoids within the massif, as well as its internal structure.

The massif was practically not studied so far, and its territory was not geologically surveyed even at scale one million. It was recognized by the Russian geologists in 1950s and 1960s, e.g. Yu.Ya. Petrovich, B.M. Kazakov in 1958. In 1965 it was assessed for piezo-quartz reserves by I.A. Kuryaev, V.M. Andreev, P.V. Osipov, N.N. Senkevich, but alkaline granites have never been reported. Our further investigations were conducted in 1960s through 1980s as part of the Soviet-Mongolian Expedition, Academy of Sciences of the USSR and Mongolia. Only In 2000s the survey indicated that KHBM is composed of typically alkaline granitoids [2, 4] and is involved in the South Gobi belt of alkaline granitoids extensively striking in sublatitudinal direction [5]. Referred to as Gobi – Tien Shan, it makes part of Late Paleozoic rifting province of alkaline rocks of the Central Asian fold belt [20].

#### **GEOLOGICAL POSITION AND STRUCTURE OF KHBM**

The first geological map of KHBM and its setting was built based on numerous walk geological routes (airborne photos were missing). The contributions were made by geologists A.V. Goreglyad, L.P. Zonenshain, M.V. Durante, O.D. Suetenko, G. Enzhiin, A.Ya. Saltykovsky. This map was published in the book [1]. After Landsats of this territory were acquired, the map was specified, and the last variant of this map

was built with spectro-zonal space images applied (see Figure 1). On the images the massif looks as craters on the lunar surface (Fig. 1a), and consists of the Western and Eastern intrusive bodies. Alkaline granites occupy low forms of relief, and hosting hornfelsic rocks produce elevated ring-like framing of the massif, in places overlapped by younger loose sequences. The form of the western and eastern bodies of alkaline granitoids is rounded, but such shape is particularly typical for the western body. The internal structure of the body is circular, emphasized by location of roof sagging and numerous ring dykes (Fig. 16). The general northeastern extension of massif agrees with the strike of the system of the latest dykes of micro-grained syenites and monzonites in the massif. (Fig. 16).

**Geological structure of the massif setting**. The KHBM of alkaline granitoids is located in the center of the depression with same name, filled with Upper Paleozoic sedimentary-volcanogenic sequences, framing the massif in the south, east and north in the exo-contact zones of the massif and in the remote periphery. The granitoids of the massif and its rocks are overlapped with Upper Cretaceous red color sedimentary sequences containing the remains of dinosaur eggs, as was stated by Mongolian geologist Hos Bayar.

The most ancient (Pre-Caboniferous) rocks of the region are greenstone siliceous and terrigenous sediments with a visible thickness 1500 m, outcropped in the southwest of KHBM (in the region of well Gashun-Sukhoy Khuduk). The lower boundary of this complex from data by Zonenshain L.P. [2] consists of greenstone altered basic volcanics, cut by numerous bodies of hornfelsic gabbro, as well as siliiceous and claysiliceous schists and jasper. The upper sequence of this complex consists of siliceousclay schists, polymictic sandstones and conglomerates. The age of this complex is defined as Pre-Carboniferous, and by formation composition this complex belongs to ophiolites (eugeosyncline formations in [2]). The upper Paleozoic stratified sequences of the Khan-Bogdinsky depression are the sequences of Carboniferous and Permian age, their rocks unconformably occur on the rocks of the ophiolite complex. The most ancient Upper Paleozoic formations of the depression make up a sedimentary sequence composed of sandstones, siltstones and conglomerates and characterized by and tuff-sedimentary sequence with Turonian-Visean fauna, the Middle Carboniferous flora. The higher horizons of the section consist of essentially volcanic rocks, which are united into andesite (differentiated) and bimodal (basalttrachyrhyolite-comendite) association of rocks [18].

Andesite (differentiated) association is subdivided into the lower sedimentaryvolcanogenic (500–600 m) and upper lava (up to 2500 m) sequences. The crosssection of sedimentary-volcanogenic sequence begins with a series of grey andesite flows, which are replaced at the top by conglomerates with pebbles of andesite porphyrites, rarely granites and gabrodiorites, and up the section by rhythmically alternating tuff sandstones, tuff siltstones and tuffites with flora of Middle-Upper



Fig. 1. Landsat-7 of KHBM of alkaline granites (a) and scheme of its geological structure (b).

**1–5 – rocks of massif:** 1 – alkaline granites of the main intrusive phase, 2 – fine-grained aegerine granites, 3 – dykes of micro-grained syenites and monzonites, 4 – late pantellerites, 5 – alkaline granite-porphyres; **6–8 – volcanic rocks of bimodal association:** 6 – comendites, trachyrhyolites, their tuffs and ignimbrites, 7 – basalts; 8 – non-dismembered rocks of association; 9 – rocks of island arc associtation (differentiated complex); 10 – Pre Late Paleozoic complexes; **11–13 – roof of massif:** 11 – hornfelsic rocks of differentiated complex, 12 – biotite and amphibole-biotite granites and granosyenites, 13 – sites of hematithized alkaline granites MIP («red granites»); 14 – faults; 15 – boundaries of volcanic flows.

The scheme marks sampling sites for isotope dating.

Carboniferous [2]. The upper parts of cross-section display flows of andesites irregularly distributed over the section.

The lava sequence is primarily composed of flows of andesites and andesitebasalts. On the area of Khan-Bogdinsky depression its structure shows facial alterations, linked with different conditions of eruptions. Its northern parts basically contain dark-cherry and violet-grey lavas, in places with well expressed slag zones, which formed under subaerial conditions. Along the southern part edge of the depression the structure of the sequence is defined by black massive andesites and andesite-basalts with infrequent layers of jasper and hyaloclastites, that indicates their formation in the aqueous medium. The cross-section is completed with beds of grayish common and sintered dacites – trachyrhyolites. Such a change of basic and medium rocks is fairly common for cross-sections of Carboniferous volcanic sequences of Southern Mongolia, that allowed to distinguish such rock associations as differentiated volcanic complex [20].

*Bimodal volcanic association (or bimodal complex)* is composed of alternating packets of rocks of markedly different composition: (i) basic lavas and (ii) tuffs, ignimbrites and lavas of comendite and trachyrhyolite composition. The rocks of the complex occur along the northeastern and southwestern termination of KHBM. The structure of the eastern lava field is marked by a significant ruggedness with volcanic sequences sloping gently into fine sand. That is why the rocks of the complex occur as thin homogeneous fragments of the cross-section. This was the cause to initially divide them into liparite and basalt-liparite sequences [2]. However the further research [18] enabled to unite these sequences into a bimodal complex.

The most complete cross-section of the complex is the case within the volcanic field of the KHBM southern termination. The rocks conformably overlie volcanic sequences of the differentiated complex, separating from them with sandstone and gravel beds of irregular thickness. At the top of cross-section there is sequence of volcanic rocks (thickness in meters): (1) basalts and andesite-basalts (400), (2) trachydacite and comendite tuffs and ignimbrites (200), (3) olivine basalts (350), (4) comendites and trachyrhyolites (100), (5) basalts (100), (6) trachyrhyolites, comendites and their sintered tuffs and ignimbrites (>800 m). The total thickness of the cross-section is in excess of 2000 m. The volcanic piles are filled with a series of bedded intrusions, of hydrothermally altered comendites and alkaline granite-porphyries.

Both volcanic fields show a series of similar structural units, indicating affinity of their cross-sections. Thus, cross-sections of both fields are topped by series of compositionally similar trachyrhyolite (comendite) lava flows, ignimbrites and sintered tuffs. The low parts display basalts and andesite basalts characterized by a number of common morphological features, such as flow thickness, abundance of almond-like, slagging and burnt rocks. They both share the feature of having subvolcanic formations, which are dykes and bedded bodies of alkaline granite porphyries, comendites and trachyrhyolites and rarely basalts. The dykes are particularly numerous in the northeast of the lava field. They cross practically entire complex of volcanogenic formations, except for the upper series of acid lavas, ignimbrites and sintered tuffs, and belong to two differently oriented and spatially separated systems. Close to the massif these are steeply dipping bodies of east – northeast strike lying on the continuation of the main faults and accompanying dykes, observed through entire massif of alkaline granites. Away from the massif there are dykes of alkaline granite porphyries, which belong to the arc-bended (circular) system of bodies, dipping under angle ~  $45^{\circ}$  in the western and southwestern directions, i.e. towards the massif of alkaline granites. The amount of dykes is truly high. In the southern part, the dykes with thickness 5–9 m occur every 50 m across the general strike of the dyke system. These bodies are more abundant towards north.

Bedded intrusions of trachyrhyolites and alkaline granite-porphyries are widely developed in the structure of both volcanic fields. In the cross-section of rocks of bimodal complex they commonly lie under the third series of ignimbrites and sintered tuffs. The composition of subvolcanic intrusions of acid composition they define the rocks which produce a continuous range of transitions from glassy to properly crystallized varieties. The endocontact zones of intrusions are composed of glassy and fluidal comendites and trachyrhyolites, which correspond to erupted rocks. The degree of crystallization increases towards the central parts of bodies, as well as along their strike towards the granite massif. The rocks acquire porphyry-like structure, defined by the development of phenocrysts of quartz and alkaline feldspar in microgranite or granophyre bulk mass with microlites of alkaline amphibole involved. Crystallized varieties of subvolcanic bodies become similar in composition and appearance of rocks to the dykes of alkaline granite-porphyries of KHBM.

The above characteristics indicate a considerable similarity of the lava fields structure over the northeastern and southwestern termination of the massif. Also considering that along the southern and eastern boundaries of the massif there is a narrow zone of exposed rocks of bimodal complex, uniting both fields by a continuous zone of exposures, it appears that both volcanic sites belong to the one larger volcanic field. It initially occupied the whole area of the granite massif, and made up no less than  $1800 - 2000 \text{ km}^2$ . Abundance of ignimbrites and sintered tuffs, continuous strike of volcanics, as well as a large thickness of cross-sections with a limited spatial distribution enable an assumption that volcanic piles formed within the structural depression of the type of a flatly lying trough with the beds dipping towards its center. In the south it is bounded by a large fault, in which volcanic formations dip relative to the structure of the basement more than by 2000 m. Its northern side is overlapped by sediments of Mesozoic depressions, but judging from the ratios observed 40 km to the northeast from the KHBM this side was for4merl;y a normal fault with the amplitude close to that of volcanic graben bottom subsidence, like along its southern boundary.

**Palaentological age of volcanic complexes**. The age of subaerial volcanogenic formations within the KHBM was defined from the distribution of plant remains of the Late Paleozoic [2]. As reported previously, the cross-section of the Late Paleozoic stratified formations of the region exhibits: in the lower part it is a sedimentary sequence, and in the center these are the rocks of differential complex, which consists of (i) volcanogenic-sedimentary sequence at the base, and (ii) lava andesite sequence at the top, and in the upper part there are rocks of bimodal complex, which were subdivided by M.V. Durante et al. [2] into lipartite and basalt-liparite sequences. The rocks of the sedimentary sequence [2] enclose flora of the Middle Carboniferous, which defines the lower age boundary of volcanics development. Some tuffites of sedimentary-volcanogenic sequences, making up the lower horizons of differential complex contain flora charactristic of the Middle-Upper Carboniferous. This flora is the evidence that eruptions of andesites and formation of magmatic rocks of the differentiated complex proceeded in the second half of the Carboniferous.

One more group of plant prints was found in the tuffs among acid volcanics of liparite sequence [2]. These prints represent fragments of cordaites typical of the Lower Permian. As was stated above, the rocks of the liparite sequence are united with the rocks of the basalt-liparite sequence, devoid of plant remains, into a single bimodal complex [16]. Therefore, the age of rocks of bimodal complex should be regarded as Lower Permian.

Settings of the Upper Paleozoic complexes formation. Stratified sequences within the KHBM depression are typical of continental volcanic formations. The early (andesite or differentiated) complex of these rocks distributed all over the area of South-Mongolian Hercynides are commonly accompanied by discordant massifs of non-alkaline granitoids. In particular, toward west and northwest from the massif, as well as to the north from the village Khan-Bogd there is a plutonic association of monzonite-syenite-granosyenite-granites (MSGG), referred to banatite [3]. This association is dominated by pinkish-grey amphibole-biotite monzonites and pink quartz and quartz-devoid syenites, cut with dykes of the same kind of rocks, as well as basalts and pink granites. Towards south-west from the massif there are separate plutonic bodies of pink leucocratic biotite granites. Such a body makes up the Mot Eligenny-Khira (coordinates of the center of the massif are 43°14'N. 106°42' E). It consists of pinkish-grey monzonites, cut by numerous veins of red amphibole-bearing granosyenites, quartz syenites, and the latter are cut by the dykes of altered basites. The contact between monzonite and red granosyenites was observed 1 km to the south from that Mount. In the vicinity of Mount Barun-Kharatu-Ula there is an intrusive contact of monzonites and red grano-syenites with volcanics close in composition.



Fig. 2. Scheme of geological structure of site «Armstrongite».

1 – alkaline granites of the main phase of the massif; 2 – pegmatoid alkaline granites; 3 – «cleavage» of ekerite and pegmatite bodies; 4 – silicified alkaline granites; 5 – quartz rocks; 6 – basic effusives; 7 – rhyolites and breccia of basic effusives in the basement of rhyolite sequence; 8–11 – rocks of schlieren pegmatites (on the inlet): 8 – ekerites with «suns» of aegerine, 9 – fine-grained ekerites, 10 – alkaline granitic pegmatites with armstrongite, 11 – zone of anchi-monomineral armstrongite; 12 – elements of rock occurrence.

These are dark-grey hornfelsic andesites, trachytes and trachydacites. All these rocks are cut by apophyse of alkaline granites near water well Tsokhon-Usu-Khuduk to the west from Khan-Bogdinsky massif.

Magmatic associations MSGG within the Hercynides of South Mongolia are more widespread, than normal granites, which predominate in Caledonides of Central Asian foldbelt [21], and in Northern Mongolia. This association refers to intrusive analogs of volcanics of the differentiated complex, that is why enclosing are earlier Pre-Late Paleozoic rocks of ophiolite ciomplexes, and similar in composition and age rocks of the Early and Mid Carboniferous volcanic sequences. Jointly with latter they produce a marginal volcano-plutonic belt, which was formed in the Early and Mid Carboniferous within the territory of South-Mongolian Hercynides, which was an active continental margin of the Paleo-Asian continent [13, 20].

In contrast to magmatic associations of the marginal belt the rocks of the bimodal complex and conjugate massifs of alkaline rocks are under structural control. They are confined to the chain of troughs, which is traced through all southern Mongolia, and is controlled by a system of faults, confined to the boundary between South Gobi micro-continent and Hercynides of the South Mongolian zone. This chain of troughs host zones of dykes, basalts, comendites, trachyrhyolites and alkaline granites, their formation is related to the large amplitude extension. Such features of structural position of rocks of bimodal association, as well as presence of widespread alkaline granites and comendites, the indicators of continental rifting setting, enabled the zone of their distribution to be distinguished as the Gobi – Tien Shan rift zone.

Roof sagging in the massif. Figure 16 displays the contribution of alkaline granites in the roof sagging. They commonly compose the peaks of hypsometric elevations. For example, the highest point of the massif Mount Khan-Bogd is composed of the roof sagging. The same are mounts Buste, Sher and others. They are underlain by the apical parts of the massif, in places occurring as schlieren pegmatoids (stock sheiders) and ekerites. The roof sagging basically consists of dark grey hornfelsic volcanic rocks, rarely cutting them by subvolcanic rhyolites, similar to the rocks of the island arc association. They are exemplified by sites Armstrongite (Fig.2) and Severny (Fig.3), in which the roof of the massif is made of stratified darkgrey glassy dacites and their breccias cut by the bodies of pink and white rhyolites. Position of these sites is shown in Figure 16. In the other sites of the massif the rocks of the roof are andesite porphyrites, dominating in the zone of the northern endocontact of the massif and rarely the tuff sedimentary rocks. The complex of rocks, involved in the structure of roof sagging is compared to the rocks of the differentiated complex, and thus indicate that formation of the massif occurred at the depths below the base the bimodal complex. Also note that no evidence assimilation between the rocks of the roof sagging was observed with alkaline granites of the massif.



Fig. 3. Scheme of geological structure of site «Severny».

1 -loose valley type sands, 2 -alkaline granites of endocontact facies, 3 -«stratified» bodies of ekerites and pegmatites, 4 -dacites and tuff breccias of dacites, 5 -rhyolites, 6 -bodies of ekerites, 7 -bodies of alkaline granitic pegmatites, 8 -elements of rock occurrence and elements of trachyte fluidization.

In contrast to the roof sagging of volcanic rocks xenoliths of pink biotite leucocratic non-alkaline granites and granosyenites close to the enclosing association MSGG, occupy low parts of relief in the northern part of the massif and on the contact of the western and eastern bodies (Fig. 16). In the northern part of the massif the roof sagging in the form of biotite granites produce concentric fields of the rounded fragments of biotite granites, as well as transitional amphibole-biotite granites, being the result of interaction of biotite granites is quite common. Presence of breccias of rounded segregations of biotite leucogranites cemented with alkaline granites reminds of interaction of «cold» magmas of biotite leucogranites with «hot» magmas of alkaline granitoids when magma is intruded into magma. That is why, it seems that magmas of these two granite types could occur in different time (biotite leucogranites are somewhat earlier, than alkaline granites), and are close in time of formation.

In the west-lying body the roof sagging disclose its circular structure, separating the internal part of the from the external one. Diverse dyke alkaline granitoids, including their rare-metal varieties, are spatially located in the roof sagging in the massif. In the eastern body the roof sagging is located in its western part, partly tracing the contact zone of the western and eastern bodies of alkaline granites. The hypsometric level of roof sagging shows that the depth of erosional cross-section of the massif is not in excess of a hundred meters. Location of the massif of alkaline granites in the upper part of the of the Upper Paleozoic island arc, which is overlapped by bimodal sequence is possibly due to a change of regime of tectonic compression in formation of island arc sequences into tectonic extension in the formation of the bimodal association with its numerous dykes and volcanic units (Fig. 16, [2]).

Contacts of the massif and its morphology. The contact surface of the massif of alkaline granites with hosting rocks of the western body in the west, south and north gently (5-15°) submerges underneath hosting rocks. The contacts of all varieties of alkaline granitoids of the massif with hosting rocks are very sharp and distinct, in places with xenoliths and apophyses. In the exocontact halo both hosting rocks roof sagging are hornfelsic. Considering all evidence, the granitoids of KHBM formed in the process of magma intrusion, with the traces of its mechanical impact on host rocks being evident due to phenomena of xenolith rotation, their pull-apart and migration. This is indicated by small fissure bodies available among relics of the massif roof and being its apophyses. The area of flatly dipping apical part of the massif does not strike, judging from the width of hornfels halo, further 1 km away from the contact. Besides, in places the hornfelses of the marginal halo are cut with vertical dykes outlining the contour of the massif in the above-lying sequences. All this evidence indicates the primary circular contour of the massif and its steep lateral constraints, as well as existence in the massif roof of circular fissure zone, being continuation of its lateral surface. The roof sagging within massif lie on the alkaline granites nearly horizontally and flatly dip towards south, west and north in the endo-contact parts. Dipping of the flat separation of alkaline granites of the massif commonly coincides with dipping of the contact surface of granites. These data indicate poorly convex form of the upper surface of the massif and shallow erosional occurrence.

We draw particular attention on the ratio of alkaline granites of the massif with the rocks of the bimodal complex, which indicate tectonic, post-intrusion contact between them. Both volcanics and granites are crushed on the contact and are transformed into a tectonic breccia. Volcanic rocks along the extensive line of the contact are fresh and are devoid of the evidence of impact produced by the granites of the massif. Besides, in the recent erosional cross-section with rocks of the bimodal complex the same hypsometric level is marked for the rocks of the differentiated complex (as roof pendants) which lie at depth about 2 km in the Late Paleozoic formations. This is the evidence that after emplacement, the KHBM experienced vertical displacement with the amplitude over 2 km.

In 1973 the KHBM was cut by the gravimetric route from SE to SW across the Mount Khan-Bogd with observation spacing 2-3 km by geophysicists under leadership by Yu.A.Zorin. The quantitative interpretation of gravimetric data [2] indicated that the massif looks as a flattened body (laccolith), with the base dipping in a step-wise manner to NW up to 7 km, producing the depth in the central part 4.5 km, and in SE part it is 1-2 km. We assume that in the thickest part of the body there was a feeding channel, and the step-like dipping of the base is due to presence of faults.

Relationships and sequence of formation of the rocks of the massif. The general sequence of formation of magmatic rocks of KHBM is as follows: (from early to late): (1) light grey to pink alkaline coarse-, medium-grained granites of the main intrusive phase (MIP) with accessory elpidite, making up the larger part of the Western and part of Eastern bodies; (2) dykes of fine-grained ekerites, porphyry-like ekerites, bedded alkaline granitic rocks, composed of alternating ekerites and pegmatites, pegmatoid alkaline granites, pegmatites, which occur in the Western body, underlining its circular structure; (3) lilac fine- to medium-grained alkaline (normally aegirine or arfvedsonite-aegirine, often myarolitic) granites with accessory zircon, making up a larger part of the Eastern body; (4) dykes of micro-grained to glassy dark-blue or dark-green pantellerites, (5) dykes of alkaline porphyry-like finegrained leucogranites, developed in the northern part of the Eastern body; (6) dykes of micro-grained red granosyenites and monzonites producing the system of extensive dyke bodies of sublatitudinal strike in the Western body; (7) rare carbonate and quartz-carbonate veins. The relationships between the rocks do not cause doubt relative to the enumerated sequence of their formation (Fig. 2). The contact between alkaline granites The contact between alkaline granites MIP of the Western and Eastern bodies is more complex, as in the contact zone granites of these bodies often become alike. In 2 km to the south-east from Mount Bust there is cutting of alkaline medium-grained amphibole-bearing granites with accessory elpidite (MIP of the western body) by violet fine-grained againe-bearing alkaline granites with accessory zircon (MIP of the Eastern body).

The fine-grained violet aegirine-bearing alkaline granites of the Eastern body produce flatly lying (subhorizontal) intrusive body about 100 m thick, without any evidence of a circular structure. In the base it contacts alkaline granites MIP of the Western body cutting them and the vein body of pegmatoid alkaline granites of spotted texture (Fig. 4), often with erosional windows, composed of MIP of the Western body.



Fig. 4. Ratios of alkaline granites of the Western and Eastern bodies.

1 -alkalune granites MIP of the Western body, 2 -pegmatoid alkaline granites, 3 -alkaline granites of the Eastern body, 4 -ekerites.

In the roof of granites of the Eastern body there is sagging of the roof of the host rocks described above. It appears that intrusion of magma of granites of the Eastern body proceeded in subhorizontal direction between the apical surface of alkaline granites MIP of the Western ring and their roof. Granites of the Eastern body practically lack dykes of the 2<sup>nd</sup> phase. However the dyke of glassy pantellerites (4<sup>th</sup> phase) was found there. It cuts both granites of the Eastern and Western bodies. In the north of Eastern body there are extensive sublatitudinal dykes of porphyry-like alkaline leucogranites. It is likely that granites of the Eastern body are nearly synchronous to dyke rocks of the 2nd phase of the Western body.

The ring structure is observed in the granites of the Western body, and it is outlined with a ring location of the roof sagging, as well as ring dykes of rocks of the  $2^{nd}$  phase. There is a spatial relationship of circular dykes of the  $2^{nd}$  phase and roof sagging. Later dykes of microsyenites and monzonites of the  $6^{th}$  phase are confined to sublatitudinal fissures concordant with a general strike of the massif. Smaller dykes of such rocks are submeridionally NW-directed. In the southern-most dyke of the  $6^{th}$  phase the structure is more complex: the earliest (endo-contact) rocks are dark-grey diorites transformed into pinkish-grey monzonites and quartz syenites, similar to analogous rocks of the island arc (banatite) association, which are cut by thick bodies of porphyry-like ekerites, cut by later fine-grained ekerites. Porphyry-like and fine-grained ekerites are referred to the  $2^{nd}$  phase.

and syenites of this complex dyke with alkaline granites MIP of the Western body are not clear. On the one hand, they compose the dyke body in MIP granites jointly with cutting ekerite. On the other hand, they can be roof sagging in ekerites, confined to the hypsometric uplift in MIP cut by ekerites. If all enumerated rocks of this complex dyke are later as compared to MIP granites, one should admit the multi-stage of island-arc magmatism in the massif vicinity (prior to alkaline granites and after them). If the island arc association of the dyke is the roof sagging, as was observed in the other roof sagging, it is more logical to accept the younger formation for the main part of island arc magmatism as compared to alkaline MIP granites. The most widespread dykes of ekerites of the 2<sup>nd</sup> phase are fairly diverse in structure and texture, not so much in composition. They involve rare metal and "barren" varieties not differing much but in lack or abundance of rare metal minerals.

With this in mind, the following scheme of the KHBM emplacement is proposed. In the early phase the rocks of the bimodal basalt-comendite association were formed, they overlap differentiated basalt-andesite-rhyolite island arc association of high alkalinity and its plutonic analogs. The next to form were medium-coarse-grained alkaline granites MIP. The magma, from which they were formed, filled the circular chamber of the Western and Eastern bodies, and the feeding channel was displaced most likely to the north-west from the center of the Western body. The magmatic chamber, consisting of MIP granites, lied between pre-Carboniferous complex and Upper Paleozoic sequences of the island arc specifics within the andesite and liparite series of the massif framing, composing its roof. This formation erupted subhorizontally in SE direction. Most likely, the cause of intrusion of abundant alkaline-granite magma was the change of compression conditions into tectonic extension, when the Upper Paleozoic trough of the KHBM originated. It is very possible the massif roof descended into a magmatic chamber, as it happens in collapse calderas due to eruption of comendites of the bimodal complex, which resulted in formation of the ring structure of the Western body. It seems that ring faults united the roof of crystallized body of alkaline granites with a residual magmatic chamber. The dykes of the  $2^{nd}$  phase intruded along them, the larger part is confined to the sub-horizontal surface in the apical zone of MIP granites. A bit later along the roof of MIP granites there was an intrusion of the new portion of alkalinegranite magma which led to formation of fine- medium-grained aegerine-bearing alkaline granites of the Eastern body.

#### ISOTOPE AGE OF MAGMATIC ROCKS IN KHBM

Some papers report K-Ar [2, 19] and preliminary Rb-Sr [1, 15] estimates of the age of alkaline granitoids of KHBM. The K-Ar estimates show a large scatter (218–362 Ma) and cannot be accepted satisfactory. They were obtained from different minerals (mica, amphiboles, alkaline feldspars), differently preserving the radiogenic argon with the processes overlapped. It is noteworthy that alkaline feldspar has the scatter of ages from 218 to 306 Ma. This mineral commonly loses radiogenic argon in the processes overlapped, so the age defined from this mineral is underestimated. Amphibole retains radiogenic argon much better. As to alkaline granites, the K-Ar age after amphibole varies from 226 to 362 Ma. The maximum age, obtained after arfvedsonite and polylithionite (362 and 342 Ma, accordingly), is defined for alkalinegranite pegmatites, for which the age is overestimated for unknown reasons even with K-Ar system [7]. Without these extreme estimates the K-Ar age of alkaline granites varies within narrow limits between 266 and 310 Ma. The age definitions were made for granites by Rb-Sr method, and they amount to 282±22 Ma [1] and 286±9 Ma [15]. But they were based on the erochrons with a fairly large scatter of isotope composition, therefore these figures are hardly reliable.

This contribution provided new data both for Rb-Sr system (e.g. for volcanics of the southern termination of the massif), and for zircons by U-Pb method, their ages being most reliable, as well as amphibole and polylithionite by Ar-Ar method.

The accessory zircon was extracted for geochronological studies by standard technique with utilization of heavy liquids at the isotope laboratory of  $H\Gamma\Gamma\mu$  RAS. Zircon was decomposed and Pb and U were chemically separated after modified technique by T. Crow [22, 23]. The level of blank experiment was not in excess of 50 pg Pb. The isotope composition of Pb and U was performed on a mass-spectrometer Finnigan MAT 261 in the static regime or with an electronic multiplier (coefficient of discrimination for Pb is 0.32±0.11). Experimental data was processed with programs "PbDAT" and "ISOPLOT" [24, 25]. When the ages were calculated, the commnly accepted values of uranium decay constants were used [27]. The corrections for ordinary lead were introduced in accordance with modal values [26]. All errors are provided at the level  $2\sigma$ . The analytical parameters of geochronological studies after other isotope systems are described elsewhere [7, 9].

In addition to the age of alkaline granites proper, the principal geochronological issue is the age of massifs of biotite non-alkaline leucogranites analogous to the roof sagging in KHBM, located to the west from the latter (Fig. 5). This is why the zircon ages, obtained by U-Pb method, thought to be as most reliable, were measured for alkaline granites MIP of the western body, rare-metal alkaline-granite pegmatites of the  $2^{nd}$  phase, for trachydacites of the roof sagging of the Buste Mount, biotite

leucogranites of the massif located to the west of KHBM (Fig. 5) analogous to xenoliths in alkaline granites. The sampling sites are shown in Figs. 1 and 5.



Fig. 5. Scheme of the structure of the central part of Gobi – Tien Shan rift zone and its position in the system of structures of the marginal belt of Southern Mongolia.

The inlet shows the structural scheme reflecting specific position of KHBM within Gobi – Tien Shan rift zone: 1-3 – complexes of rift zones: 1 – KHBM, 2 – fields of development of bimodal volcanic associations and alkaline granites, 3 – eroded and rift zone sites overlapped by younger rock sequences; 4 – area of Hercynides development and complexes of active continental margin; 5 – Caledonide; 6 – complexes of South Gobi microcontinent; 7 – Meso-Cenozoic depressions; 8 – massifs of biotite non-alkaline granitoids; 9 – faults of Dzun-Bainsky system; 10 – faults; 11 – boundary between Mongolia and China; 12 – sites of sampling rocks for dating and their age in million years.

**Rift zones:** Gobi – Tien Shan, Main Mongolian lineament. **Blocks of South Gobi microcontinent:** TIII-Ul – Toto-Shan-Ulanul, Ts-Ul – Tsagan- Ul. Dz-B – Dzun-Bainsky system of faults.

For MIP granites of the western body the U-Pb age (sample XE-1807) is measured from zircon as  $290\pm1$  Ma, for rare-metal alkaline-granite pegmatites of the  $2^{nd}$  phase (sample XE-1745) its was  $292\pm1$  Ma, and for pink biotite granites of the massif, located to the west of KHBM (XE-4448), and xenoliths in granites MIP it was  $290\pm1$  Ma, for trachydacites of roof sagging in the massif (sample XE-1802) the age was  $330\pm1$  Ma (Fig. 6). Considering these figures it follows that alkaline granites MIP and their pegmatites have practically the same age about 290 Ma, which is estimated with a high frequency. Pink biotite non-alkaline granites have the same age. They produce both individual massifs and large xenoliths in alkaline granites MIP. We shall underline that the age of pink granites is defined not from xenoliths but from the massif away from Khan-Bogdinsky alkaline granites Maccuba, which the influence of the latter on pink granites impossible. Besides, trachydacites of roof sagging have the age 330 Ma. For comparison, we also identified Ar-Ar ages (Fig. 7) for amphibole of alkaline granite MIP (sample XE-1808) and for polylithionite of rare-metal pegmatite of the  $2^{nd}$  phase (XE-1433). They turned out to be close, although younger against zircon dates, e.g.  $283\pm4$  Ma and  $285\pm7$  Ma, respectively. The spectrum of the ages of amphibole of sample XE 1808 is marked by well developed segment of plateau (Fig.7b), which age may be regarded as the age of amphibole crystallization. The spectrum of polylithionite ages is similar to that of mixing components with the ages about 295 Ma (polylithionite proper) and 229 Ma (secondary phase), possibly representing the time of superimposed event (Fig. 7a).

The Rb-Sr age of the island arc sequence of the southern termination of the massif, defined from two points (basalt and dacite), is 329±5 Ma (Fig. 8c). This estimate is tentative because it is defined from two points only. However, it practically coincides with the age of zircon from dacite of roof sagging, and shows similarity of the age of roof sagging with island arc sequence of the massif framing. The Rb-Sr age of bimodal sequence of the southern termination is evaluated from the isochron (Fig. 8a): all varieties of volcanics of bimodal sequence have the age 291±4 Ma (CKBO=52), for acid volcanics it was 293±15 Ma (CKBO=122, absent in the figure). The Rb-Sr isochron age for alkaline granitoids MIP of the Western body of the massif (Fig. 8 b) was determined to be 287±3 Ma (CKBO=1,8). Thus, we have got close ages for alkaline granitoids of KHBM measured by U-Pb method by zircon (290±1 Ma), Ar-AR method after amphibole and polylithionite (283±4 Ma and 285±7 Ma), Rb-Sr isochron method (287±3 Ma), close to 290 Ma. This age is close to that after zircons of red biotite granites (290±2 Ma) and Rb-Sr age of bimodal sequence of the southern termination of the massif. Geological data indicates that red biotite granites were formed somewhat earlier than alkaline granites, but the difference in their age is not detected by radiological methods, and in the first approach they can be considered synchronous in time. The earlier magmatic formations are the rocks of the differentiated complex within the massif and in roof sagging, which formed about 330 Ma ago. Available geochronological data agree well age determinations obtained from floristic prints [2]. Thus, the age of trachydacites from the rocks of the differentiated complex defined as 330 and 328 Ma, corresponds to the geological estimate of time when the complex was formed - Late Carboniferous. The Early Permian age of rocks of the bimodal complex also corresponds to the results of geochronological investigations, which showed that formation of these rocks and related alkaline granites occurred about 290 Ma.





a - alkaline granite MIP, b - alkaline granitic pegmatite of the 2nd intrusive phase, <math>c - biotite non-alkaline granite to the west of massif of alkaline granites, <math>d - trachyrhyolite of roff sagging of alkaline granitic massif.

#### GEODYNAMIC MODEL OF FORMATION OF KHAN-BOGDINSKY VOLCANO-PLUTONIC ASSOCIATION

Available geological and geochronological data point out that KHBM of alkaline granitoids, massifs of non-alkaline biotite leucogranites and associated volcanic sequences of calc-alkaline and bimodal complexes originated within the active
continental margin of South-Mongolian Hercynides. The specific feature of its development was that prior to the border 330 Ma ago it underwent the event of basalt-andesite-rhyolite magmatism [13], which resulted in formation of magmatic rocks (differentiated) series, typical for marginal magmatic belts [11, 12]. It was followed by rifting processes with specific bimodal basalt-comendite magmatism with pantellerites and alkaline granites. These rifting processes within Gobi - Tien Shan zone commenced from the border 318 Ma in the region of Tost-Nuru Mounts [9], i.e. at least 10 - 15 million years after the marginal magmatic belt ceased to evolve, and proceeded to 290 Ma (time of formation of KHBM). As a result, the extensive Gobi-Tien Shan rift zone came to existence, which was superimposed onto the marginal belt striking for over 1000 km.



Fig. 7. Ar-Ar age spectra of polylithionite XB-1433 (a) and amphibole XB-1808 (b) of rare-metal alkaline granite pegmatites of KHBM.

Formation of the Gobi-Tien Shan rift zone was in fact formation of the largest continental rift systems of the world – Late Paleozoic – Early Mesozoic rift system of Central Asia. The specifics of its development was a sequential displacement of rifting zones from the end of the Late Paleozoic North-Asian paleo-continent into the depth with formation of a series of sub-parallel rift zones, their age varying from Late Carboniferous – Early Permian (Gobi-Tien Shan zone, margin of continent) to Late



Fig. 8. **Rb-Sr erochron for volcanic rocks of bimodal association within the framing of KHBM (a), isochrons of alkaline granites MIP (b) and hosting island arc volcanic rocks of the complex (c).** 

Permian (North Mongolian zone, mostly remote from the edge of the continent). Such a displacement of rifting zones enable to relate the cause of the rift system origination and, in particular, Gobi- Tien Shan rift zone, with a sequential overthrusting of continental edge onto mid-oceanic ridge of Hercynian paleo-continent or hot spot of the mantle within its outlines [20] (Fig. 9, model A, B and C).

The existence of the mantle hot spot in the oceanic sector is indicated by the involvement of high-Ti basalts within Middle Paleozoic ophiolites of South Mongolia [14] with geochemical characteristics of OIB-type basalts (our unpublished data). These rocks, characterized by the evidence of formation in the settings, remote from transport areas, produce tectonic slabs. In the recent structure they are tectonically compatible with of rock complexes of island arcs, which evidently occurred due to accretionary processes in the formation of the South Mongolian Hercynides. It appears that presence of such rocks amongst Hercynides indicates existence at least in the Middle Paleozoic paleo-Asian ocean of the islands or lava plateau, their formation being controlled by the mantle plume, by analogy with recent islands.

Scheme 9 displays two possible variants of development of events after collision of the edge of the North Asian paleo-continent with the mantle hot spot (B and C). In the one case (B) it is assumed that oceanic islands blocked the subduction zone, causing break-off of subduction slab. The asthenospheric window was intruded by mantle plume, which affected both mantle wedge, and lithosphere, triggering tectonic and magmatic activity.

In the second variant (C) we assume that subducted lithosphere acquired additional buoyancy under the influence of ascending hot plume and it did not "sink" immediately, but gently proceeded in asthenosphere under the lithosphere base, and only at a significant distance from the edge of the continent it subsided into low mantle. As is known from the example of Cordilleras of Southern America, above the sites of flat (subhorizontal) movement of subduction plate along the lithosphere base the magmatic activity of AKO type is not the case. Such a cessation of magmatic activity occurred in the end of Carboniferous - Early Permian within the southern active margin of the North-Asian paleo-continent. However in its back part in Central and North Mongolia it was the onset of activity of the continental margin type. Formerly these changes in redistribution of continental margin magmatism were related to overflipping of continental lithosphere above subduction zone [6]. It seems that the model proposed (Fig. 9 C) is preferred for it considers contribution of mantle plume to the general geodynamic setting of magmatism formation in the marginal part of paleo-continent. In this version, plume affected continental lithosphere indirectly through partial melting of oceanic lithosphere and mantle of the mantle wedge. These both models assume that within the south Mongolian continental margin rifting processes proceeded after magmatism of continental margin type, when conditions of magma formation had not changed much. It is believed that this is the evidence to explain specific magmatism of Gobi - Tien Shan rift zone. Its structure formation was contributed by the two kinds of magmatism: bimodal and the one typical for suprasubduction settings. This assumption is exemplified by existence of andesites of calcalkaline specifics, involved in the structure of bimodal association along with basalts of OIB type [8]. Noteworthy, these are common, rather than alkaline granites, occurring within KHBM and having the same age, as alkaline granites. The same can be referred to calc-alkaline island arc associations or crustal anatectic formations, formed under the influence of hot mantle magmas. In the former scenario thuis could be associated with contribution of metasomatic mantle of mantle wedge into melting processes, initiated by mantle plume. In the second case, the same effect can be caused by plume affecting non-dehydrated oceanic lithosphere and above-lying mantle of the wedge.



Fig. 9. Model of fomation of KHB volcano-plutonic association.

Solid arrows show directions of block movement, dashed lines denote possible ways of the upper mantle convection.

Considering the issue of KHBM formation one should keep in mind its huge size, which is the evidence of presence of a great magmatic chamber in its basement. The conditions of its origination were possibly defined by the features of structural position of the massif. As follows from Figure 5, the massif is located in the site of knee-like bending of the rift zone. This bend is adjusted to the fault zone, which crosses Hercynides under a sharp angle and structures of South Gobi microcontinent, and it is traced by superimposed Meso-Cenozoic depressions of Dzun-Bain trough. The zone of faults displaces the Tsaganulinsky and Totoshan-Ulanul blocks of the microcontinent, being a large strike slip fault. At the same time, the structures of the rift zone become adjusted to the newly formed boundary of the microcontinent, they follow in parallel to it and include NE-striking dyke swarms, concordant with the strike (see Fig. 5) of Dzun-Bain system of faults. Considering that this fault zone originated after formation of sublatitudinal structures of the active margin, and that it controls the boundaries of the rift zone within KHBM and north-east off it, one concludes that the time of its emplacement was approached to the time of the rift zone formation, and, consequently, it can be correlated with transform faults. The site of juncture of this fault with the rift zone created conditions for a maximum extension of the rift zone in the basement with formation of the large cavities, capable to enclose large volumes of mantle magmas, its subsequent differentiation eventually ensured formation of voluminous alkaline granite melts, rare-metal ones included. It is important that the volumes of initial magmas for bimodal associations and alkaline granites are in this case not very large and are comparable with moderately-sized trappean formations.

It should be reminded that rifting magmatic processes in the study region commenced with simultaneous eruptions of basic and alkaline-salic lavas. The results obtained while studying the lava fields of the bimodal volcanic association in the other parts of the rift zone [18, 8] indicate that the eruptions of basic lavas proceeded from fissure channels, whereas formation of alkaline-salic rocks was associated with a large volcano of central type. In the history of volcano formation there is a completion phase of ignimbrite volcanism, during which the sequence with thickness over 800 m originated. Its outcrops are observed over the massif framing, therefore we assume that initially it occupied that part of the territory, which is currently occupied by the alkaline granite massif, where the central part of volcano, erupting alkaline-salic melts, most likely was located. The area of the sequence the area of the sequence distribution, consequently was no less than 2000 km<sup>2</sup>, and the volume was in excess of 2000 km<sup>3</sup>, with the account of a high thickness of volcanic complex

lying close to a volcano. Voluminous ignimbrites are commonly linked with formation of gigantic calderas of Vellis type [10, 17], their mechanism of formation involves collapse of the roof of magmatic chamber, feeding volcanic activity, and formation of the ring system of faults, which bound the descending block and defined the outlines of the above-forming caldera (Fig. 9D and E). The caldera probably formed due to an extension regime, existing within the rift zone and causing formation of faults, which drained the deep magmatic chamber and thus, could cause a fast outflow of fluidized (ignimbrite) melts onto the surface.

Descending of supra-roof block was obviously accompanied by formation of detachment cavities, lying from its lateral boundaries both inside descending (caldera) block, and in the enclosing framework following pull-apart mechanism (Fig. 9E). Widespread bedded intrusions of comendites and granite-porphyres within the massif framing are likely the evidence of such an interaction between the rocks of the frame and collapsed caldera block. In the structure of the latter the detachment cavities could be fairly sizable due to differentiation of movement of its different sites, due to heterogeneity of friction forces along the lateral surface of the block causing its splitting. Considering the stratified structure of the crust of the caldera block the largest cavity could originate above the folded-metamorphic massive basement in stratified sequences of the Early and Middle Carboniferous rocks, where it should enclose melts ascending from deeper magmatic chamber. It appears that this was the way how the intermediate magmatic chamber was formed. It was turned into the massif of alkaline granites after differentiation and hardening of crystallized melts. Presence of two isometric bodies in the structure of the massif is obviously the indication of one more act of caldera formation, which could be looked upon in terms of the model, proposed by Williams [28, 29]. It explains formation of subsidences above shallow laccolith due to redistribution of melts in the horizontal line, when the magmatic chamber was expanding. Following this model, it is feasible that in KHBM magma rushed in from the laccolith in horizontal direction along the weakened zone controlled by northwestern system of faults and dykes. A repeated subsidence of the caldera block should have created the concentric system of dislocations in the structure of the massif. It had to be used in formation of ring dykes within the massif. We assume that deep differentiation of alkaline-granite magmas was due to the fact that the massif located within the rift zone, was not affected by stresses which could violate the course of processes of differentiation.

Emergence of the massif on the surface (Fig. 9F) was related to the uplifting of the within-caldera block, which occurred at the post-caldera stage of the territory development, possibly after folding involving volcanic sequences of the bimodal complex, before the Late Mesozoic platform cover was formed in the region. The data is lacking on a precise evaluation of time of that event, so it is not feasible so far to relate this process with some geological event in post- Carboniferous history of South Mongolia. Nevertheless, we may assume that it was due to processes of collisional compression, which occurred along the continent margin because of closing of Paleo-Asian ocean and affecting deep crust level, including deep magmatic chamber and causing deformations with a redsitribution of plastic material underneath caldera block with a subsequent squeezing upwards.

# CONCLUSIONS

1. This investigation provided results on the geological position, internal structure and tectonic position of KHBM of alkaline granitoids, one of the largest in the world. It lies in the south of the Gobi desert (Mongolia) and occupies the central part of the Late Paleozoic trough, in which the island arc calc-alkaline (varying alkalinity) differentiated volcanic sequences are replaced by rifting bimodal basalt-comendite alkaline granite associations. Position of KHBM is controlled by the intersection of sublatitudinal Gobi-Tien Shan rift zone with the oblique transverse fault, also controlling bimodal magmatism.

2. The massif is composed of Western and Eastern ring-like bodies; it has sharp crossing magmatic contacts with the island arc sequence and tectonic fault boundaries with bimodal sequence. The internal ring structure is characteristic of the western body and is specified by location of ring dykes and roof sagging of the hosting island arc sequence. The sequence of formation of the massif rocks is as follows (from early to late):

(1) light-grey to pink alkaline coarse-, medium-grained granites of the main intrusive phase with accessory elpidite, composing the larger part of the Western and part of the Eastern bodies; (2) dykes of coarse-grained ekerites, bedded alkalinegranitoid rocks, composed of alternating ekerites and pegmatites, pegmatoid alkaline granites, pegmatites, which are developed in the Western body, emphasizing its ring structure; (3) lilac small- to medium-grained alkaline (commpnly aegirine or arfvedsonite aegirine, often miarolitic granites with accessory zircon, making the larger part of the Eatern body; (4) dykes of micro-grained to glassy dark-blue or dark green pantellerites, (5) dykes of alkaline porphyry-like small-grained leucogranites widespread in the northern part of the Eastern body; (6) dykes of micro-grained red granosyenites and monzonites forming the system of extensive dyke bodies of sublatitunal strike in the Western body; (7) rare carbonate and quartz-carbonate veins.

From preliminary gravimetric data the massif looks as a flattened body (laccolith), its basement dipping in a step-wise manner towards north-west. Reliable geochronological data was obtained for the rocks of the massif and its framing: for

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7	Size of fraction (mkm) and	Weight	Conte	ent, /o			sotope ratio	S		Rho		Age (Ma	•
	characteristic	(mg)	Pb 6	(5 U <sup>2(</sup>	$^{6}\mathrm{Pb}/^{204}\mathrm{Pt}$	<sup>207</sup> Pb/ <sup>206</sup> Pb	a 208 Pb/206 Pt	$a^{207} Pb/^{235} U$	<sup>206</sup> Pb/ <sup>238</sup> U	20	$^7Pb/^{235}U$	<sup>206</sup> Pb/ <sup>238</sup> U	<sup>207</sup> Pb/ <sup>206</sup> PI
1		)			Bioti	te granite (s	ample XB-4	448)					
	-150+100	0.74	11.9 2	41 4.	22	0.0524±1	0.2407±1	$0.2861\pm 6$	0.0396±1	0.59 2:	56±1	250±1	302±4
	-85+60	0.48	12.2 2	34 6	37	0.0522±1	0.2275±1	0.3189±10	0.0443±1	0.50 23	81±1	279±1	296±6
	-150+100, A 20%	0.30	12.5 2	08 2	69	$0.0519\pm3$	0.2477±1	0.3277±22	0.0458±1	0.25 28	88±2	288±1	282±15
	-150+100, A 50%	0.72	4.23 7	73.8 4	03	0.0526±3	0.2740±1	0.3321±19	0.0458±1	0.57 29	91±2	288±1	313±11
	-150+100, НРФ	*,	U/Pb=	19.3 1	792	0.0521±1	0.2273±1	0.3309±7	0.0460±1	0.72 2	90±1	290±1	291±3
٦			Ra	re ear	th alkaly	ne pegmatit	e 2 <sup>nd</sup> phase (	sample XB-	1745)				
	1 grain	0.64	9.54 2	9 013	05	0.0521±1	0.0337±1	0.3337±8	0.0464±1	0.57 29	92±1	292±1	294±4
	3 grains	0.92	9.97 2	211 6	08	0.0520±1	0.0191±1	0.3330±8	0.0465±1	0.52 29	92±1	293±1	285±5
1					Trac	hydacite (s	umple XE-18	802)					
	>100, A 20 %, 10 grains	*	U/Pb=	16.94	60	0.0533±3	0.2362±1	0.3865±26	0.0526±2	0.65 3.	32±2	331±2	341±12
	>100, НРФ Зч.	*,	U/Pb=	17.8 2	362	$0.0529\pm1$	0.1787±1	0.3836±8	0.0526±1	0.62 3.	30±1	330±1	326±4
0	>100, НРФ 2ч.	*,	U/Pb=	17.5 1	301	0.0533±1	0.1868±1	0.3854±12	0.0525±1	0.53 3.	31±1	330±1	341±5
1					Alkaline	granite <b>Г</b> И	<u>Ф (sample X</u>	(B-1807)					
1	-150+100, 9grains	*,	U/Pb=	15.1 1	47	0.0518±4	0.4283±1	0.2671±24	0.0374±1	0.42 2	40±2	237±1	276±19
0	>150	0.65	23.4 4	9 861	27	0.0521±1	0.1834±1	0.2915±6	0.0406±1	0.69 20	50±1	256±1	290±3
(m)	>150, A 20%	0.17	34.2 6	526 6	53	0.0521±1	0.2965±1	0.3179±8	0.0443±1	0.54 23	$80{\pm}1$	279±1	289±5
4	-150+100, НРФ	*,	U/Pb=	20.2 4.	29	0.0520±3	$0.1074\pm1$	0.3303±24	0.0461±2	0.44 2	90±2	290±1	286±15

alkaline granites MIP, the ages were measured by U-Pb method after zircon  $(290\pm1 \text{ Ma})$ , Ar-Ar method after amphibole and polylithionite  $(283\pm4 \text{ Ma} \text{ and } 285\pm7 \text{ Ma})$ , Rb-Sr isochron method  $(287\pm3 \text{ Ma})$ , close to 290 Ma. This age is approached by the age after zircon of red biotite granites  $(290\pm1 \text{ Ma})$  and Rb-Sr age of bimodal sequence of the southern framing of the massif (table 1). The earlier magmatic formations include rocks of the island arc differentiated complex in the massif framing and in the roof sagging, which formed about 330 Ma ago. Available geochronological data on hosting rocks agree well with the age determinations obtained from floristic prints [2]: island arc sequence is dated as Middle-Late Carboniferous and bimodal sequence as Early Permian.

3. The geodynamic model of KHBM formation assumes collision of the Hercynian continent with the hot spot of paleo-ocean as two scenarios. In the first version after collision of the edge of the North-Asian paleo-continent with the hot spot of the mantle the hot spot of plume reworked the subducted lithosphere, having formed something like asthenosphere window, which was the source of rifting magmatism and the KHBM proper. In the second variant the hot and emerging mantle of plume smoothed down the subducted plate, stopped the island arc magmatism, and possibly created local convective system in the asthenosphere of the mantle wedge, which contributed to formation of the sources of rifting magmatism.

4. A huge volume of alkaline-granite plutonic and volcanic rocks of KHBM and its ring structure are conditioned by the caldera mechanism of introduction and evolition of its magmas. The initial basite magma had to form the volume, exceeding the volume of alkaline granitoids and their volcanogenic analogs, which is commensurable with the volume of magmas of moderately sizable trap provinces.

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# ENRICHED MANTLE RESERVOIRS ARE THE SOURCE OFALKALINE MAGMATISM

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Alkaline magmatism has occurred since 2.5–2.7 Ga and its abundance has continuously increased throughout the Earth's history. Alkaline rocks appeared on the Earth with changes in the geodynamic regime of our planet, i.e., when plume tectonics was supplemented by plate tectonics. Globalscale development of plate tectonics at the Archean–Proterozoic boundary initiated subduction of already significantly oxidized oceanic crust enriched in volatiles and largescale mantle metasomatism caused the formation of enriched reservoirs as sources of alkaline and carbonatite magmatism. Study of metasomatized mantle material showed the occurrence of traces of primary carbonatite melts, which are strongly enriched in rare elements, according to ionmicroprobe analyses. The results obtained allowed us to propose a new two-stage genetic model for Carich carbonatites including (1) metasomatic wehrlitized mantle with formation of carbonaterich melts or three immiscible liquids (at high alkali contents), i.e., silicate, carbonatitic, and sulfide (at high sulfur activity).

## **INTRODUCTION**

The appearance of alkaline magmatism in the Earth's history is related to the initiation of plate tectonics and, as a consequence, of largescale mantle metasomatism, which led to the formation of enriched mantle reservoirs. Ionmicroprobe studies of products of carbonate metasomatism in mantle material from some localities demonstrated that the metasomatizing mantle fluids had very high rareelement concentrations, which are comparable to those in the Lovozero giant deposit and some carbonatites. The results allowed us to propose a new two-stage genetic model for Carich carbonaties and related deposits. According to current data of seismic tomography and heat flow, the metasomatizing fluids enriched in rare elements were probably derived from plumes ascending from anomalous mantle existing in the Earth's interior.

## **ENRICHED RESERVOIRS IN THE EARTH'S HISTORY**

Alkaline igneous rock associations are the most productive for various valuable elements. Modern data on geochemistry, isotopy, and seismic tomography show that alkaline magmas were generated from enriched mantle reservoirs. Giant complex deposits of rare metals (Nb, Ta, Zr, REE, Sr, Ba), radioactive

elements, phosphorus, and aluminum are concentrated in alkaline rocks. Almost 100% of world reserves of magmatic Sr and >85% of Nb, Zr, and Ce are related to alkaline igneous rocks and carbonatites. Demand for these elements continuously increases, particularly in industrial coun tries; thus, alkaline rocks can be regarded as materials of the future.

The alkaline magmatism is most typical of stable regions (platforms) and is controlled by rift structures and zones with abruptly decreasing thickness of the continental lithosphere. Note, however, that alkaline rocks originate in almost all geodynamic settings, including oceanic ones. Evolution of the magmatic regimes of the Earth is primarily determined by global differentiation of the mantle, as the thickest layer in the Earth's interior.

The postaccretionary history of the Earth comprised multiple episodes of mantle melting resulting in the formation of continental and oceanic crust. During this process, the mantle gradually lost CaO, Al2 O3 ,TiO2 , rare lithophile elements and transformed into strongly depleted substrate, which could produce only highly magnesian magmas depleted in rare elements. However, such evolution is not typical in the Earth's history. On the contrary, many data show enrichment of mantle magmas (basalts and picrites) in Ti, LREE, and incompatible elements [1].

Isotopic studies show the existence of mantle reservoirs significantly enriched in rare elements, which could not be possible in the model of continuous global melting of mantle in the postaccretionary Earth evolution. The ideas developed in the last decades suggested active mantle–crust interaction [2] with largescale exchange of materials and formation of the mantle reservoirs enriched in rare elements. The appearance of alkaline magmatism with extreme concentrations of 4 KOGARKO rare elements is undoubtedly related to the generation of such zones in the mantle.



Fig. 1. Distribution of intensity of continental alkaline magmatism with time.

The analysis of literature and our data [3] shows that alkaline magmatism appeared in the Earth's history at about 2.5–2.7 Ga (Fig. 1), and alkaline rocks older than 2.7 Ga are extremely rare. The figure shows continuously increasing

activity (abundance and volume) of alkaline magmatism with the Earth's evolution. For example, the total volume of young platophonolites in Kenya exceeds 50 000 km 3 [4], which is much larger than the volume of alkaline rocks of all other geologic epochs. The increasing scale of alkaline magmatism with time was also noted by Lazarenkov [5]. A similar tendency is characteristic of the evolution of oceanic magmatism. The database for igneous rocks of oceanic islands and seamounts of the Atlantic Ocean, primarily of alkaline composition, helped us to reveal that the intensity of intraplate magmatism significantly grows with time. The appearance of alkaline rocks at the Archean–Proterozoic boundary coincides with several global geologic events, and this boundary became an important benchmark in the Earth's history.

Most researchers believe that the geodynamic regime of our planet changed at this boundary, and plume tectonics was supplemented by plate tectonics [6].

The oxygen atmosphere of the Earth originated particularly at this boundary [7], generally due to hydrogen dissipation to cosmic space and the activity of living organisms.

At the earlier stages, the Earth's atmosphere consisted of ëç4 ,H2 , and ëé2 [8].At the Archean–Proterozoic boundary, the continental crust extensively grew and stable, cratonic regimes were developed.

The oxygen bearing atmosphere caused the formation of oxidized marine sediments (jaspilites, carbonates) and the oxidation of paleosoils [7]. Note that giant sedimentary pitchblende and pyrite deposits the Archean of the Witwatersrand type have no analogues in the later epochs, because the minerals composing their ores are unstable in an oxygen atmosphere [7, 9]. Global development of plate tectonics at the Archean-Proterozoic boundary initiated the subduction of already significantly oxidized oceanic crust with elevated contents of volatiles (generally H2 O and CO2), which, in turn, resulted in the involvement of these components in mantle cycles. Active mantle-crust interaction increased the concentrations of volatile components in the mantle, which had lost these components during hightemperature accretion at the earlier stages of Earth evolution. Intensive studies of the mantle fluid regime revealed mantle heterogeneity with respect to oxidation-reduction potential. According to recent data [11], oxygen fugacity in the mantle substrate varies within five-six logarithmic around the quartz-fayalite-magnetite (OFM) units buffer system. However, despite the significant redox heterogeneity of the lithospheric and astenospheric mantle of the Earth, there are some regularities. Most researchers accept the idea that the oldest Archean lithospheric mantle (a diamond source) had low f O2 values. Diamonds are much older than host kimberlites [12-14], >3 Ga in some cases. They contain inclusions of Nifree metallic iron [15], moissanite, magnesiowuestite [16], and methane [17], which indicates strongly reduced conditions in the old Archean mantle. These data are consistent with data on the V contents in Archean tholeiites [18] and the significant growth of the Ti/V ratio during differentiation caused by fractionation of pyroxenes concentrating reduced trivalent V. Our recent studies revealed the occurrence of Cubearing Fe–Ni alloys in the mantle, i.e., the phases composing the Earth's core [19]. Their stability requires very low oxygen fugacities, by three logarithmic units lower than the wuestite-iron buffer [20]. Such conditions probably existed in early geological history at about 4.6 Ga during the core formation. Arculus and Delano [10] experimentally determined the oxidationreduction regime for mantle xenoliths using electrochemical cells and demonstrated that low  $f \neq 2$  values for some rocks and megacrysts in basalts are also close to the wuestite-iron buffer. On the contrary, Bullhaus [21] recently revealed the more oxidized state of most mantle peridotites (from -2 to 0 logarithmic units with respect to QFM), particularly the mantle rocks strongly modified by metasomatic processes (2 and higher logarithmic units relative to QFM). Islandarc andesites and basalts (+3 logarithmic units relative to QFM) closely related to the subduction of oceanic crust and the alkaline rock of oceanic island (+2 logarithmic units relative to QFM) are the most oxidized igneous rocks [21]. Our data demonstrate a significant degree of oxidation of the highly alkaline lavas of Trindad Island [22] with the oxygen fugacity exceeding  $f \neq 2$  of the QFM buffer by three orders of magnitude. Study of the oxidation degree of volcanic gases [8] determined f é2 and CO/CO2 at 0.01 and 0.03 respectively, which indicates high oxygen fugacities during magma formation in the modern mantle. All these data emphasize the significant heterogeneity of the mantle with respect to redox conditions, verify the existence of relicts of older reduced mantle domains within more oxidized mantle material, and may indicate global processes of mantle oxidation in the Earth's history.

Mantle oxidation might be generally caused by iron disproportionation into the Earth's core [23]. According to Galimov [23], about 95% of the iron core was formed over about 100 m.y. of the earliest evolution of the Earth. Further growth of the core by consumption of FeO from the mantle resulted in oxygen generation over the next 100-500 m.y. of the Earth's history in the amount sufficient for mantle oxidation to QFM buffer values. Some researchers believe and our calculations verify their ideas that subduction accompanied by influx of oxidized H2 O - fluid into mantle cycles could also increase oxygen fugacity in mantle fluids. This process is distinctly pronounced in zones of global mantle metasomatism. A significant increase of volatile influx into the mantle due to the activation of subduction at the beginning of the Proterozoic stimulated largescale mantle metasomatism, because the oxidized fluid phase (O +CO2) is capable of transporting great amounts of silicate material [24]. According to [25] and our calculations, annual addition of H2 O and CO2 into mantle due to the subduction of oceanic crust with hydrous minerals and carbonates is 10 14 g mol/y and 6,10 12 g mol/y, respectively. Note that carbonatites containing 20% C compose no more than 6–27% in alkaline igneous associations. Relation of carbonatites with the oceanic crust recycling is verified by the wide scatter of  $\Box$ ë13 values for carbonatite complexes worldwide [26]. Thus, changes in the

geodynamic regime of the Earth and the strong activation of the mantle-crust interaction led to the global metasomatic migration of material and the formation of zones enriched in rare elements in the mantle substrate, a source of alkaline magmas. Coincidence of these events could not be accidental and probably indicates a basic relation of these processes. Alkaline igneous rocks appeared at the end of Archean as a result of the above events and were not abundant during the whole Proterozoic, at least until the end of the Vendian (Fig. 1). Some small peak occurred only at the end of the Middle Proterozoic, between 1300 and 1000 Ma. Instead, the activity of alkaline magmatism continuously increased from 600 Ma, with peaks at about 550, 400, and 250 Ma, and particularly strongly from 200-180 Ma, reaching the maximum at about 40 Ma in the Eocene. A similar increase in the abundance of alkaline rocks is also typical of individual continents, i.e., North and South America, Eastern Europe, Siberia, and Africa. The role of kimberlites, lamproites, and carbonatites also strongly increased since the end of the Ordovician with a maximum in the Late Cretaceous and Early Paleogene. It should be mentioned, however, that the primary distribution could be disturbed because of the significant consumption of old alkaline rocks of oceanic islands and seamounts in subduction zones. It primarily concerns the Mesozoic igneous rocks of the eastern Pacific Ocean. Thus, continental alkaline massifs are probably best preserved. If the subducted rocks could be taken into account, the temporal distribution of alkaline rocks would be less contrasting, but woold not principally change. Hence, this distinctly heterogeneous distribution with a strong increase of alkaline rock abundance in the Panerozoic requires some specific interpretation. Most alkaline rock massifs were emplaced in intraplate environments due to mantle plume activity. Only some of them are located at spreading axes above hot spots, e.g., the Azores archipelago in the Atlantic Ocean. According to seismic tomography data, there is a close relationship between descending slabs of the subducted oceanic lithosphere and nearby ascending mantle plume jets at the deep mantle level near the mantle-core boundary. As a result, even the volcanics emplaced within the lithospheric plates show isotopic and geochemical signatures of recycled crustal material, which was revealed, for example, for Hawaiian volcanoes [27]. One more suggestion should also be considered. It is possible that periods of the whole-mantle convection and two-layer convection (separately for the lower and upper mantles) alternated in the Earth's history. The epochs two-layer convection dominated during of assembly of supercontinents (Pangeas), while the whole-mantle convection operated during the supercontinental epochs and breakup initiation. The alkaline rocks appeared with the formation of the first (sufficiently well documented) Pangea0 supercontinent. According to Ringwood's ideas [28], the slabs subducted during periods of twolayer convection could be retained at the upper and lower mantle boundary, where they could be entrapped by plumes. There is a dependence between mantle viscosity and convection structure, and probability of the whole-mantle convection increases with increasing viscosity caused by gradual cooling of the Earth.

The whole-mantle convection is probably operating now, while the two-layer convection was typical of the Late Archean, and plumes were initiated at the base of the transitional zone between the lower and upper mantles at depths of about 660–670 km. Significant changes in deep geodynamics are suggested at the end of the Neoproterozoic. They corresponded to the initiation of the Rhodinia breakup (early Late Riphean) and may have been expressed in dominant whole-mantle convection. The latter may have further developed later, at 400–250 and, particularly, 200–180 Ma, during the epoch of the initiation of Wegener's Pangea breakup. According to seismic tomography data, many slabs of the subducted oceanic lithosphere (eastern Pacific periphery, Mongolia–Okhotsk and South Anui paleooceans [29]) began to reach at that time the D" layer at the mantle base and to stimulate plume ascent from this depth. Thus, the fluid-rich material from the mantle base (1000–1200 km), rather than from the lower/upper mantle boundary (670 km), could ascend to the surface.

The demonstrated increase in the activity and abundance of alkaline magmatism in the Earth's history may have been caused by the gradual involvement of progressively deeper mantle horizons into the wholemantle convection and advection (plumes), progressively increasing material exchange between the Earth's surface and crust and its deeper interior due to recycling, and progressively increasing filtration of metasomatizing fluid flows through the mantle.

# GLOBAL MANTLE METASOMATISM AND GENESIS OF

# **ENRICHED MANTLE RESERVOIRS**

Large-scale interstitial pellicular melting by the mechanism suggested by Ringwood [28] may have occurred as a geochemical consequence of the global degassing of the subducted material and the release of oxidized fluids (H2 O and CO2). We imply that the melts enriched in H2 O and CO2, generated at lowmelting degree and capable of transporting great amounts of incompatible rare elements and alkalis, presented the main metasomatizing agent in the mantle. Experimental data show [30] that aqueous fluid can contain at high pressure a few tens of percent of alkalirich silicates.

It was demonstrated earlier (e.g., [31]) that metasomatic processes play a highly important role in the generation of alkaline magmas. Strong enrichment of alkaline magmas in rare lithophile elements and depletion with respect to radiogenic isotopes can be accounted for only by the metasomatic influx of rare lithophile elements into zones of alkaline magma generation. This idea can be verified, for example, by our isotopic data [32] on the worlds largest alkaline complexes on the Kola Peninsula (Khibina and Lovozero massifs) accompanied by giant raremetal deposits. These data show strong depletion of the initial mantle source of alkaline magmas with respect to rare elements. In the  $\Box$ Nd  $-\Box$ Sr

diagram, the rocks and ores of the Khibina and Lovozero massifs fall within the field of depleted compositions [31, 32]. This paradox can only be explained by the very rapid addition of metasomatizing material strongly enriched in incompatible rare elements to the melted mantle domain when there was not enough time for the accumulation of radiogenic isotopes (87 Sr and 143 Nd), and the Sr and Nd compositions. initial isotopic ratios still corresponded to the depleted Recent detailed studies of mantle material from various regions [33, 34] and our data [3, 37] demonstrated the significant role of metasomatism in the geochemical history of the mantle. Crystallization of amphiboles, phlogopite, apatite, primary carbonates, minerals of the lindsleyite-mathiasite series, and other minerals concentrating rare elements is related to metasomatic reactions in mantle substrate, most likely, to low-degree melts. As was shown experimentally by Fransis [35], partial melting of strongly amphibolized metasomatized mantle can generate alkaline melts. Highalkali glasses in metasomatized mantle xenoliths were also described by Edgar [36]. When studying strongly metasomatized and carbonatized nodules from volcanics of the Montana Clara Islands (Canary Islands) and Fernando de Noronha (Brazil), we found glasses (in association with primary carbonates) corresponding in composition to phonolites and trachites oversaturated in alkalis. Thus, alkaline magmas could have been generated during the partial melting of metasomatized mantle.

Recently, we actively studied the processes of mantle carbonate metasomatism in various regions. In the oceanic mantle of some Atlantic islands [3, 37], we revealed primary carbonates developed metasomatically by the following reactions:

2MgSiO3 + CaMg(CO3) = 2Mg2SiO4 + CaMgSi2O6 + 2CO2,

 $3CaMg(CO \square)2 + CaMgSi2 O6 = 4CaCO3 + 2Mg2 SiO4 + CO2$ .

Metasomatic clinopyroxenes and olivines are products of these reactions. Using microprobe, we determined CAMECA ion concentrations of a metasomatic pyroxenes in oceanic and continental carbonatized mantle (Canary Islands, Fernando de Noronha, and East Antarctica). With the partition coefficients of rare elements in pyroxene-carbonate melt equilibria [38, 39], we calculated the concentrations of rare and rare earth elements in the metasomatizing mantle carbonate melts. Note that the average composition of the Lovozero massif accompanied by giant rare metal deposits fall within the interval of concentrations calculated for metasomatizing mantle fluids (Fig. 2). These data are consistent with the compositions of carbonate melts in the metasomatized mantle of some other regions [39]. All melts are enriched to a various degree in LREE and incompatible elements. This REE distribution (with LREE enrichment) indicates that metasomatic meltfluids were generated in crystal melt equilibria, because the LREE partition coefficients in most mantle rockforming minerals, particularly in garnets, are notably lower than the HREE partition coefficients. Most calculated compositions of metasomatic meltfluids show minima of HFSE (Ti, Zr, Nb) (Fig. 2). This is consistent with experimental data [38] on the low solubility of these elements in hightemperature carbonatite liquids. However, HFSE maxima (Ti, Zr) were revealed in some cases (Fig. 2), which could indicate alkaline carbonate– silicate compositions of metasomatizing agent, because HFSE solubility strongly increases in highly alkaline silicate melts [39]. Despite some differences in the compositions of metasomatizing melts, all of them are enriched in incompatible rare elements and LREE. Universal enrichment of mantle fluids in rare elements is a basic geochemical feature of the Earth's mantle.



Fig. 2. Spidergram for the mantle metasomatizing meltfluids (calculated by ion microprobe analyses of metasomatic clinopy roxenes of mantle rocks from Fernando de Noronha Island, East Antarctica, and Canary archipelago).

Stars—average composition of rocks of the Lovozero alkaline massif, asterisks—average composition of carbonatites.

The ascending mantle plumes induce low-degree melting of the surrounding mantle (particularly at the base of the plumes), which produces carbonate melts enriched in rare elements and volatile components. These melts have very low viscosities, rapidly penetrate to upper horizons, and metasomatically replace mantle peridotites (stage 1). Further ascent of the diapir causes the partial melting of metasomatized mantle and the generation of alkaline and carbonatite magmas (stage 2). The proposed model of the formation of enriched reservoirs is generally consistent with the geochemistry of metasomatizing meltfluids, but requires very low melting degrees of mantle material. If the mantle source of giant rare-metal deposits of the Kola Peninsula was strongly depleted in rare elements [32], the mass of the depleted material producing metasomatizing fluids should be huge. It is more logical to suggest that the plume itself, significantly enriched in rare elements as compared even to primitive mantle, was the source of rare elements for the giant alkaline intrusions. Let us consider these problems in the next section.

# **DEPTH OF FORMATION OF ENRICHED MANTLE RESERVOIRS**

According to modern concepts, two global reservoirs, upper and lower, can be distinguished in the Earth's mantle. However, recent data of seismic tomography [40] indicate the occurrence of new structural heterogeneity in the lower mantle below about 1000 km until D" layer at 2300 km. Pwave velocities significantly decrease in this interval [40]. These data suggest the existence of three dynamic regimes in the Earth's mantle. The above heterogeneity continues over about two thirds of the transitional lower mantle–core layer (Fig. 3) and comprises 20–30% of the whole Earth's mantle. The heterogeneity is related to elevated viscosity and density of the material at the base of the lower mantle.



Fig. 3. Structural scheme of the Earth's upper and lower mantle (after [40]).

The size and shape of this region (megaplume) may indicate the chemical heterogeneity of the lower mantle. If this seismic heterogeneity had thermal origin, it could not exist for a long time because of tremendous heat flux from the core. Density in the anomalous zone probably increases due to the decomposition of silicates and the formation of high-density oxides (magnesiowuestite, stishovite) at pressures >650 kbar. Crystallization of magnesiowuestite enriched in Fe is

another mechanism of the formation of heavy mantle material. The suggestion of the existence of an anomalous layer in the lower mantle is consistent with modeling of the thermal regime of the Earth. The heat flux of the whole Earth is estimated at 44 TW; the heat flux of the crust is 6 TM due to high concentrations of radioactive elements [40], and the remaining 38 TW should be supplied by the mantle and core. Even if the whole Earth was chondritic in composition, its heat flux would be only 31 TW [41]. It is established now that 25–90% of mantle substrate (MORB sources) is depleted in rare and radioactive elements, and, thus, heat flux of the Earth should be lower than 31 TW [41], but it is actually much higher (44 TW). Hence it follows that a layer producing huge heat amount corresponding to U concentration of 25.6 ppm should exist in the mantle [41].

The origin of this zone enriched in rare and radioactive elements in the mantle is not clear. A proposed model suggests the longterm development of a reservoir isolated since the early stages of the Earth's evolution and originating either during the evolution of the global magmatic ocean or by the recycling of Archean crust enriched in mafic components and incompatible rare elements. Elevated contents of radioactive elements should cause active heat generation and very high temperatures in the D" zone. Under the effect of heat flow from the core, hot and dense material of the lower mantle could rise and produce superplumes, e.g., African or East Pacific. We also considered the possi bility of magma generation for the Kola alkaline igneous province from an anomalous zone in the lower mantle. According to [42], the total volume of alkaline rocks of the Khibina and Lovozero massifs is about 11000 km 3. Using the average contents of rare elements in these rocks (e.g., Zr and Th), we can estimate the volume of the primitive or depleted mantle as sources of these rocks. According to our isotopic studies [31, 32], the giant alkaline intrusions of the Kola Peninsula were generated from a mantle source depleted in rare elements. The further calculation showed that the volume of mantle material for the formation of the Khibina and Lovozero massifs should be about 3.7106 km 3 for primitive mantle and 3.8107 km 3 for depleted mantle. These volumes are not realistic and are comparable or even exceed the volumes of the largest mantle plumes. For example, the volume of the Siberian superplume is about 7, 10 6 km 3 [43]. Moreover, there is no geological evidence of the relation of the Kola alkaline province with any large plume. Thus, our calculations may indicate a much more enriched source of alkaline magmas for the Kola alkaline province. They could be derived from an anomalous layer of the lower mantle. General regularities in fractionation of rare and radio active elements in mantle reservoirs allow us to estimate the distribution of incompatible elements in the anomalous layer. As was mentioned above, U content in the anomalous zone is 25.6 ppm [40]. If all incompatible elements are proportionally fractionated in mantle processes, their contents should be (like the U content) by about two orders of magnitude higher than in the primitive mantle. These estimates are semiguantitative. However, such concentrations at 15% melting degree of anomalous lower mantle domains (this melting degree is

normally accepted for superplumes producing flood basalts) and the addition of 5-30% of metasomatizing fluid could provide for the generation of alkaline magmas with concentrations of rare elements close to their average concentrations in the ultrabasic alkaline magmas parental for the ultrabasic alkaline rock associations in Polar Siberia and the Kola Peninsula. The calculations suggested Nb contents of 0.7 ppm in the primitive mantle, 70 ppm in the anomalous mantle, 490 ppm (calculated value) in a metasomatizing melt, and 131 ppm in parental magma [42] and assumed 5% melting degree of a metasomatically modified substrate. At these Nb concentrations, the calculated contribution of metasomatic material is 1.3%. At lower melting degree of 1%, the amount of the metasomatizing melt-fluid will be very low (about 0.013%). Our studies, together with literature data [44] demonstrate that the concentrations of metasomatic minerals (amphibole, mica, apatite, etc.) in the mantle are low (a few fractions of percent). Metasomatic processes in the mantle frequently do not cause the specific minerals. formation The addition of rare of elements bv metasomatizing melts can be low and only increases their concentrations in rockforming minerals.

Therefore, the evolution of alkaline magmatism in the Earth's history was determined by the intensity of mantle metasomatism causing the formation of enriched mantle reservoirs as sources of alkaline magmas. The global development of metasomatism in the mantle was related to the subduction of oceanic crust initiated at the Archean–Proterozoic boundary. The genesis of alkaline and carbonatitic magmas and related giant raremetal deposits may have been caused by mantle metasomatism induced by the partial melting of the anomalous substrate in the lower mantle. Such anomalous zones are revealed in recent studies using seismic tomography [40, 41].

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# UDK 552.48 + 552.231.6 Garnets of anomalous oxygen isotope composition in diamondiferous xenoliths (Nyurbinskaya pipe, Yakutia)

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The oxygen isotopic data suggest that many eclogite xenoliths, especially the group C eclogites, are representatives of subducted crust. The presence and preservation of coesite in eclogites from Udachnaya and other kimberlite pipes of Yakutia gives a strong proof of the involvement of oceanic crust in subduction by the lithosphere formation. Presented in this paper data are interesting considering their implication to this problem at least in three aspects: 1) first of all, because of their wide variations in composition of studied samples, 2) in connection with the unusual peculiarities of kimberlites of this pipe, and 3) as samples containing diamonds.

Oxygen isotope data were obtained from more than 125 clean garnet mineral separates of xenoliths from the Nurbinskaya pipe by laser fluorination techniques. Garnets were separated from the different types of diamondiferous xenoliths including eclogites, garnet websterites (pyroxenites), and peridotites. The majority of the peridotites garnet samples lie within the range of average mantle values  $(5.5\pm0.45 \ \%)$ , except one with  $\delta^{1\Box}$  value – 6.57%. Garnets from xenoliths of suggested websteritic paragenesis in most cases have  $\delta^{1\Box}$  values above 6.0‰ and only in two samples oxygen isotope estimations were 5.89 and 5.98‰. It should be pointed that in two samples of this paragenesis garnets have  $\delta^{1\Box}$  value 7.30 and even 8.59 ‰. The restricted range of  $\delta^{18}$ O garnets from eclogites is 4.73 to 9.65 ‰ and lie predominantly in the range reported for eclogite xenoliths from Siberian platform (between 2.8 and 8.0 ‰). We not find any marked difference in oxygen isotope compositions of garnets from different varieties of eclogite xenoliths but about 2/3 of investigated samples have  $\delta^{1\Box}$  values above 6.0‰.

It is possible to speculate from these evidences that many eclogites and part of garnet websterites (pyroxenites) are products of subducted oceanic crust or experienced a strong metasomatic influence. There is ample evidence that latter statement is more confident that is confirmed by wide distribution of zoned garnet in investigated parcel of samples and intensive development of phlogopite and other metasomatic minerals.

KEY WORDS: xenoliths, eclogites, pyroxenites. diamonds, oxygen isotopes, subduction

## **INTRODUCTION**

As xenoliths in kimberlites are potential source rocks for diamonds and representatives of deep mantle levels, and therefore provide important information about the mantle petrology and conditions of diamond formation, they have been the focus of numerous conventional oxygen isotope studies. Data on oxygen fugacity of southern African eclogites predominate and include studies on Roberts Victor [18, 23, 31], Bellsbank [6, 30], Orapa [10] and Southern Africa in general [39]. Oxygen isotope data for mantle peridotite xenoliths from Southern African kimberlites have been reported by Kyser et al. [26]. New laser-fluorination  $\delta^{1\square}$  data for constituent minerals from more than 100 xenoliths in South African and Siberian kimberlites have been reported in Mattey et al. [29] and Jacob et al. [22]. Analyses of eclogites and peridotites or their minerals from Yakutian kimberlites are much less abundant and were done predominantly for the xenoliths from the Udachnaya and Obnazhennaya pipes [41, 42, 53, 57, 58].

It is well known that whereas peridotite and eclogite xenoliths occur in varying proportions in kimberlites, diamondiferous eclogite xenoliths are significantly more abundant world wide than peridotite xenoliths with diamonds. In contrast to garnets from most mantle peridotites, which have  $\delta^{1\square}$  values in the very small range + 5.36±0.18 [27], many mantle eclogite xenoliths have  $\delta^{1\Box}$ values that are significantly outside of the range accepted as normal for mantle rocks. The  $\delta^{1\square}$  of garnets in mantle eclogites show wide variation + 2.3 ‰ to + 9.2‰ [11, 29]. By analogy with a section rocks from ophiolite sequences [19], eclogites (or their constituent garnets and clinopyroxenes) with  $\delta^{I\square}$  values above those of "common mantle" are thought to represent subducted ocean basalts that have been altered by interaction with sea water at temperatures below about 350 °C, whereas eclogites with lower  $\delta^{1\square}$  values represent matic rocks that have interacted with higher temperature hydrothermal fluids [28]. Taylor et al. [57] have suggested that similar mixing of subducted oceanic crust with more refractory mantle material was an important process in the evolution of the eclogite suite xenoliths from the barren kimberlite pipe Obnazhennaya (Siberia).

Most workers interpret high and low oxygen isotope ratios of mantle eclogite xenoliths as the result of subduction and prograde metamorphism of oceanic lithosphere that has undergone oxygen isotope exchange with sea water on or near the sea floor [23, 28]. But discussion on possible reason of wide variations of  $\delta^{1\square}$ values in eclogite xenoliths is still continues. From this point of view any new data especially if they were provided for the representative collection of mantle xenoliths could have important implication for the deciphering of mantle processes and explanation of possible variation in oxygen isotope composition of mantle substrate. Presented in this paper data are interesting considering their implication to this problem at least in three aspects: 1) first of all, because of their wide variations in composition of studied samples, 2) in connection with the unusual peculiarities of kimberlites of newly discovered Nakynsky field, and 3) as samples containing diamonds. In result it allows us to place constraints on the source material and processes involved in formation of diamonds in eclogite and pyroxenite xenoliths from the Nyurbinskaya pipe. These results could have important implication to the understanding of diamonds genesis and lithosphere evolution.

# GEOLOGICAL BACKGROUND

The first deposit of diamonds of the Nakynsky kimberlite field was found in 1994. In the passed ten-year period much attention was paid to the study of geological structure, mineralogy of kimberlites and other specific features of the Nakynsky kimberlites [1, 8, 35, 50]. Nonordinary essence of the new field kimberlites is demonstrated in these works both by isotopic characteristics of kimberlite rocks, which draw them together with kimberlites of group 2, and by specific features of indicator minerals composition, garnets and chrome spinels in the first turn. Sufficient differences in composition of both the lithospheric mantle of the new Nakyn field and the low crust in this region were revealed, which point to their distinction by a set of petrographic types of rocks and their tectonic-magmatic evolution from neighboring Malo-Botuobinsky and Daldyn-Alakitsky fields [50].

Already the first evidence on studying external morphology of diamonds from kimberlites of pipe Botuobinskaya indicated that they occupy intermediate position by spectrum of morphological groups among diamonds from kimberlites of Malo-Botuobinsky and Daldyn-Alakitsky fields [8]. Increased quantity of rounded diamonds, coated diamonds or crystals in "jacket" and diamonds with cavern, channels, crystals with etching and other indications of natural dilution and resorption were revealed in diamond population of the Nakynsky field kimberlites [50]. During visual study of mineral inclusions in diamonds from this field increased content of orange garnets and other minerals, presumably of eclogite association, was indicated in them, which pointed to possibility of higher distribution of eclogitic mineral inclusions among diamonds of these pipes in comparison with diamonds from other kimberlite fields. This statement was confirmed by investigation of mineral inclusions in diamonds from the Botuobinskaya pipe [48]. These data have shown the urgency of more detailed research of diamond-bearing associations, which became the basis for fulfilling this work.

The Nakyn kimberlite field represents on the whole a tectonic block in the central part of Vilyuy-Markhinskaya zone, with 380 km<sup>2</sup> area, which is constrained by faults of north-eastern strike, along which trap dykes preceding emplacement of intruded kimberlites. By now two kimberlite pipes (Botuobinskaya and Nyurbinskaya) have been discovered within the Nakyn field, having cylindrical shapes extended in north-eastern direction, geological characteristics of which is given in [8]. Both pipes were formed in two intrusive phases: subvolcanic (the first phase) and volcanic. Both kimberlite pipes of the new field are characterized both by resemblance of their material composition and diamond content and are infilled with porphyric kimberlites and autolithic kimberlite breccias, with sharp prevalence of the latter.

Two varieties of kimberlite rocks – autolithic kimberlite breccia and porphyric kimberlite, were distinguished in pipe Nyurbinskaya. According to a number of features they fall into a category of unique ones among earlier discovered kimberlite pipes of Yakutian diamondiferous province [25]. First of all it is due to their highest productivity, specific features of morphology and quality of diamonds, petrographic and mineralogical characteristics of kimberlites, and due to composition of diamond indicator minerals. Strong alteration is characteristic feature of kimberlitic rocks, which is expressed in absence of olivine relics, intensive chloritization of phlogopite to the depth of 200 m, and practically carbonate composition of the basic matrix on upper levels [50].

Kimberlites of the Nyurbinskaya pipe differ by exclusively low content of chrome spinels, picroilmenite and perovskite. They are depleted in pyrope and enriched in almandine garnets. The latter, most probably, is conditioned by low content of abyssal xenoliths and high – metamorphic crustal rocks. As by petrochemical features of kimberlites and concentration of heavy fraction minerals, in comparison with their distribution in other pipes, Nyurbinskaya testifies similarity with kimberlite rocks of Aykhal (Alakit-Markhinsky field) and Lomonosovskaya (Zolotizky field) pipes [50].

By content of major elements kimberlites of both pipes are close to each other and sharply differ from kimberlite rocks of other pipes of Yakutian province. These are low-titanium (less than 1.0 weight % TiO<sub>2</sub>), low-ferruginous, mediummagnesia and relatively high-potassic rocks. Average chemical composition of Botuobinskaya and Nyurbinskaya kimberlites differs from composition of kimberlite rocks of Malo-Botuobinsky and Daldyn-Alakitsky region pipes by increased content of alumina and potassium oxide at small quantity of titanium oxide. The age of kimberlite pipes determined on phlogopite by Rb-Sr method constituted 366±9 Ma years [1], and on phlogopite by  ${}^{40}$ Ar/ ${}^{39}$ Ar method - 380±12 Ma [25].

# SAMPLES AND ANALYTICAL TECHNIQUES

More than 150 samples of mantle xenoliths from the kimberlite pipes Nurbinskaya, which situated in newly discovered Nakynsky field, of the Yakutian kimberlite province were studied. Modal analyses have been performed for most of the xenoliths. Major-element analyses were performed for garnets from all samples and partly for the other rock-forming minerals. All samples were classified into different varieties of eclogites, websterites and peridotites according to their petrographic features and chemistry of garnet.

Major element compositions of garnets and some other minerals in all xenoliths were determined with a Superprobe JXA-8800R electron microprobe at the ALROSA Co Ltd. (Mirny, Russia). Care was taken to analyze only those minerals from the samples that did not appear to contain inclusions or exhibit alteration. Analytical conditions included an accelerating voltage of 15 keV, a beam current of 20 nA, beam size of 5  $\mu$ m, and 50 seconds counting time for all

major elements and 100 seconds for Na, Mn, Ti, and Ni. All analyses underwent a full ZAF correction.

Fresh garnet grains were selected from over 120 samples eclogite and peridotite xenoliths collected from the Nurbinskaya kimberlite pipes. The mineral was crushed and sieved to 200 µm and picked using a binocular microscope. After picking the clean, optically gem-quality garnet grains were washed with water and acetone. A split of each sample was used for EMP analyses. Pure separates were obtained using heavy liquids, magnetic separation, and careful hand picking under a binocular microscope. Another 10 mg portion was set aside for oxygen isotope analyses by laser-fluorination. The oxygen-isotope analyses were performed on the garnet separates, approximately 1-2 mg per run, using a 32 W CO2 laser, BF5, and a dual-inlet Finnigan MAT/251 mass spectrometer at the University of Wisconsin. Replicate analyses were performed on many samples, particularly those that had initial oxygen values outside that of mantle. The general procedures are those discussed by Spicuzza et al. [54]. All values are reported with respect to V-SMOW.

# RESULTS

# Petrography and sample description

As mentioned above we have studied a representative collection of more than 150 xenoliths with visible diamonds from the Nurbinskaya pipe. Most of the xenoliths are presented by the samples of rounded shape or sometimes by the monomineralic garnet megacrysts containing ingrowths of diamonds. It should be noted that 2 and even more diamond crystals were found in some samples. It is necessary to emphasize the unique nature of this collection of xenoliths and megacrysts with diamonds, which is determined both by the fact that they were found in kimberlites of recently discovered pipe in the new kimberlite field, and by multiplicity of the samples. Though kimberlites of such known pipes as Mir and Udachnaya were developed for many years general quantity of xenolith finds with diamonds in them does not prevail 100 samples in each individual pipe as a rule. This is true at least for the xenolith populations from studied in detail kimberlite pipes, mantle associations of which are known and minutely described in literature [22, 49]. The quantity of xenoliths and garnet megacrysts containing diamonds, which are characterized in kimberlites from the pipes of South-African craton [24], is within the same number.

Leaving comprehensive characteristics of diamonds present in xenoliths we'll note only that in most of the samples they are represented by crystals of octahedral forms (Fig. 1). The following ratio of morphological diamond groups was established for the whole collection: octahedral – 65%, rounded and transitive forms – 22%, intergrowths and boart – 8%, various types of twins – 5%. Two xenoliths, comparatively similar in composition of basic mineral garnet and containing in one case octahedron with polycentric structure of facets, and in the

other case – sharp-edged octahedron, are shown in Fig. 1 a, b. It should be noted that in both cases diamonds are deeply submerged into matrix of hosting garnet but at the same time one can trace margin around them represented by chloritizated phlogopite. Similar but more pronounced margin consisting of monomineralic completely chloritizated phlogopite is distinctly seen around corroded crystals of irregular shape (Fig. 1d). Some of the studied samples contain diamonds in the quantity of up to five and more crystals. In some xenoliths, therewith, there accumulations of diamonds forming individual chains and veinlets are occur (Fig. 1c). Diamonds in some samples are presented by coated crystals which usually are of yellow-green colour. The size of diamonds varies from 0.5 mm to 4-5 mm in the long axis. It is necessary to emphasize that many diamonds contain channels of etching and other indications of corrosion (see Fig. 1d). The quantity of such crystals in the studied collection on the whole constitutes about 30%.



#### Fig. 1. **Diamondiferous xenoliths from the Nurbinskaya pipe.**

a) Sample N-6, by the garnet composition belongs to the magnesian (Group A) eclogites. Diamond is presented bv colorless smooth faced octahedron, size - 2.4x1.8 mm; b) Sample N-4, by the garnet composition belongs to the garnet websterites. Diamond is presented by colorless polycentric octahedron, size - 3.0x2.0 mm; c) Sample Nthe 17. bv garnet

composition belongs to the high-aluminian (Group C) eclogites. Diamonds are presented by  $5^{\text{th}}$  arc wise distributed crystals of octahedral form; size of crystals varies from 2 to 3 mm; d) Sample N-20, by the garnet composition belongs to the ferrian (Group A) eclogites. Diamond is presented by grey heavily corroded crystal of irregular form, size – 3.5x2.5 mm. It is obvious channels of corrosion, which cut a crystal and also is seen a rim around diamond that is presented by phlogopite, which intensely replaced by chlorite.

What is concerning the host-rock of xenoliths they look in many cases as garnetites or garnet megacrysts because of strong rock alteration, which caused only garnet presence as a fresh mineral relicts in some cases with little amount of corundum or rutile that are stable to alteration or phlogopite that have obvious late origin. No fresh relicts ortho- or clinopyroxene were met in studied samples even after their cutting and polishing. Reasoning from this, classification of the studied xenoliths has been done predominantly on the base of the garnet composition of samples.

## Chemistry of garnets from Nurbinskaya xenoliths

Garnet is a major mineral of eclogite and pyroxenite xenoliths especially garnet websterites. Chemical composition of this important constituent of eclogites largely reflects their nature. Garnets from eclogites show a wide variation in content of Mg, Fe and Ca oxides. Eclogitic garnets usually are poor in chrome content but in websterites they could contain up to 1.0 wt%  $Cr_2O_3$ . As it was noted earlier grains of garnet were selected from all the xenoliths and analyzed on the subject of determining major components, in most cases at that definition of composition of garnet central and peripheral zones was performed with the goal of revealing possible zonality.

Table 1.

		Average of	compositio	on of garn	et cluster	groups		
Sort 8	G1	G2	G3	G4	G5	G6	G7	G8
SiO <sub>2</sub>	39.26	39.73	40.05	40.48	40.23	40.62	40.52	40.72
TiO <sub>2</sub>	0.27	0.45	0.29	0.38	0.40	0.30	0.37	0.15
$Al_2O_3$	21.22	21.28	21.75	21.96	21.85	22.05	21.98	15.39
$Cr_2O_3$	0.07	0.08	0.08	0.10	0.10	0.11	0.17	9.87
MgO	9.38	10.23	12.43	14.10	9.00	15.53	20.59	19.09
CaO	5.09	8.27	5.05	10.20	16.33	3.82	3.37	6.12
MnO	0.48	0.39	0.38	0.19	0.14	0.37	0.34	0.47
FeO	23.07	18.22	18.83	10.76	11.03	16.26	10.65	7.46
Na <sub>2</sub> O	0.10	0.16	0.12	0.22	0.17	0.12	0.12	0.04
NiO	0.01	0.01	0.01	0.01	0.01	0.01	0.01	0.01
Total	98.94	98.83	99.00	98.41	99.26	99.19	98.12	99.31
		Nun	nber of sa	mples and	l relative 9	%		
Sort 8	G1	G2	G3	G4	G5	G6	G7	G8
samples	10	17	51	8	13	40	10	12
%	6.2	10.6	31.7	5.0	8.1	24.8	6.2	7.5

Average composition of garnet cluster groups of diamondiferous xenoliths from the Nyurbinskaya pipe

Based on the chemistry of garnets we have tried to classify all samples on their belonging to one or another variety of mantle xenoliths. The received data of compositions of this mineral were calculated with the help of factor analysis, which made it possible to distinguish 8 groups of garnets, clearly varying by these or those major components (Table 1). The Table contains average compositions of cluster groups and the quantity of analyzed samples and their percent content in the studied parcel. It is necessary to note that the groups of garnets distinguished by the results of cluster analysis sufficiently closely correspond to these or those varieties of garnets in specified types of mafic and ultramafic xenoliths. It should be noted that the distinguished groups correspond to xenolith varieties of mantle rocks by a number of indications. In particular group G8 completely responds to garnets of dunite-harzburgite paragenesis according to their high C<sub>2</sub>O<sub>3</sub> content, garnets of group G7 are differ by much less C<sub>2</sub>O<sub>3</sub> and low CaO content with similar high MgO and correspond to garnets of lherzolite and wehrlite or some pyroxenite paragenesis, though they are distinguished by decreased chromium content. The group of G6 garnets corresponds to this mineral from garnet websterite xenoliths, which is less magnesian in comparison with garnet of two previous groups but also has low CaO content. Not excluded that some garnets of this group could belong to Group A eclogites. The group of G5 garnets completely corresponds to the garnets from eclogite xenoliths of high-aluminous variety represented usually by grospidites, kyanite and coesite eclogites or corundumbearing samples, what is also confirmed by the finds of individual corundum grains in the samples of this group. The high CaO in garnet is a criterion for recognizing of these kinds of xenoliths (Spetsius, 2004). Garnets of 4<sup>th</sup> group are close to the previous by Mg# and differ by less content of CaO and by their chemistry could belong to different kind of Group B eclogites. Groups G1, G2 and G3, for which low chromium content, increased FeO and varying contents of calcium oxide are characteristic, correspond to garnets from the varieties of magnesian and magnesian-ferruginous eclogites. Garnets of group G6 with increased magnesia and decreased calcium content may belong both to xenoliths of magnesian eclogites (Group A) and include a group of garnets that are characteristic for garnet websterites. Garnets of groups G6 and G7 are in somewhat indefinite position, being distinguished by increased magnesia content and by low content of calcium and chromium. Similar garnets occur both among pyroxenite xenoliths and can be correlated with garnets of megacryst association. Probably it is expedient to isolate them into a group of transitive rocks. In any case the first five groups, which unite garnets with moderate Mg# and increased FeO, respond to varieties of mantle rocks of eclogite composition. More than 60% of samples in the



parcel of investigated mantle rocks with diamonds account for this type of diamondbearing xenoliths. It should be pointed that samples of for sure dunite-harzburgite association according of garnets chemistry constitute less than 10% of the whole population of diamondiferous xenoliths (see Table 1).

#### Fig. 2. Garnet composition in diamondiferous xenoliths from the Nurbinskaya pipe.

For comparison field of garnets in diamondiferous eclogites from the Udachnaya pipe is shown.

Wide variations of garnet composition of investigated xenoliths is evidenced from the triangular diagram that shows as distribution of mineral in the whole parcel and prevalence of garnets with moderate contents of all three elements (Fig. 2). It should be pointed that according to garnet composition the most diamondiferous samples from the Nyurbinskaya pipe occupy fields of groups B and C eclogites. In the whole the distribution of diamondiferous xenoliths is close to the distribution of diamondiferous eclogites from the Udachnaya pipe [see 49].

It is well known that garnets in the majority of eclogites and websterites exhibit intra- and inter-grain homogeneity [49, 57] but it is not a case for the garnets of diamond-bearing xenoliths from the Nyurbinskaya pipe. Analyses of garnets were made in different types of eclogites with the attention to possible zoning of this mineral (Table 2). We presented here only data that show a rather wide difference in Mg/(Mg+Fe) ratio from sample to sample and obvious variation in content of major elements from core to rim. It should be pointed that zoning in garnets probably could be formed as a result of at least two different processes: (1) magmatic crystallisation or subsolidus transformation and (2) zoning could be a result of metamorphic mineral transformation or influence of metasomatic events.

Table 2.

Sample	SiO <sub>2</sub>	TiO <sub>2</sub>	Al <sub>2</sub> O <sub>3</sub>	$Cr_2O_3$	FeO	MnO	MgO	CaO	Na <sub>2</sub> O	Total
N3c	41.03	0.31	22.37	0.08	20.34	030	12.29	4.06	0.13	100.61
N3r	41.53	0.34	22.58	0.07	20.01	0.31	12.47	4.01	0.10	101.11
N16c	41.76	0.32	22.82	0.04	16.22	0.38	15.91	2.88	0.12	100.06
N16r	41.53	0.36	22.62	0.07	15.73	0.36	16.35	2.91	0.12	99.69
N24c	41.72	0.29	22.85	0.10	17.07	0.36	14.87	4.01	0.12	101.03
N24r	41.87	0.31	23.00	0.07	16.88	0.33	14.54	4.00	0.11	100.77
N34c	39.64	0.23	22.47	0.04	18.24	0.45	8.68	9.54	0.20	99.04
N34r	40.41	0.25	22.55	0.04	17.58	0.49	8.75	9.80	0.19	99.56
N35c	40.69	0.42	22.79	0.02	18.53	0.27	11.18	7.16	0.18	100.98
N35r	41.16	0.53	22.63	0.03	17.14	0.44	11.71	7.53	0.20	100.92
N59c	40.48	0.26	22.62	0.09	17.17	0.21	13.21	5.16	0.17	99.16
N59r	40.58	0.32	22.41	0.07	17.34	0.32	12.61	5.34	0.13	98.79
N60c	41.52	0.49	22.63	0.05	15.23	0.39	14.40	6.13	0.15	100.59
N60r	40.71	0.53	21.92	0.06	16.93	0.40	12.27	7.26	0.16	99.84
N66c	41.17	0.20	22.41	0.19	9.46	0.14	12.31	13.71	0.08	99.52
N66r	41.44	0.49	22.31	0.21	9.50	0.23	16.21	9.25	0.11	99.52
N77c	40.42	0.40	22.68	0.07	10.42	0.16	11.24	14.20	0.14	99.55
N77r	40.76	0.40	22.83	0.01	10.31	0.17	10.67	14.63	0.13	99.74
N113c	40.01	0.39	22.36	0.01	21.62	0.34	9.88	6.16	0.19	100.61
N113r	40.19	0.46	22.31	0.02	21.24	0.33	9.84	6.49	0.22	100.79
N-150c	40.81	0.46	21.9	0.08	12.58	0.21	13.87	9.23	0.15	99.29
N-150r	41.91	0.44	22.39	0.04	12.5	0.29	14.49	8.84	0.15	101.05
N-155c	41.33	0.48	22.06	0.00	13.69	0.53	17.81	3.64	0.16	99.70
N-155r	40.53	0.45	21.87	0.01	14.2	0.56	17.14	3.39	0.15	98.30

Zoned garnet in diamondiferous xenoliths from the Nyurbinskaya pipe

Note; Abbreviations (here and further): N-pipe Nyurbinskaya; c: core; r: rim.

In the result of zonality that forms during magmatic crystallisation usually both garnet and other minerals exhibit decreasing of Mg# and increasing of iron content from core to the rim of grains. In contrast vase verse Mg# increased from core to the rim in our case and subsequently as rule increased content of CaO. The content of FeO decreased to the periphery of garnet grains in most samples. Such major elements behaviour suggests that the most probable scenario for the formation of zoning garnet was the changing of Mg, Fe, Ca and others element concentrations under influence of introducing metasomatic fluids. Last statement is obvious from the picture (Fig. 3) that demonstrates a relationship of garnet zonality with the intergranular space and boundaries between the mineral grains and cracks. It is obvious that zoning and changing in garnet composition from the core to the periphery of grains correlates with the intensity of metasomatic processes and was formed under fluids incorporation.



Fig. 3. Zoned garnets in eclogite xenolith from the Nurbinskaya pipe (Sample N-60). Notice non similar rim thickness. Analyses of core and rim see in Table 2.

Oxygen isotope composition of garnets Analyses of oxygen isotopes were obtained for 125 clean garnet mineral separates by laser fluorination techniques [59]. All corrected  $\delta^{I^{\square}}$  values of garnets from studied samples are given in Table 3. The distribution of estimations oxygen isotopes of garnets from the whole parcel of xenoliths is shown on Fig. 4 from were is obvious wide variations of  $\delta^{I^{\square}}$  values and unusually big shift of most samples to higher level from the range accepted as normal for the mantle rocks.

It should be noticed that 7 investigated garnets were taken from ultramafic xenoliths. This is obvious as from their specific color and other features of garnets and as well is proven by high chromium content of garnets (8 wt.%  $C_2O_3$  and

more) that confirming the belonging of these samples to the lherzolite-dunite suite xenoliths from the kimberlite pipes [43]. These garnets have  $\delta^{18}$ O value that varies Table 3.

G1.	d <sup>18</sup> O	G1-	d <sup>18</sup> O	G	d <sup>18</sup> O	G	d <sup>18</sup> O
Sample	corr.	Sample	corr.	Sample	corr.	Sample	corr.
N-3	6.38	N-35	6.64	N-72	6.52	N-104	8.59
N-4	6.74	N-36	6.39	N-73	6.76	N-105	6.72
N-5	6.30	N-37*	5.05	N-74	6.72	N-106	6.28
N-6	6.46	N-38	5.51	N-75	6.27	N-107	6.29
N-7	6.73	N-40	6.40	N-76	6.59	N-109*	5.57
N-8	6.35	N-42	6.82	N-77	5.39	N-110	5.98
N-9	6.48	N-43	9.65	N-39	6.56	N-111	5.89
N-10	7.08	N-44	6.52	N-78	6.71	N-112	6.59
N-11	6.63	N-45	6.31	N-79	7.17	N-113	6.78
N-12	6.51	N-47*	6.03	N-80	6.63	N-117	6.79
N-13	6.65	N-48	6.40	N-81	6.47	N-118	6.21
N-14	5.51	N-49	6.70	N-84	4.78	N-119	6.43
N-15	5.40	N-50	6.55	N-85	6.56	N-120	6.82
N-16	6.24	N-51	7.24	N-86*	6.03	N-121	5.25
N-17	5.43	N-53*	6.57	N-87	4.73	N-122*	5.36
N-18	6.44	N-54	7.14	N-88	6.23	N-124	6.94
N-19	5.54	N-55	6.46	N-89	6.25	N-125*	5.24
N-20	6.44	N-56	6.60	N-90	6.35	N-127	6.81
N-21	6.52	N-58	6.89	N-91	6.42	N-129	7.30
N-22	5.29	N-59	6.68	N-93	6.76	N-131	6.56
N-23	6.31	N-60	6.96	N-94	6.65	N-132	6.70
N-24	6.55	N-61	6.89	N-95*	5.06	N-133	6.64
N-25	6.42	N-62	6.98	N-96	6.23	N-135	6.37
N-26	6.75	N-63	6.35	N-97*	4.98	N-136	6.23
N-27	6.15	N-64	5.33	N-98*	5.07	N-140	6.08
N-28	6.64	N-65	6.75	N-99	4.98	N-141	6.50
N-30	6.35	N-66	6.02	N-100	5.07	N-143	6.61
N-31	5.43	N-67	6.57	N-101	6.94	N-145	6.38
N-32	6.17	N-68	6.66	N-102	6.85	N-157	5.54
N-33	6.57	N-69	6.80	N-103	6.71	N-126	6.33
N-34	6.75						

Representative corrected $\delta^{18}$ O values of garnet in diamondiferous
xenoliths from the Nyurbinskaya pipe

**Note**:\*- xenoliths of ultramafic compositions with high content of chromium in garnet ( $\geq 8$  wt.% Cr<sub>2</sub>O<sub>3</sub>)

in a narrow range 4.98 to 5.36‰. These entire garnet samples lie within the range of average mantle values (5.5±0.45 ‰, [29]). Garnets of more probable lherzolitic association have usually restricted variation  $\delta^{1\Box}$  values inside mantle estimations range but do not all have equivocally "normal" peridotitic values except one sample N-53 with  $\delta^{1\Box}$  value – 6.57‰. Garnets from xenoliths of suggested websteritic paragenesis in most cases have  $\delta^{1\Box}$  values above 6.0‰ and only in two samples oxygen isotope estimations were 5.89 and 5.98‰. It should be

pointed that in two samples of this paragenesis garnets have  $\delta^{||}$  value 7.30 and even 8.59 ‰. Garnet of eclogitic paragenesis has shown the widest variation (4.73 – 9.65 ‰) and subsequently the highest as well  $\delta^{||}$  value. No very light values less than 4.0 ‰ are observed, as reported from some eclogites [28]. It should be pointed that those garnets with strikingly different from the most samples high  $\delta^{||}$  signatures in pyroxenite and eclogite xenoliths, the garnets having values of 8.59‰ and 9.65‰, and it is suggested by comparison with other eclogitic garnets that these 2 xenoliths have no obvious distinction in their major elements composition or other features.



#### Nurbinskaya Garnets

Fig. 4. Oxygen isotope values of garnets in mantle xenolith from the Nurbinskaya kimberlite pipe.

Estimated  $\delta^{\square}$  values have shown any obvious correlation with major elements (FeO or CaO) of garnet, as reported for eclogite diamondiferous xenoliths from the Udachnaya kimberlite pipe by Jacob et al. [22]. We didn't found any correlation of oxygen isotopes behavior with the different varieties of xenoliths or with other criteria as well. But it is evidenced from their chemistry that garnets from most samples showing high  $\delta^{\square}$  value (above 6 ‰) as a rule exhibited strong secondary alteration and zoning. These are obvious from comparison of major elements contents and oxygen isotope analyses of garnets from given samples.
### DISCUSSION

Mantle xenoliths in kimberlites worldwide are represented by mafic and ultramafic rocks. Both these types are coupled and it is most probable that they were formed in different terrains of Yakutian kimberlite province at the same time. The lowest peridotite osmium isotopic compositions require Re depletion in the mid-Archaean (3.2 Ga) according to [33] and this age is interpreted as the time of differentiation of the Siberian lithospheric mantle. Ancient origin for Siberian lithosphere supported by Re-Os and Sm-Nd model ages for eclogite and peridotite xenoliths from pipe Udachnaya is obtained by different authors. Age of formation of parent xenoliths mantle rocks and their differentiation in mantle according with isotopic data and other petrologic evidence is not less than 3.0 Ga (e.g. [34], and references therein). Besides the Re/Os estimations of diamonds of ultramafic paragenesis from Yakutian kimberlites give an age 2.8-3.4 Ga [34]. If it is correct for the whole province and for the xenoliths from different pipes the conditions of PT-formation of similar mantle rocks as well their isotope systems should be similar. New investigations of Re-Os systematics in mantle samples and diamonds show that the Re/Os isotope system holds the key to further progress in understanding continental mantle evolution, not only with reference to mantle keel depletion but also to the addition of eclogitic components via subduction [20].

The carbon isotopic ratios of diamonds from kimberlites reveal that many diamonds have  $\delta^{13}$ C values distinct from typical mantle [10-14, 17]. The reasonable explanation for the high and very low  $\delta^{13}$ C values involves an ultimate source from the crust. This crustal source is also supported by sulfur isotopes of sulfide diamond inclusions [15]. In addition, oxygen isotopic data suggest that many eclogite xenoliths, especially the group C eclogites, are representatives of subducted crust. The presence and preservation of coesite in eclogites from Udachnaya and other kimberlite pipes of Yakutia provides evidence of the development of the sub-continental lithospheric mantle by subduction of ancient oceanic crust [22, 46].

The distribution of  $\delta^{||}$  values of garnets from different mafic and ultramafic mantle xenoliths from the Nyurbinskaya kimberlite pipe in comparison with others is shown on Fig. 5. As were shown previously for the xenoliths from the Udachnaya and Obnazhennaya pipes some estimations of oxygen isotopes of peridotitic garnets give  $\delta^{||}$  values of garnets that are outside the average mantle values [53]. A significant number (~20%) of garnets of more than 100 samples from these pipes has been lies outside the mantle window both above and below. Some garnets with high  $\delta^{||}$  from these pipes belong to the xenoliths of mantle pyroxenites and partly lherzolites [53]. Basing on account of these data it was concluded that besides eclogites some xenoliths of pyroxenites and peridotites can represent the remnants of subducted crust. Oxygen isotope compositions of garnets from different mantle xenoliths of the Udachnaya and other kimberlite pipes of Yakutia evidenced that not only eclogites as well some ultramafic xenoliths of peridotite-pyroxenite suite and alkremites contain garnet with  $\delta^{18}$ O values lying beyond of the mantle range that could be interpreted as a result of subduction [53, 57].

Variations in oxygen isotope composition of garnet in diamondiferous mantle xenoliths population from the Nurbinskaya pipe that are containing diamonds and are parent rocks for their origin are wide and predominantly anomalously high (see Fig. 4). In many cases garnets have values  $\delta^{18}$ O that are far outside of mantle range and estimations positioned on much more higher levels than even minerals or rock of possible crustal formation or with rather obvious evidence for their relationship with subduction processes. If we will state on conclusions that all samples of eclogites and websterites with oxygen  $\delta^{18}$ O values of garnets clearly above the mantle range (even 6‰) we will have anomaly large quantity of xenoliths of possible crustal origin. It is obvious that about 1/2 of the whole eclogitic and websteritic xenoliths population could be originated via subduction (Fig. 4). Anomalously wide variation and high  $\delta^{18}$ O values of garnets in diamondiferous xenoliths from the Nyurbinskaya pipe that predominantly lie beyond of the mantle value is clearly seen in comparative distribution with previous estimations for the xenoliths from the kimberlites of other pipes of Yakutian and South African provinces (Fig. 5).



Fig. 5. Comparison of oxygen isotope data of garnets in mantle xenolith from different kimberlite pipes.

**Note**: Oxygen isotope values for eclogitic xenoliths garnets from the Yakutian kimberlites (after Snyder at al. [41] with addition).

As it was shown earlier high  $\delta^{18}$ O values have been met more often in garnets from kyanite- and coesite-eclogites as in kimberlites of Yakutia and South Africa as well [21, 40, 53]. It is possible to speculate from different petrologic evidences that some types of eclogites are products of subducted oceanic crust. It is confirmed first of all for the coesite-bearing and some diamondiferous eclogites. Coesite-bearing eclogites present in the Udachnaya, Zagadochnaya, Sytykanskaya and other kimberlite pipes of Daldyn-Alakitsky region, which represent the central part of Yakutian kimberlite province as well the central part of the Siberian platform [37, 53]. It should be noticed that the wide distribution of coesite in lithospheric mantle under Siberian platform also confirmed by often presence of coesite as inclusions in diamonds from kimberlites Mir, Udachnaya, Zarnitsa and others [4, 5]. These data give a strong evidence for addition of crustal component during formation of eclogites and lithospheric mantle under Siberian platform as well as for the presence of crustal component in surrounded medium during formation of diamonds of eclogitic paragenesis.

Another line of evidence for subduction is the preservation of isotopically light diamonds in some mantle xenoliths as well in kimberlites. This is the most important signature from the diamonds carbon isotopes study that could be identified for the possibility of crustal protoliths and subduction by their formation [22]. The C isotopic ratio of diamonds from the Yakutian kimberlites has shown that many diamonds have  $\delta^{13}C$  distinct from the typical mantle value [17]. Such crystals are widespread in all well studied kimberlite pipes of Daldyn-Alakitsky region and present in the Mir pipe and others kimberlites of the province. Besides, in many cases isotopically light diamonds are presented by cubic crystals that widespread in kyanite and other types of highly aluminous eclogites. It should be pointed that proportion of isotopically light diamonds increase from the south to the north of the province. The most reasonable explanation for the high  $\delta^{13}C$  is a contribution of carbon with appropriate isotope ration in time of diamond formation from the subducted oceanic crust. It should be pointed that cubic crystals are often met among diamonds of the Nyurbinskaya pipe and they comprise about 3% of the whole diamond population in this pipe.

In case with oxygen isotope composition of garnets in diamondiferous xenoliths from the Nyurbinskaya pipe not only garnets of Group C eclogites have high oxygen isotope values but more than 2/3 parts of Group B eclogites and garnet websterites show  $\delta^{18}$ O values far above on mantle range. From our point of view this phenomenon is not possible to explain only by subduction model of their origin.

Although it has been suggested that mantle fractionation models might be able to explain derivation of low  $\delta^{I^{\Box}}C$  values from "normal" mantle carbon reservoirs with initial  $\delta^{I^{\Box}}C$  near -5‰ [7], this type of model does not account for the anomalously high oxygen isotope values of coexisting silicates. Oxygen isotope crystal–liquid fractionation is minimal at mantle temperatures, and differences in pressure do not cause a measurable change [9]. Igneous processes in

the upper mantle thus cannot account for the wide range of  $\delta^{1\square}$  alues of mantle eclogites. As shown above, most workers interpret the anomalously high and low oxygen isotope ratios of mantle eclogite xenoliths as the result of subduction and prograde metamorphism of oceanic lithosphere that has undergone oxygen isotope exchange with sea water on or near the sea floor [23, 28]. But discussion on possible reason of wide variations of  $\delta^{1\square}$  values in eclogite xenoliths still is far from the finishing. On a base of obvious correlation of high oxygen isotope values of garnets in many studied samples with zonation of garnets and strong metasomatic alteration of samples with diamonds from the Nyurbinskaya we can conclude that oxygen composition of garnets in many samples experienced fractionation through metasomatic influence. It is possible to speculate from these evidences that many eclogite and garnet websterite xenoliths are products of subducted oceanic crust or experienced a strong metasomatic influence. There is ample evidence that latter statement is confident and is confirmed by wide distribution of zoned garnet in investigated parcel of samples and intensive development of phlogopite and other metasomatic minerals in studied samples.. Another line of evidence for the possible influence of the metasomatose and alteration on oxygen isotope composition of garnet is the wide distribution of samples with well developed textures of deformation and cataclasis among studied xenoliths and development of veins of partial melt products, which are similar to described in the Udachnaya pipe [53]. Taking all these observations together indicates that the presence of different kind of metasomatic fluids is widely recorded in eclogite and websterite xenoliths. In some cases these fluids probable have had a strong subduction signature leading to a high  $\delta^{\Box}$  values as found in many eclogites from kimberlites worldwide [30, 41, 53].

It is not possible to define at that moment what were the conditions of these processes and at what time it has happened but it is the most reasonable explanation that metasomatic fluids are account for the anomalously high oxygen isotope values of garnet in diamondiferous xenoliths of this pipe. It should be pointed that diamonds and xenoliths as well kimberlites itself contain strong evidence of unusual features of mantle representatives of this newly discovered kimberlite field.

# CONCLUSION

The Nurbinskaya pipe in Nakynsky field (Yakutia) is an unusual occurrence of diamondiferous xenoliths with the wide variations in major element chemistry of garnet the main constitution of these rocks. Wide distribution of eclogitic and websteritic or probable megacrysts varieties is revealed.

New results, obtained on oxygen isotope composition of garnets in diamondiferous xenoliths from the Nurbinskaya pipe, show a wide variation and predominantly high  $\delta^{18}$ O values of garnets in different varieties xenoliths of eclogite and pyroxenite composition that never have been met before in one given pipe.

Presence of zoned garnet in many investigated samples and other evidence give a strong support for the influence of metasomatose on garnets oxygen isotope composition and other peculiarities of xenoliths from this pipe. These evidences allowed explaining the wide oxygen isotope composition of garnets from the Nurbinskaya pipe as a result of introducing of metasomatic fluids on one stage of such events probably connected with the global mantle metasomatose.

Anomalous wide variations and high  $\delta^{18}$ O values of garnets in diamondiferous xenoliths from the Nyurbinskaya pipe that lie beyond of the mantle value and rather different of previous estimations for the xenoliths from the kimberlites of other fields of Yakutian province suggests about intensive lithospheric mantle evolution beneath this field.

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### RECONSTRUCTION OF THE MANTLE SEQUENCES AND THE STRUCTURE OF THE FEEDING AND VEIN MAGMATIC SYSTEMS BENEATH THE KIMBERLITE FIELDS OF SIBERIAN PLATFORM

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Monomineral thermobarometry including 4 new methods for clinopyroxene, garnet Cr – spinel and ilmenite correlated with the othopyroxene thermobarometry was used for the reconstruction of the mantle columns structure beneath 5 kimberlite fields in Siberian platform. Heavy minerals concentrate of 18 kimberlite pipes 500 analyses in average for PT estimates for each pipes. Position of the geotherms found using each monomineral methods are close and allow to determine mantle layering consisting from 10-12 primary subduction horizons the features of mineralogy are depending on the degree of metasomatism and melt interaction. Each mantle column have in the bottom thermally excited horizon, then 3-4 horizons with the dunite – harzburgite – lherzolite – eclogite sequence where the degree of depletion is rising to the bottom and 3-5 horizons at the upper part of the columns divided with the pyroxenite horizon at the middle part.

Splitting of the geotherms is characteristic for the most of TP diagrams referring to the multistage melt percolation processes typical for the mantle beneath kimberlite pipes. The most large pipes are usually diamond bearing and have the most depleted peridotite lens under the asthenospheric layer. The sub adiabatic geotherm (or PT path of the melts) and the dense concentration of the pyroxenites bodies dispersed in the peridotite columns corresponds to the high concentration of the melts beneath large pipes. The structure of the mantle sequence may highly very in kimberlite field and even within the single pipe groups.

Clinopyroxene geotherms as a rule are more low –temperature. They obviously mark the pyroxenite lens in the middle part of sequence dividing the low column part with K – metasomatism and upper part with the Na – melt influence on the peridotites. Garnets give 2-3 geotherms the most hi-T one corresponds to the peridotites with signs of interaction with protokimberlite melt. Location of magma feeding system in pre – eruption stage is marked by the ilmenite thermobarometry for megacrysts. Cr –Mg rich varieties show the TP values for metasomatites and pyroxenites and Mn – Fe – rich ilmenites for eclogites and pyroxenites.

## INTRODUCTION

The reconstruction of the structure of the lithospheric mantle allow to decipher the processes of the creation of the continents and influence of the plums at the lower horizons of the lithosphere what in turn is determined by the convective mantle [29] motions. Major part of the diamond was captured by the kimberlites from the lower part of the lithospheric mantle [11, 20, 36, 38, 43, 46, 47, 54, 55, 62]. In the early works devoted to kimberlites and diamonds the study of the mantle xenoliths and mantle structure play important work [19, 38] what allow mapping mantle lithosphere [32]. Due to high alteration degree of xenoliths the structure of mantle was obtained mainly for the Udachnaya [22], Obnazhennaya pipes [68] and Mir [14, 15, 18, 43, 57, 62]. Thermobarometric estimates using garnet [32, 34, 52] made by W.Griffin and his colleagues [14] allow to compile the profile of the lithospheric mantle and determine the evolution of the mantle in the Northern part of the Siberian platform [32, 53]. This variant of garnet thermobarometry gives rather preliminary results for the pressures which sometimes are far from those determined with the polymineral methods. Using only garnets do not allow obtaining the complete image of the mantle sequences.

More then 15000 original microprobe analyses of the xenocrystic minerals from Siberian kimberlites from 6 kimberlitic regions of Siberian platform were made for the 3 reports on the exploration of mantle structure and dynamics of the evolution for the ALROSA stock made by the authors for this article.

# Methods

In this work only results of calculation the monomineral thermobarometry for 5 major minerals in mantle rocks: orthopyroxene [22, 44], clinopyroxene [7,8,47], garnet [4, 6, 25, 35, 58], chromite [10, 69] and ilmenite [3, 11, 31, 69] were used. Four last new methods were calibrated on the opx's estimates and include new variants of barometers and thermometers transformed to monomineral version [10,11, 69]. Their usage show the agreement with the polymineral methods commonly used in the mantle thermobarometry [24, 25, 35, 40, 45, 46, 47, 51, 58, 69]. Detail mantle sections fro the more then 30 kimberlite pipes in Siberian platform revealed the layered structure of the mantle beneath kimberlite regions and the position of the feeding and vein systems [10-17].

### **RECONSTRUCTION OF THE MANTLE COLUMNS BENEATH THE KIMBERLITE FIELDS OF SIBERIAN PLATFORM Daldyn region**

Three types of the mantle section may be recognized there. (fig. 1). The first type is characteristic for Udachnaya [5] and Zarnitsa pipes [5, 13]. The last one show discrete interval for the pressures and contrast layering what possibly is explained by the presence of pyroxenite horizons in the basement of each paleo slabs



Fig. 1. **TP estimates for the mantle columns beneath the kimberlite fields of the** Daldyn region

1. garnet thermobarometry for peridotites [4,6]; 2- the same for pyroxenites; 3.- clinopyroxene thermobarometry (Cr) [47]; 4- clinopyroxene thermobarometry (Jd- Di) [7,8]; 5. orthopyroxene thermobarometry [24, 44]; 6. Sp thermobarometry [10]; 7. Ilm thermobarometry [11].

connected with the lithospheric keel (eclogites of B-type) [18, 36, 61, 62]. In the middle part of the section the anatexic and hybrid pyroxenites (40-50 kbar), built up the pyroxenite lens [5, 42, 43, 53]. Pyroxenite bodies are also abundant in the spinel facie and near Gar-Sp transition. [5, 53]. Hardly depleted dunites [55] are found within two intervals in 45-60 kbar [5,13]. More abundant in Cr- spinel rocks are tracing the most low temperature conditions in he 32-58 kbar interval for

#### Table 1.

Representative analyses of garnets from Yukutia kimberlites												
Element	SiO <sub>2</sub>	TiO <sub>2</sub>	Al <sub>2</sub> O <sub>3</sub>	Cr <sub>2</sub> O <sub>3</sub>	FeO	MnO	MgO	CaO	Na <sub>2</sub> O	K <sub>2</sub> O	Total	
1	2	3	4	5	6	7	8	9	10	11	12	
				DAI	DYN F	ELD				•		
AEROMA	GNITN	AYA PI	PE									
AerG01	39,44	0,65	12,90	11,64	6,51	0,43	19,01	9,22	0,10	0,00	99,90	
AerG04	41,30	0,02	16,26	9,66	7,05	0,43	23,46	0,91	0,02	0,00	99,10	
AerG05	40,94	0,41	16,39	8,82	7,84	0,45	19,01	6,10	0,06	0,00	100,02	
AerG10	40,79	0,09	17,06	8,45	7,80	0,47	19,02	6,11	0,03	0,00	99,82	
AerG14	41,49	0,62	17,70	6,25	7,07	0,39	20,75	5,48	0,07	0,00	99,81	
AerG34	41,62	0,09	20,50	4,07	8,12	0,44	19,58	5,82	0,03	0,01	100,27	
AerG38	41,88	0,24	20,67	3,33	7,88	0,38	20,55	4,67	0,06	0,01	99,67	
OSENNYA	Y PIPE	2										
OSGa066	41,63	0,25	7,39	18,02	6,58	0,37	8,68	19,42	0,07	0,00	100,41	
OSGa067	41,79	0,25	16,42	7,92	7,59	0,49	20,07	5,78	0,06	0,01	99,37	
OSGa068	42,11	0,30	16,16	8,39	7,84	0,43	18,76	6,30	0,04	0,00	99,33	
OSGa071	43,55	0,73	20,64	1,70	8,11	0,30	20,85	4,33	0,06	0,01	99,28	
OSGa072	42,73	0.37	21,18	0.80	11.10	0.51	18,82	4,71	0.06	0,02	99,29	
OSGa073	42.64	0.01	19.07	5.22	7.32	0.39	19.78	5.79	0.04	0.00	99.26	
OSGa078	41.17	0.04	15.35	10.32	7.49	0.53	18.41	6.81	0.06	0.02	99.19	
DOLGOZHDANNAYA PIPE												
DIG28	41 47	1 00	19.01	3 31	9.85	0.42	19.06	5.07	0.08	0.00	99.26	
D1G29	41 41	0.93	19.91	2.28	9.65	0.42	19 74	4 58	0.09	0.00	99.00	
DIG32	41.44	0.94	19.55	2.37	9.84	0.37	19.81	4.68	0.08	0.00	99.07	
DIG1	42.62	0.01	19.19	6.19	6.61	0.35	22.91	2.31	0.03	0.01	100.23	
DlG2	41,20	0,02	17,65	7,32	7,58	0,42	18,52	6,36	0,02	0,01	99,09	
DIG5	41,82	1,02	19,40	2,80	9,85	0,40	19,22	4,90	0,08	0,00	99,48	
DlG6	41,28	0,02	17,66	7,34	7,56	0,48	18,30	6,44	0,02	0,00	99,10	
ZARNITS	A PIPE				1	1	1		1			
ZarG04	40,86	0,47	18,76	3,47	12,45	0,47	12,92	10,63	0,05	0,00	100,08	
ZarG10	42,28	0,08	18,53	5,79	7,83	0,32	20,30	5,10	0,02	0,00	100,25	
ZarG20	41,44	0,17	18,37	6,48	7,93	0,45	19,74	5,06	0,05	0,00	99,69	
ZarG19	41,37	0,43	17,05	7,76	7,61	0,46	19,49	5,76	0,08	0,00	100,01	
ZarG24	40,53	0,34	16,61	8,07	10,96	0,44	17,85	5,18	0,08	0,00	100,07	
ZarG06	40,21	0,35	14,41	9,99	10,18	0,66	15,23	8,22	0,07	0,00	99,31	
ZarG17	39,90	1,82	11,66	11,54	8,47	0,37	17,36	7,79	0,13	0,00	99,04	
			MA	LOBOTU	UOBINS	KY RE	GION					
MIR PIPE												
MirGar28	40,49	0,17	16,98	7,97	6,64	0,41	17,51	8,44	0,06	0,01	98,68	
MirGar56	41,50	0,40	17,37	6,73	7,38	0,45	17,80	8,19	0,12	0,00	99,94	
MirGar16	41,67	0,11	19,35	4,97	8,17	0,41	19,94	4,62	0,06	0,01	99,31	
MirGar12	41,42	0,27	19,59	5,07	7,48	0,45	20,63	4,62	0,07	0,01	99,61	
MirGar26	41,85	0,60	19,93	3,77	6,87	0,33	21,09	4,62	0,08	0,02	99,16	
MirGar53	41,56	0,18	20,91	3,29	9,05	0,53	19,68	4,98	0,04	0,00	100,22	
MirGar52	41,11	0,44	22,37	0,00	13,63	0,41	16,69	4,71	0,15	0,00	99,51	

Table 1. Continued

1	2	2	4	5	6	7	0	Δ	10	11	10			
		3	4	3	0	1	δ	9	10	11	12			
DACHNAY	A PIPE	Ľ	r		1		1							
DacGr74	40,18	0,05	15,55	9,92	7,66	0,59	17,51	7,48	0,03	0,00	98,97			
Dac23	41,48	0,13	17,90	7,55	7,34	0,48	19,56	4,59	0,13	0,04	99,20			
DacGr63	41,89	0,13	20,25	4,37	7,32	0,48	21,17	3,52	0,04	0,00	99,17			
Dac19	42,05	0,45	20,60	3,50	6,87	0,39	21,00	4,72	0,25	0,01	99,84			
DacGr70	41,51	0,09	20,94	2,84	8,76	0,47	19,37	5,11	0,03	0,00	99,12			
INTERN	ATIO	NAL'	NAYA I	PIPE										
IntG01	40,29	0,10	14,10	12,40	7,79	0,58	18,05	6,37	0,06	0,00	99,75			
IntG05	41,15	0,27	17,71	7,93	7,49	0,47	20,11	5,09	0,00	0,00	100,23			
IntG10	41,68	0,30	19,54	4,89	8,00	0,45	19,84	4,89	0,05	0,00	99,64			
IntG11	42,17	0,11	21,13	2,82	9,19	0,52	19,68	4,51	0,03	0,00	100,16			
IntG13	42,28	0,07	22,28	1,56	9,15	0,50	19,84	4,44	0,03	0,00	100,15			
AMAKINS	KAYA	PIPE												
AmGG02	40,84	0,03	16,86	9,03	7,63	0,49	20,32	4,95	0,05	0,00	100,20			
AmGG03	41,28	0,20	18,17	7,05	8,17	0,52	19,13	5,66	0,06	0,01	100,24			
AmGG04	40,64	0,17	18,36	6,46	8,05	0,57	20,14	5,60	0,07	0,02	100,07			
AmGG09	41,92	0,14	20,58	4,04	8,12	0,48	20,23	4,66	0,04	0,00	100,22			
AmGG10	41,33	0,19	20,65	3,10	8,13	0,39	20,89	4,52	0,07	0,01	99,27			
ALAKITE FIELD														
AYKHAL PIPE														
Aikh001	40,34	0,02	12,44	14,44	6,62	0,47	18,21	7,11	0,01	0,03	99,69			
Aikh006	40,20	0,47	10,08	13,40	5,86	0,33	7,49	21,69	0,05	0,00	99,57			
Aikh11	41,52	0,29	17,17	7,26	7,69	0,35	19,54	6,00	0,04	0,00	99,86			
Aikh025	41,86	0,03	20,43	3,74	8,31	0,42	19,66	5,13	0,02	0,00	99,61			
Aikh44	41,38	0,89	20,01	2,01	10,86	0,43	19,10	5,08	0,06	0,00	99,81			
Aikh77	41,97	0,76	21,43	0,55	11,33	0,48	19,19	4,66	0,07	0,01	100,45			
KOMSOM	OL'SKA	AYA PI	PE											
KmGa107	41,49	0,07	17,55	7,88	7,30	0,42	22,18	3,04	0,00	0,01	99,94			
KmGa125	41,23	1,30	16,32	6,69	7,88	0,33	19,89	6,27	0,09	0,00	100,00			
KmGa011	41,98	0,13	19,18	5,76	7,19	0,36	20,30	5,30	0,07	0,00	100,27			
KmGa139	41,95	0,91	18,25	4,89	8,98	0,35	19,81	5,30	0,09	0,01	100,53			
KmGa094	42,57	0,22	20,65	2,64	7,57	0,32	21,33	4,67	0,04	0,00	100,01			
YUBILEY	NAYA F	PIPE												
YbG146	41,59	0,04	16,76	8,48	7,68	0,43	18,97	6,11	0,04	0,01	100,11			
YbG201	41,23	0,03	17,63	7,08	7,54	0,34	19,51	6,11	0,06	0,00	99,52			
YbG150	41,70	0,10	17,86	6,96	7,70	0,46	19,17	5,71	0,04	0,002	99,70			
YbG138	42,17	0,23	19,50	5,13	6,84	0,40	20,85	4,73	0,06	0,01	99,92			
YbG007	42,41	0,02	20,89	3,70	7,97	0,48	21,27	3,50	0,01	0,002	100,26			
SYTYKAN	SKAYA	PIPE									•			
SytG001	41,70	0,08	18,53	6,60	7,11	0,41	21,51	3,40	0,06	0,02	99,41			
SytG067	41,84	0,36	19,02	5,85	6,92	0,38	20,94	4,37	0,08	0,01	99,78			
SytG072	41,94	1,03	18,89	3,60	8,34	0,29	20,58	4,87	0,08	0,01	99,63			
SytG062	41,89	0,23	21,26	2,68	8,42	0,39	20,28	4,56	0,08	0,01	99,80			
SytG043	42,31	0,47	21,04	1,93	7,98	0,42	21,15	4,27	0,03	0,03	99,64			

Table 1. Continued

							-				r		
1	2	3	4	5	6	7	8	9	10	11	12		
OZERNAY	'A PIPE	2											
OZG324	42,45	0,424	16,67	7,37	7,35	0,30	19,47	5,96	0,09	0,02	100,09		
OZG311	42,97	0,207	18,86	6,60	6,70	0,35	19,34	4,18	0,08	0,01	99,30		
OZG290	42,26	0,635	18,36	5,92	7,64	0,40	18,81	5,70	0,11	0,002	99,84		
OZG270	42,59	0,152	20,38	3,52	7,17	0,32	21,16	4,84	0,03	0,01	100,17		
OzG070	41,43	1,16	20,18	1,53	11,01	0,35	19,51	4,75	0,10	0,02	100,05		
				UPPER	MUNA	FIELD							
DEIMOS P	PIPE												
DeiG02	40,84	0,36	14,82	11,36	6,30	0,41	18,23	7,11	0,08	0,00	99,51		
DeiG03	41,65	0,07	15,91	10,79	7,16	0,46	23,33	0,95	0,02	0,00	100,35		
DeiG09	42,09	0,80	17,97	6,19	7,61	0,33	19,53	5,43	0,12	0,01	100,09		
DeiG13	41,91	1,16	19,07	3,10	9,85	0,36	19,21	5,27	0,10	0,00	100,03		
DeiG16	41,85	0,28	22,18	1,31	10,41	0,45	18,37	5,57	0,07	0,00	100,49		
NOVINKA	PIPE												
NovG05	41,83	0,06	15,41	9,75	6,52	0,36	21,64	4,21	0,02	0,01	99,81		
NovG07	41,71	0,11	16,78	8,86	6,29	0,33	22,92	2,64	0,03	0,01	99,68		
NovG11	42,39	0,11	18,36	6,57	6,30	0,33	22,93	2,81	0,02	0,01	99,83		
NovG16	42,00	0,19	19,45	4,41	7,08	0,29	21,03	5,16	0,04	0,01	99,65		
NovG22	42,11	0,91	20,38	1,72	8,43	0,30	21,31	4,38	0,20	0,00	99,74		
ZAPOLYARNAYA PIPE													
ZapG01	41,96	0,15	17,53	7,90	7,04	0,32	20,59	4,68	0,05	0,00	100,22		
ZapG02	41,45	0,42	15,46	8,55	6,96	0,30	19,64	6,14	0,03	0,00	98,95		
ZapG05	41,58	0,43	18,42	6,27	6,20	0,30	20,87	4,86	0,07	0,00	99,00		
ZapG12	42,84	0,27	21,09	3,29	6,38	0,34	21,67	4,29	0,10	0,01	100,27		
ZapG13	43,23	0,93	20,87	1,63	8,11	0,25	20,63	4,38	0,12	0,01	100,17		
KOMSOM	OL'SKA	AYA M	AGNITNA	AYA PIPI	E								
KomMG01	42,06	0,00	15,17	10,70	5,72	0,31	24,06	1,69	0,04	0,03	99,21		
KomMG04	41,14	0,16	16,48	8,80	8,53	0,30	20,38	3,43	0,04	0,03	99,31		
KomMG05	41,64	0,00	16,23	9,19	6,51	0,33	23,66	0,99	0,02	0,02	99,96		
KomMG08	41,51	0,24	18,55	6,03	6,34	0,40	18,64	7,74	0,11	0,01	99,78		
KomMG11	42,10	0,41	20,23	3,72	6,77	0,33	21,55	4,41	0,08	0,01	99,80		
				NAI	KYN FI	ELD							
NYURBIN	SKAYA	PIPE											
13n	41,03	0,09	13,51	13,21	6,86		18,96	6,02	0,04		99,72		
бn	40,89	0,10	13,73	12,77	7,01		18,81	6,17	0,03		99,50		
GNur417	42,31	0,46	17,88	6,07	7,40	0,30	20,27	5,29	0,07		100,05		
GNur403	41,47	0,33	18,65	5,51	6,67	0,36	21,91	4,62	0,06		99,59		
160m50	42,00	0,19	22,03	1,70	9,22		20,11	3,84	0,04		99,13		
BOTUOBI	NSKAY	A PIPE								l			
37B	41,82	0,08	15,87	10,93	7,08	0,47	22,85	0,94	0,01	0,00	100,04		
botgar38	42,11	0,23	19.62	5,43	7,83	0,45	20.08	3,85	0,09	0,01	99.69		
botgar51	41.74	0.07	21.91	2.27	9.33	0.55	19.75	4.15	0.05	0.01	99.83		
botgar17	41.08	0.19	16.02	9.27	7.80	0.50	18.25	6,84	0.05	0.01	100.00		
36B	42.41	0.10	17.61	8.13	7.17	0.39	22.36	2.08	0.05	0.00	100.30		

Udachnaya pipe, garnet bearing rocks are more H-T for the same pressures and ortho – clinopyroxenes give the ranges of the temperatures in the same pressures. [5]. For the mantle sequence beneath Zarnitsa pipe three discrete intervals of the pressures were found estimated using garnet and pyroxenes [13]. Cr-spinel marks the layers upper and lower then this three. The ilmenites show all the intervals showing the branches close to those marked with garnets with the different degree of heating. It is possible that magma did not captured only some selected intervals but contrast in lithology mantle sequence gave the different pressure intervals for Cr-spinels and garnet – pyroxenes (fig.1).

The pyroxenite layer definitely seen on the TP for Udachnaya is not evident on the diagram for Zarnitsa pipe. In this mantle lithosphere sections the heating to the SEAG [52] at the level of Sp-Gar transition and upper was estimated. The subvertical sub-adiabatic TP gradient was determined on orthopyroxene for Udachnaya pipe comparing to the other mantle columns where the nearly conductive TP gradients are prevailing.

Ilmenite TP estimates for Udachnaya refer to more deep (up to 40 kbar) parts of mantle lithosphere and more heated while for Zarnitsa [13] they are tracing step by step all the pressure interval 65 to 25 kbar.

The second type of the mantle section is found for Ozernayaya pipe [68] where the more fertile Cpx- rich rocks are making up the interval 40 to 55 kbar, what is close to the estimates for ilmenites. Lower then 55 kbar the depleted dunite horizon is represented in concentrate by coarse chromite grains giving heated conditions ~60 kbars. The Garnet estimates are low – temperature. Ilmenite TP points allow to suggest the broad distribution of the HT<sup>o</sup>C Ilm – Cpx metasomatites within the long pressure interval.

For the Aeromagnitnaya pipe (3-type), which have the similar structure of the mantle column the close position of the Cpx and Ilm TP estimates within 65-45 kbar also suggest the presence of metasomatic rocks. But upper Ilm are splitting on to three branches with the different heating degree. Cr- spinel estimates are corresponded to low temperature conditions but at the level of 40 kbar they mark three geotherms 35 40 and 45 mv/m<sup>2</sup> in the pyroxenite layer.

Mantle structure beneath the Dolgozhdannaya pipe is close to Osernyaya pipe. But Irelyahskaya show only rich in the Cpx mantle basement and transition from convective to conductive branches.

Judging on the example of the Daldyn region it is possible to say that the same field the structure of the close pipes may be varying in high scale [1, 5]. The pipes even in the same field may carry quite varying set f the mantle peridotite xenoliths and of course eclogites and metasomatites.

#### Alakite region

The most detail reconstructions of the mantle were obtained for the kimberlite pipes of the Alakite field (fig.2) [2, 5, 16, 65]. The TP diagram for the Aykhal pipe reveal the structure similar to those calculated for the South Africa pipes. [48, 49,



Fig. 2. **TP estimates for the mantle columns beneath the kimberlite fields of the Alakite region.** Signs are the same.

60]. Essential heating from 38 to 45 mv/m<sup>2</sup> was found for the lower part 80-50 kbar including the horizon of the garnet dunites 50-60 kbar. The evident 8-9 pressure intervals are marked by the discrete pressure intervals for garnets and Close structure reveal the mantle column beneath the clinopyroxenes. Yubileynaya pipe [2, 5], essentially enriched by clinopyroxene comparing with the others. The 2 horizons of the essentially depleted rocks are found at the basement of the mantle column 58-65 kbar. The high temperature garnet branch (~ 45  $mv/m^2$ ) to 40 kbar, most probably corresponds to the pyroxenites or those from the contact zones of the deep seated feeding channels. The lower part of the mantle section beneath the Sytykan pipe are represented by garnet harzburgites and dunites at the level 60~ 50 kbar, several lenses of the eclogites- like rocks are suggested upper judging by the mineralogical features of the eclogites – like rocks. Pyroxenite lens was found at 45-35 kbars, upper the essentially depleted (harzburgite layer is suggested). TP estimates of the Cr-spinel are located along the low temperature conductive branch for Sytykanskava Aykhal [16] and Ozernava mantle columns. But in lithosphere beneath Komsomolskaya and Yubileynaya they trace several (3-4) separate layers with the vast range of heating. More HT estimates usually are located near the pyroxenite layer. In mantle beneath Sytykanskaya [19, 56] spinel peridotites compile the lens between the heated depleted layer 65-58 kbars and essentially pyroxenitic horizons starting from 50 kbar. Mantle column under the Komsomolskaya pipe are more enriched comparing with the mantle horizons beneath the southern kimberlite pipes. And the layering became similar to those from the mantle Upper Muna field. The deviations for the clinopyroxene TP estimates to the lower temperatures are typical for the eclogites and similar rocks with high Na2O content in pyroxenes.

The TP conditions for the ilmenites from Aykhal pipe are tracing part of mantle columns wit the temperatures upper 35  $mv/m^2$ , were determined using chromites and starting from 50 kbar with pyroxenes. For the Yubileynaya pipe the deviations to the high temperature field on TP diagram are more pronounced in the upper part of the mantle column. The mantle column beneath the Ozernava pipe which is located just near the Yubileynaya shows the HT polybaric ilmenite trend with the strong deviations from conductive branch. The TP estimates for ilmenites beneath Sytykanskaya pipe in the lower part of the mantle section refer to the low temperature gradients beneath the Sytykanskaya pipe and practically coincide with garnet TP points suggesting broad ilmenite - clinopyroxene metasomatism what was found as the hydrous Cpx-Ilm veinlets in the xenoliths. But in the upper part of the mantle section it gives more heated conditions then gradients for the common mantle peridotites. The mantle section for the Komsomolskaya pipe lower part also represented by metasomatites close in TP conditions to the common harzburgites. Heated metasomatites are located upper. Protokimberlite melt intrusion in several stages and accompanied metasomatism is a most probable model for these phenomena.

REPRESENTATIVE	ANALYSES	OF PYR	OXENES	FROM	VUKUTIA	KIMBERLITES
KEI KEDEI IIIII		OF I IN		LICONT	IONOIN	<b>KINIDEKETE</b> D

Element	SiO <sub>2</sub>	TiO <sub>2</sub>	Al <sub>2</sub> O <sub>3</sub>	Cr <sub>2</sub> O <sub>3</sub>	FeO	MnO	MgO	CaO	Na <sub>2</sub> O	K <sub>2</sub> O	Total
1	2	3	4	5	6	7	8	9	10	11	12
DALDYN H	FIELD										
AEROMA	GNITNA	YA PIP	E								
AeCp096	54,64	0,17	2,20	4,23	2,04	0,11	15,17	17,20	3,38	0,02	98,17
AeCp098	54,12	0,35	1,83	0,60	4,20	0,15	19,73	15,42	1,50	0,04	98,70
AeCp103	54,35	0,15	3,39	2,48	2,48	0,13	13,66	18,23	3,57	0,00	98,86
AeCp104	54,57	0,32	2,53	1,74	3,03	0,14	15,49	18,56	2,67	0,03	98,54
AeCp105	53,85	0,20	0,89	2,73	1,79	0,04	15,74	20,94	2,04	0,02	98,81
AeCp106	53,93	0,30	2,30	4,78	2,23	0,10	14,46	16,62	4,07	0,00	98,50
AeCp107	54,31	0,10	1,32	0,51	1,72	0,08	16,24	23,16	1,12	0,00	98,72
AeCp109	54,30	0,56	4,70	0,57	3,99	0,08	15,96	14,27	3,78	0,01	99,20
OSENNYA	Y PIPE										
OsD64	52,90	0,36	3,37	1,58	2,61	0,09	14,72	21,74	1,60	0,00	98,97
OsD68	54,08	0,09	1,31	4,93	2,59	0,10	14,52	17,64	3,20	0,02	98,48
OsD69	53,77	0,04	2,83	3,93	2,25	0,12	14,52	17,75	3,59	0,00	98,81
OsD70	55,22	0,09	2,33	4,28	2,50	0,05	14,48	17,54	3,23	0,01	99,73
OsD71	53,35	0,32	3,55	1,81	2,48	0,08	14,43	21,37	1,84	0,00	99,23
OsD72	55,17	0,04	2,57	4,04	2,66	0,07	14,05	17,38	3,39	0,02	99,39
OTHOPYRO	DXENES										
OsE02	51,88	0,05	4,60	0,28	16,20	0,37	27,32	0,29	0,08	0,00	101,06
OsE03	51,65	0,06	4,70	0,32	15,69	0,29	26,04	0,33	0,03	0,00	99,12
IRELYAK	H PIPE										
IrD-2-1	55,08	0,20	1,40	1,33	3,21	0,12	17,12	19,64	1,68	0,01	99,79
IrD-2-5	54,79	0,20	1,57	1,43	3,56	0,09	15,28	20,18	1,83	0,00	98,93
IrD-2-7	55,58	0,22	1,35	1,29	2,46	0,07	15,72	21,39	1,39	0,01	99,48
IrD-2-8	54,84	0,31	1,29	1,29	2,92	0,09	17,46	19,17	1,39	0,00	98,76
IrD-2-11	54,98	0,19	1,72	0,20	3,98	0,06	15,38	21,62	1,39	0,00	99,52
IrD-2-12	55,21	0,23	1,26	2,20	2,88	0,08	16,07	19,39	1,88	0,02	99,22
MALOBO	TUOBIN	SKY RE	GION								
MIR PIPE					-						
MirD17	54,22	0,14	1,93	2,62	2,09	0,10	15,12	19,82	2,61	0,00	98,65
MirD09	54,66	0,13	2,66	2,81	2,32	0,09	14,54	18,23	3,21	0,01	98,66
MirCp55	54,79	0,14	2,21	4,81	1,97	0,09	14,57	16,84	3,67	0,04	99,13
MirCp30	54,99	0,20	3,37	2,58	1,22	0,06	14,60	19,64	2,94	0,01	99,61
Mir16	54,88	0,16	8,76	0,09	2,34	0,05	10,58	16,10	5,08	0,00	98,04
MirCp01	54,12	0,18	2,73	0,28	1,74	0,06	16,08	21,85	1,57	0,01	98,62
MirCp27	54,66	0,03	0,65	0,32	2,84	0,07	17,17	23,01	0,62	0,01	99,38
DACHNAY	A PIPE	r				1			1	r	
DachD01	53,66	0,41	2,30	0,07	5,75	0,15	16,97	17,16	2,13	0,03	98,62
DachD06	54,05	0,48	5,52	3,67	1,43	0,07	12,20	16,15	4,98	0,00	98,55
DachD16	53,37	0,16	3,80	5,99	2,79	0,10	13,28	13,78	5,97	0,02	99,25
DachD14	54,32	0,05	3,38	3,38	2,97	0,11	13,59	16,85	4,06	0,01	98,72

Table 2. Continued

1	2	3	4	5	6	7	8	9	10	11	12		
DachD21	53,95	0,24	2,28	2,83	2,44	0,09	15,84	18,26	3,06	0,04	99,03		
DachD22	50,27	0,97	2,89	0,68	8,28	0,21	15,84	19,02	0,25	0,00	98,40		
DachD28	54,23	0,14	2,70	4,69	2,16	0,06	13,54	16,91	3,90	0,01	98,34		
DachD33	54,10	0,28	3,75	1,06	1,26	0,04	15,34	20,18	2,55	0,01	98,56		
DachD46	54,17	0,13	2,78	4,12	2,27	0,09	14,07	17,38	3,61	0,00	98,61		
DachD49	53,76	0,55	1,64	0,69	7,36	0,22	16,75	18,75	0,24	0,00	99,96		
INTERNAT	TONAL	'NAYA I	PIPE										
IntD01	54,78	0,17	2,82	2,90	2,11	0,07	14,59	18,78	3,01	0,01	99,24		
IntD02	54,91	0,17	3,21	1,84	2,38	0,05	14,61	19,18	2,86	0,00	99,21		
IntD06	54,75	0,17	2,71	3,04	2,24	0,08	14,34	18,68	3,09	0,01	99,10		
IntD07	53,88	0,19	2,27	1,81	2,67	0,09	14,95	20,66	2,32	0,01	98,85		
IntD10	55,50	0,03	7,64	3,70	3,85	0,13	10,27	10,41	7,37	0,01	98,91		
IntD11	54,65	0,19	2,27	5,02	2,41	0,11	14,17	16,78	3,65	0,02	99,27		
IntD12	54,55	0,19	4,24	3,19	1,47	0,06	13,61	18,03	3,78	0,01	99,13		
IntD13	54,37	0,14	2,92	3,37	2,22	0,08	14,30	18,17	3,30	0,00	98,87		
IntD14	53,84	0,18	2,44	0,49	4,31	0,11	15,14	19,87	2,23	0,04	98,66		
IntD16	54,84	0,03	3,42	6,43	2,67	0,07	12,61	14,17	5,23	0,01	99,48		
IntD18	54,76	0,16	2,54	2,60	2,39	0,06	14,78	19,75	2,65	0,03	99,71		
ALAKITE FIELD													
AYKHAL P	PIPE						1						
AiD008	55,13	0,09	2,59	3,29	2,50	0,12	14,89	18,19	3,13	0,01	99,94		
AiD011	55,36	0,16	3,03	1,95	2,91	0,09	15,25	18,28	3,13	0,02	100,17		
AiD013	55,43	0,20	2,51	3,33	2,56	0,12	15,16	17,67	3,19	0,03	100,21		
AiD014	55,09	0,13	2,45	2,26	1,94	0,08	15,49	20,41	2,31	0,01	100,17		
AiD015	55,00	0,39	1,84	0,61	4,80	0,17	18,61	16,85	1,61	0,05	99,92		
AiD020	54,55	0,17	3,17	0,98	1,72	0,03	15,70	22,16	1,58	0,01	100,07		
OZERNAYA	PIPE		[		r	1	1						
OzCp029	52,67	0,47	2,76	0,70	6,92	0,15	16,06	19,43	0,24	0,01	99,40		
OzCp126	55,38	0,19	3,18	3,75	2,69	0,03	14,34	16,28	3,90	0,03	99,77		
OzCp128	54,79	0,09	0,27	2,39	2,24	0,02	15,85	22,23	1,70	0,01	99,59		
OzCp131	55,10	0,10	0,26	2,86	2,12	0,00	15,46	21,59	1,95	0,01	99,45		
OzCp135	55,19	0,23	2,59	0,98	3,68	0,04	15,37	19,83	2,33	0,02	100,26		
OzCp136	54,80	0,09	0,51	2,18	2,66	0,05	15,94	21,58	1,79	0,00	99,60		
OzCp137	54,47	0,08	0,51	2,22	2,66	0,00	15,83	21,46	1,95	0,02	99,21		
YUBILEYN	AYA PI	PE	1.0.6	1.00		0.01	1 4 9 1	00.01	1 50	0.04			
YuCrD129	55,24	0,02	1,36	1,98	2,29	0,06	16,81	20,21	1,69	0,04	99,70		
YuCrD238	54,92	0,06	3,44	3,32	2,33	0,08	13,72	17,42	3,60	0,00	98,90		
YuCrD15	55,43	0,19	2,65	3,79	2,33	0,09	14,86	16,59	3,49	0,03	99,46		
YuCrD146	54,71	0,12	0,43	2,41	2,35	0,07	15,97	21,67	1,79	0,01	99,52		
YuCrD216	54,84	0,25	2,37	3,57	2,36	0,09	14,54	18,10	3,07	0,03	99,23		
YuCrD39	55,07	0,15	2,40	4,96	2,38	0,10	14,42	16,00	3,79	0,04	99,30		
YuCrD101	55,00	0,16	1,60	2,62	2,41	0,07	15,17	20,20	2,41	0,00	99,64		
YuCrD217	55,19	0,08	0,55	3,04	2,42	0,07	15,41	20,73	2,15	0,01	99,65		

Table 2. Continued

-	-			-		-	0		10		
1	2	3	4	5	6	7	8	9	10	11	12
SYTYKAN	SKAYA	PIPE			-						
SytD028	54,16	0,12	0,70	1,82	2,17	0,08	16,18	22,02	1,63	0,04	98,92
SytD030	54,36	0,09	2,87	3,65	2,25	0,09	13,96	18,47	3,33	0,08	99,15
SytD034	54,48	0,13	0,51	2,18	2,01	0,06	15,94	22,27	1,54	0,06	99,18
SytD035	54,77	0,35	1,73	4,30	2,50	0,12	14,35	17,86	3,37	0,09	99,43
SytD036	53,98	0,22	2,95	0,73	3,21	0,07	14,71	21,02	2,18	0,06	99,13
SytD041	54,04	0,22	2,98	0,73	3,32	0,08	14,99	20,84	2,33	0,07	99,59
SytD043	54,05	0,21	2,91	0,77	3,30	0,08	14,47	20,62	2,36	0,06	98,83
SytD045	54,10	0,23	2,88	0,76	3,21	0,07	14,68	20,44	2,36	0,07	98,81
SytD047	54,08	0,20	3,05	0,77	3,16	0,06	14,53	20,91	2,24	0,03	99,03
SytD049	53,81	0,20	2,97	0,74	3,17	0,08	14,49	20,86	2,23	0,04	98,59
SytD050	54,42	0,05	2,82	2,94	2,23	0,07	14,01	18,91	3,09	0,03	98,58
SytD052	54,36	0,05	2,83	3,01	2,18	0,10	14,27	18,96	3,06	0,03	98,85
KOMSOM	OL'SKA	YA PIPI	E								
KmsDi063	53,83	0,24	2,43	2,55	2,88	0,07	14,92	19,65	2,68	0,03	99,28
KmsDi065	55,07	0,14	2,33	2,30	2,10	0,07	15,82	19,87	2,71	0,03	100,44
KmsDi066	52,36	0,20	2,76	1,73	2,10	0,06	15,51	21,09	2,98	0,01	98,79
KmsDi067	54,97	0,10	3,45	3,97	1,79	0,09	13,84	18,39	3,64	0,00	100,24
KmsDi070	54,82	0,23	2,33	5,56	2,33	0,10	13,93	16,69	4,03	0,03	100,05
KmsDi071	54,96	0,20	1,66	2,75	2,11	0,08	15,42	21,27	2,39	0,02	100,86
KmsDi075	54,19	0,31	4,53	2,74	1,60	0,06	13,92	18,94	3,47	0,00	99,76
KmsDi076	54,54	0,27	3,30	4,47	2,75	0,09	13,65	16,03	4,44	0,02	99,55
UPPER MU	JNA FIE	LD									
DEIMOS P	IPE										
DeiD01	53,95	0,16	0,66	2,48	3,21	0,07	15,55	20,66	2,03	0,01	98,78
DeiD02	55,08	0,13	0,64	0,92	3,50	0,06	16,12	21,13	1,49	0,01	99,08
DeiD06	51,84	0,15	5,54	0,96	1,89	0,08	14,71	21,46	1,76	0,03	98,41
DeiD08	50,78	0,30	3,35	0,12	7,65	0,13	12,71	22,81	0,48	0,00	98,34
DeiD10	54,71	0,11	1,30	0,82	1,52	0,06	16,95	23,24	0,88	0,00	99,58
DeiD11	52,96	0,32	0,81	1,26	3,45	0,08	16,25	23,42	0,80	0,02	99,37
DeiD14	53,24	0,32	4,41	1,17	1,84	0,05	14,84	21,41	2,06	0,01	99,34
DeiD17	54,47	0,14	2,22	1,46	1,60	0,05	15,90	21,54	1,67	0,05	99,10
DeiD19	54,77	0,14	0,41	2,75	2,27	0,06	15,96	21,89	1,90	0,00	100,14
NOVINKA	PIPE										
NovDi01	52,87	0,17	2,66	4,42	2,30	0,11	14,59	15,60	6,96	0,02	99,71
NovDi04	54,33	0,32	1,72	1,40	2,85	0,12	18,29	16,97	3,51	0,05	99,55
NovDi07	55,29	0,29	1,83	3,39	1,95	0,07	15,87	17,18	3,10	0,04	99,01
NovDi09	52,14	0,23	3,15	0,03	8,91	0,17	13,45	19,95	0,94	0,04	99,01
NovDi11	55,07	0,29	1,54	0,97	2,27	0,08	17,81	19,38	1,54	0,05	99,01
NovDi12	54,48	0,26	1,88	4,03	1,99	0,07	15,64	17,72	2,92	0,04	99,01
ZAPO	OLYAR	NAYA P	IPE	•			•	•	•	•	·
ZpD134	55,65	0,25	1,55	3,07	1,50	0,09	16,11	20,35	2,21	0,06	100,85
ZpD137	55,98	0,27	2,07	0,87	3,37	0,15	19,08	16,79	1,78	0,04	100,40

Table 2. Continued

1	2	3	4	5	6	7	8	9	10	11	12
ZpD138	55,72	0,30	1,79	2,36	1,88	0,07	16,13	19,78	2,11	0,04	100,20
ZpD139	55,40	0,34	1,53	0,98	2,44	0,09	17,75	20,07	1,51	0,06	100,16
ZpD141	54,84	0,27	1,51	3,30	1,45	0,03	16,04	20,42	2,23	0,05	100,14
ZpD142	55,41	0,20	1,49	3,52	1,76	0,08	16,17	19,04	2,42	0,05	100,15
ZpD144	55,53	0,02	1,43	2,35	1,66	0,06	16,62	20,66	1,74	0,02	100,10
ZpD145	55,02	0,21	1,99	3,83	2,27	0,08	14,91	18,91	2,82	0,05	100,09
ZpD146	54,56	0,33	1,48	1,57	2,44	0,10	17,46	20,42	1,64	0,03	100,02
KOM	ISOMO	L'SKAY	A MAGN	ITNAY	PIPE						
KomMG01	54,68	0,197	1,90	0,58	4,42	0,091	19,66	15,20	1,61	0,026	98,36
KomMG03	53,88	0,253	1,42	1,90	2,67	0,065	17,21	19,29	1,63	0,042	98,36
KomMG04	54,56	0,270	1,81	2,31	2,43	0,076	16,64	18,47	2,10	0,037	98,71
KomMG11	52,07	0,295	5,18	2,04	2,30	0,105	20,79	18,06	0,69	0,000	99,46
KomMG12	55,35	0,213	2,00	0,51	4,07	0,126	19,77	15,40	1,60	0,036	99,04
KomMG13	54,88	0,309	1,45	1,01	3,14	0,125	17,99	18,37	1,37	0,029	98,67
KomMG16	54,54	0,267	0,83	1,32	2,19	0,067	18,07	21,35	0,83	0,007	99,46
NAK	YN FIE	LD									
NYU	RBINSK	AYA PI	PE				-				
CrDNu072	56,16	0,19	2,42	2,87	2,27		15,19	18,76	2,83	0,02	100,70
CrDNu059	55,39	0,21	2,13	3,93	2,41	0,20	14,22	17,40	3,07	0,03	98,99
CrDNu051	54,88	0,17	1,54	2,59	2,19	0,11	15,50	19,97	1,90	0,16	99,01
CrDNu035	54,06	0,30	1,33	2,38	2,53	0,13	17,11	18,55	1,94	0,05	98,38
CrDNu061	55,48	0,20	1,58	0,30	3,47	0,11	16,27	21,07	1,26	0,02	99,76
CrDNu042	55,48	0,40	1,60	1,53	2,68	0,15	17,26	19,96	1,68	0,19	100,92
CrDNu048	55,53	0,37	1,65	1,07	3,48	0,17	16,68	18,47	1,69	0,09	99,19
CrDNu029	54,80	0,31	2,04	2,39	2,14	0,10	15,19	19,85	2,23	0,03	99,08
CrDNu005	56,04	0,15	2,15	1,81	2,80	0,13	17,44	17,91	2,27	0,06	100,78
ORTH	IOPYRO	XENES									
CrDNu001	55,13	0,00	4,37	0,62	4,89	0,18	32,20	0,78	0,02	0,00	98,19
CrDNu043	54,93	0,13	3,73	0,68	4,69	0,11	33,59	0,80	0,02	0,07	98,75
BOTUOBI	<b>NSKAY</b>	A PIPE		-							
BoCp204	54,64	0,18	1,68	0,31	3,66	0,05	16,06	21,05	1,73	0,02	99,38
BoCp207	55,09	0,17	1,62	0,2	3,67	0,03	16,38	21,63	1,39	0,04	100,22
BoCp208	55,25	0,22	1,4	1,08	3,59	0,08	17,92	17,91	1,47	0,03	98,95
BoCp211	55,02	0,19	1,66	0,22	3,68	0,07	16,23	21,12	1,53	0,05	99,77
BoCp212	53,49	0,01	3,55	0,91	1,41	0	16,3	22,7	0,94	0,02	99,33
BoCp219	55,25	0,23	1,51	0,61	3,84	0,08	17,17	19,45	1,39	0,04	99,57
BoCp222	55,2	0,23	1,12	1,14	2,51	0,05	16,39	21,04	1,18	0,03	98,89
BoCp223	53,85	0,18	1,57	0,24	3,25	0,06	16,81	21,3	1,38	0,05	98,69
ΩΦΤΙ	ΙΛΡΥΡΛ	VENEC									
BoE202	54.95	0.00	4.18	0.60	4.78	0.11	34.12	0.50	0.03	0.01	99.28
BoE239	56.43	0.00	2.36	0.54	4.88	0.06	34.36	0.52	0.07	0.00	99.22
	20,15	0,00	-,50	0,0 .	.,00	0,00	1 2 ., 50	0,02	0,07	0,00	<i>,</i>

## Malo – Botuobinsky field

Reconstructed in details the mantle sequence beneath the Mir pipe [14, 15] based on the original data for minerals from concentrate (530) and data for about



Fig. 3. **TP estimates for the mantle columns beneath the kimberlite fields of the Malo-Botuobinsky region [67].** Signs are the same.

30 xenoliths was completed by the literature data [18,57]. Despite on the difference in the chemical features of the minerals from Mir and Internationalnaya pipe in general the structure of the lithosphere is close. (Fig.3). Highly depleted rocks are distributed beneath the Mi pipe in 60 - 45 kbar interval. In upper part lherzolites including the metasomatic minerals (phlogopite and amphibole) are more typical. The lens of eclogites and pyroxenites may be found there and the hot pyroxenites are found at the basement. In mantle section under the Internationalnaya pipe the heating to 1400oC is found from basement to 50 kbars. Three large peridotitic units it is possible to recognize in the lower part of the mantle section of both pipes. For Internationalnaya the lower units contain essentially eclogitic or hybrid peridotite. The spinel TP trajectories trace from the bottom the conductive geotherm but upper it is splitting and branching further to the top TP trajectories appeared in TP plot. The structure of the mantle beneath Dachnaya pipe should be very close to the Mir pipe. Spinel horizons are compiling there only conductive geotherm 35  $mv/m^2$  and at the level 35  $mv/m^2$  it show heating. Pyroxenes are giving there very LT conductive estimates with the clusters (fig.3).

The structure of the mantle lithosphere beneath Amakinskaya and Taezhnaya probably is close though the data are not sufficient. Chromite PT estimates refer to the middle part of the section and clinopyroxene to the basement of lithosphere.

The ilmenite trends for these two pipes refer to the lower part of section with gradient clothe to conductive marking metasomatites. For Mir pipe ilmenites give the TP conditions of HT metasomatites in the basement but upper as well as for Dachnaya pipe they are tracing the fractures of feeding system rising to the Moho with the clusters correspondent to the mantle layering. In the mantle column of the Internationalnaya pipe the ilmenites from the bottom show good correspondence to the conductive geotherms and from 50kbar the sub vertical sub adiabatic gradients.

# Upper Muna field

Detail TP diagrams were obtained for Deimos, Zapolyarnaya and Novinka pipe [9] and fewer representatives for Poiskovaya, Inter-Kosmos, Mechta, and Debyut. All this pipe reveal rather different TP sequence from the other region showing the presence of the enriched in clinopyroxene horizons in the basement of the lithosphere what is more typical for the Mesozoic kimberlites of South Africa and other parts of the post Gondwana continents [48, 49]. They reveal gentleinclined geotherms at the basement of the lithosphere. They are repeated tree or more times in the lower part of the mantle section (fig.4).

Three types of the mantle sections it is possible to determine there. Mantle column beneath Zapolyarnaya pipe contain wide range of the rocks of the metasomatic and pyroxenitic signatures within 65 - 30 kbar interval. The HT Ti-rich garnet branch is likely referred to the contact associations in the mantle column developed near the feeding rising protokimberlite systems. The separate

chains of the TP points probably correspond to the PT paths of the mantle melts. The ilmenite TP estimates are plotting sometimes as for Deimos on the three different TP trajectories regularly marking the starting of finishing parts of the geotherms referring to the different magmatic events or vein sys tem of different scale. HT clinopyroxene rocks are typical for the Novinka and Zapolyarnaya pipe in the lithosphere basement but are rather rare in Deimos.



Fig. 4. **TP estimates for the mantle columns beneath the kimberlite fields** of the Upper Muna region [7]. Signs are the same.

Inclined geotherms for the lithosphere beneath Poiskovaya and inter- Kosmos are tracing by the spinels; they are repeated three times in the lower mantel sequence. Ti – rich garnets widely developed there are more HT in the lower part of the mantle columns. In the upper part of mantle column near 30 kbar the HT branch close to SEAG [52] or to 60-65 mv/m<sup>2</sup> is outlined by several points. Ilmenite rocks are marking the separate discrete levels probably of the mantle metasomatites at the basement of the lithosphere near the pyroxenite lens (40 kbars) and Gar- Sp transition referring to the various TP gradients of the mantel diapers and magmatic systems.

### Nakyn field

Peridotite mantle beneath the Nakyn field is heated referring to the judging on the garnet TP estimates to geotherms 40 mv/m<sup>2</sup> (Fig.5) an possibly to 47 mv/m<sup>2</sup> for the mantle column of Botuobinskaya pipe if calculate the pressure using pyroxenite scheme. Pyroxene geotherms in the lower part are tracing the inclined convective branch which is reflect the heating of basement from 43 to 35 mv/m<sup>2</sup>, more evidently represented for the Botuobinskaya mantle sequence. In the mantle



### Fig. 5. **TP estimates for the mantle columns beneath the kimberlite fields** of the Nakyn region Signs are the same.

section beneath Nyurbinskaya pipe the pyroxene estimates from 55 to 40 kbar are tracing heating from 35 to 39 mv/m<sup>2</sup> geotherm. The main part of the spinel TP points are plotted in the middle part of the mantel section reflecting the inflection at the geotherm near 50 kbars. Lower part starting from this level is essentially heated what was probably the main reason of the high diamond potential. TP diagrams for both Nyurbinskaya and Botuobinskaya pipes are similar in the configuration but the later has more depleted mantle column what was the reason of the rarity f the pyroxenes and there low Al composition. Ilmenite TP estimates for Nyurbinskaya pipe is located within 60-50 kbar interval coinciding with the enrichment in the pyroxenes (fig.5).

### DISCUSSION

### Models of the mantle column evolution

Major conclusions and simple schemes of the evolutions were developed as a result of the study of monomineral thermobarometry and chemistry of the xenocrysts from 25 kimberlite pipes. Mantle dolmens in each region have their specific features though the concrete structure of the mantle column in the kimberlite field may vary in a high scale. Interaction of the deep seated melts and fluid s with the lithospheric mantle in each region has their specific features

For Daldyn field the contrast layering with the presence of the peridotite and eclogite slices with contrast mineralogy is typical for majority of pipes. [1, 5, 12, 13, 44, 46, 49]. Peridotites are moderately depleted containing low Na, Al, Cr content [11, 14, 46]. The Alakite field is characterized by the wide distribution of the metasomatism in the mantle column [6, 17, 19, 36, 49, and 55]. Mantle peridotites in the Malo- Botuobinsky region in the middle part of the section are characterized by the moderately enriched composition which is typical for continental mantle. Similar features were recognized for the Nakyn region wit the abundant eclogites and metasedimentary rocks [7]. Specific features in the Upper Muna field [9] are abundant metasomatites and moderately enriched characteristic of the pyroxenes at the deeper part of the mantle section [8, 13]. In the mantle column of Prianabarie the highly depleted and more Fe-rich [5, 17], with the wide distribution of the Phlogopite- Ilmenite metasomatites and local fertilization under the influence of the

Broad variation of the mantle column structure may occur in the same kimberlite field what probably is a result of the inclined position of the mantle layers or motley compositions of the paleo subduction slabs coupled in the lithospheric keel and probably local distribution of the zones of metasomatism in the zones of the fluid flames.

There is an impression that the composition of the mantle rocks in the basement seriously influences the degree of the metasomatic alteration and on the diamond potential. The pipes highly enriched in diamonds have thick depleted dunite horizon in the basement and are originally enriched in the subduction diamonds probably at the stage of the lithospheric craton growth. It is well known that dunites are beautiful conductors of the melts [23, 37], hence they should concentrate the subduction and then different plum melts including the stage of the kimberlitic magmatism. The melts flowing along the bottom of the lithosphere are choosing the dunites for rising and then penetrating in the upper part they create the series of he pyroxenites in the different levels and heat the mantle column what in the favorable FO2 and presence of the carbon in the rocks should bring to the growth of the good diamonds. Part of the diamonds are from the PT conditions refer to the pyroxenites lens with the more high heating degree. The composition of the basement determines also the type of the interaction. Interaction with dunites bring to the creation of impregnated peridotites of porhyroclastic peridotites. Intrusion of the melt brings to the shearing. Such a structures and rocks were found in the contact zones with the megacrystalline ilmenite rocks [50]. It case of the essentially pyroxenitic rocks or fertile peridotites the rocks brings to the HT magmatic substitution like found in the Thumb diatreme, Colorado [60] or to creation of hybrid low -Cr pyroxenites [48] which are typical for the Mir pipe and the South Africa and Angola and in our case pipe Dolgozhdannaya, Irelyahskaya.

Only Mir pipe from he fife studied pipes in the Malo – Botuobinsky region have judging of the frequency of the deep seated subcalcic garnets [55], have large

dunite lens in the basement. This is the largest diamond – bearing pipe in this region.

Large pipes also have steeper geothermal gradient sometimes close to the subadiabatic what probably is a result of the heating of the mantle columns by large portions of the rising melts. This heating g is a favorable factor for the diamond growth. The high developing of the pyroxenite association and the contact interaction is another one characteristic feature of large pipes.

The common feature of the mantle in Siberia and other World regions is the layering [16, 34] consisting from 10-12 slices. The thickness of the layers and other characteristics are very close for the lithospheric keel in the World. This conclusion was checked at least for the 40 pipes in Siberia. The exclusion is Prianabarie where the Mesozoic pipes are sampling the mantle column subjected to the influence of the Permo – Jurassic superplume, like in case of the mantle of the Aldan shield [17] where in the lower part of the mantle the Fe dunites with the Fe-Ti metasomatites are prevailing.

In some mantle column like in Alakite region the metasomatism penetrating through all the mantle columns are prevailing which are smoothing the contrast layering but the rising melts create at the boundaries the layers of the pyroxenites and metasomatites. The pyroxenite lens found practically in all the regions of the World at the level near ~40 kbar is the source region of the partial melting. The character of the metasomatism from K-type in the lower section to the Na type [2] upper probably is due to the decomposition of primary subduction amphibole on this layer. For example Mantle column beneath Yubileynaya pipe demonstrate high developing of the K – metasomatites with richterite and phlogopite including K – pyroxenes in the lower part of the mantle section and Na-pargasite in the lower part [55].

Large magmatic systems which are interacting with the continental lithosphere are referred to several types. H2O rich lamproite melts and H2O rich basaltic easy pass through the continental lithosphere to upper asthenospheric level where they are responsible for the heating to SEAG geothermal gradient and fertilization and additionally to Fe- metasomatism like in mantle columns beneath the Udachnaya pip and the diatremes of the Aldan shield [17]. Tholeitic plum melts which often became Mg-rich and as a result of the increasing of heating Adiabatic upwelling brings to essential depletion of the degree due to the lithospheric mantle if the whole volume. Protokimberlite and essentially carbonatite melts which appear as a result of the long differentiation and liquation course the heating mainly in the lower part of the mantle section up to 40 kbar to the 40-45  $mv/m^2$  and the Ti – metasomatism, this interaction possibly is a favorable factor of the diamond growth. Besides the high temperature komatiite melts that are responsible for the creation of the HT branches of geotherms 1600°C at the basement of the lithosphere.

### RESUME

The major models of the thermobarometric study are following. Mantle domens in each region have their specific features of the structure and interaction of the deep seated magmatic systems with the peridotite mantle. The primary structure of the lithosphere is forming at the subduction stage and in period of the compression of the lithosphere of the oceans and continents. The fragments of the mantle lithosphere in each geodynamic environment have their specific features with the close compositions of the rocks and minerals which leave their major characteristics after the coupling to the continental craton lithosphere. The melt and fluids flows of the subduction and plum nature may essentially change the geochemical features of the most permeable zones in the lithospheric keels of cratons. Usually mantle columns are separated on to two parts at the level 40 kbar by pyroxenite layer for due to dehydration of the subduction slabs near 120-130 km.

The most large pipes were formed in the most permeable zones in the lithosphere which usually are represented by the essentially dunite slices. The advection geothermal gradients are resulted partially due to the upwelling of the large magmatic masse in the continental keel what may be a favorable condition for the diamond growth [20].

The common feature for the Paleozoic kimberlites of Siberia is the layering consisting from 10-12 slices. For the region subjected to the influence of the Permian – Jurassic superplume in Anabar and Aldan shield. [17] is deep depletion and Fe- metasomatism of the mantle column lower 40 kbar.

Protokimberlite melts (which should be essentially carbonatitic at the last stages of the evolution are responsible) for the creation of the megacrystalline kimberlite associations at the stage of the creation of the feeding system which is accompanied by the Ti – metasomatism usually to the 40 kbar with the thermal gradient to 40-45 mv/m<sup>2</sup>, what may be one of the serious factor of the diamond growth.

### CONCLUSIONS

1. Using of the methods of the monomineral thermobarometry for 5 separate minerals show the agreement of the TP estimates and possibility to determine the general features of the mantle rocks of the mantle columns.

2. Ilmenite trends and associated Ti – metasomatites and contact associations are the result of the forming of the magmatic feeding systems at the stages preceding the eruptions. They are polybaric and were created in several pulses.

3. Mantle sections containing large amount of the eclogites have as a rule peridotites with the hybrid features and sometimes metasomatites and anatexic pyroxenites

4. Layered mantle structure is characteristic for the most of the kimberlite pipes and the circles of there creations are resulted from periodicity of the plum activity which cut the slabs

5. The largest diamondiferous pipes reveal the specific features of he mantle column structure- presence of the dunite horizons, sub adiabatic gradients and large amount of the pyroxenites in the mantle structure.

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# UDK 552.3/5+552.321.6 (571.56) Effects of the Yakutian plume on processes within the upper mantle of the Siberian Craton: geochemical data

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Ingenious data on distribution character of incompatible rare elements (Nb, Zr, Hf, Ti, Y, REE) in Cr-poor garnet megacrysts, in garnets and clinopyroxenes of hightemperature deformed peridotites and low-temperature granular peridotites from the Udachnava pipe kimberlites, are given. Geochemistry of HFSE and REE in Gnt megacrysts, in Gnt, Cpx from deformed peridotites best agree with hypothesis of asthenospheric melt percolation through asthenosphere and lower part of continental lithospheric plate. Origin of these melts is due to uplift of the Yakutian plume to the lithosphere basement. Geochemistry of HFSE and REE in Gnt and Cpx from lowtemperature granular peridotites suggests that lithospheric mantle of the Siberian Craton in the early Middle Paleozoic kimberlite cycle "was washed" by reduced fluids arriving from asthenospheric melt chambers. Arising fluids intensively extracted incompatible rare elements from rocks and minerals, enriched in them and left them on oxidizing barriers in reaction type metasomatites. Metasomatic diamond and graphite might be a product of this reducing process. The ancient olivineorthopyroxene cumulates complementary to high-pressure majorite-Mg-silicateperovskite cumulates buried in the upper mantle transition zone and enriched in HFSE represented a protolith substance of lithospheric mantle. Lithospheric protolith was enriched in LREE and depleted in HFSE, respectively.

# INTRODUCTION

The Upper Devonian kimberlite magmatism the most productive for diamonds is suggested to be related to the Yakutian plume [1]. It is supposed that the plume arrived to the base of rigid continental plate, produces asthenospheric melts, parent associations of Cr-poor megacrysts and realizes a specific magmatic substitution of substance in upper sections of asthenospheric layer and lower parts of lithospheric plate. These processes were demonstrated with geochemical characteristics of megacrysts and high-temperature deformed peridotites from kimberlites of South Africa and Yakutia [10, 11, 21]. B. Harte, C. Hawkesworth [15], S. Burgess, B. Harte [10] give evidence on refertilization of lithospheric base by asthenospheric melts. The paper [12] shows that zoning relative to major and rare elements in marginal garnet parts from low-temperature granular harzburgites in the Southern Africa kimberlite is due to the fluid agents affect shortly before their capture by kimberlites. It is developed in "Iherzolite" compositions of Gnt at the expense of those of "harzburgite". It can be supposed that the process is related to chemical fluid effect above the zone of direct percolation of asthenospheric liquids. At the same time all these problems are not finally resolved and invites further investigation.

The present work discusses data on geochemisty of series of incompatible rare elements (Nb, Zr, Hf, Ti, Y, REE) in Cr-poor garnet megacrysts, in garnets and clinopyroxenes of high-temperature deformed peridotites and low-temperature granular peridotites from the Udachnaya pipe kimberlites. The authors' aim is to show the distribution character of HFSE, Y, REE in Gnt and Cpx from deformed peridotites and in Gnt megacrysts, determine geochemical evolution of melts equilibrated with these minerals and discuss their possible nature. Hypothesis proposed previously for the mantle lithosphere reworking by reduced fluids arriving from asthenospheric melts at early stage of kimberlite-forming cycle is verified with geochemistry of minerals from granular peridotites [5].

### **INVESTIGATION METHODS**

All xenoliths studied and megacrysts were analyzed in detail in samples and thin sections. Gnt megacrysts as well Gnt and Cpx from deformed and granular peridotites in central and marginal parts of grains were analyzed for major oxides using X-ray microanalyzer "JXA-33" of the "Jeol" firm in the Geochemisty Institute of SB RAS (Irkutsk). Contents of rare elements in garnet and clinopyroxene are obtained using the method of secondary-ion spectrometry (SIMS) with microanalyzer "Cameca IMS ion probe" in the Institute of microelectronics of RAS ( Yaroslavl) using the method [3]. The later method provided the measurement accuracy of rare elements with concentrations > 0.1 g/t - 10-15 % and 40-50 % for concentrations < 0.1 g/t.

### **DEFORMED PERIDOTITES AND GNT MEGACRYSTS**

*Deformed peridotites*. Among studied xenoliths of deformed peridotites, four (samples 00-90, 02-49, 00-99 and 01-297a) are represented by Gnt lherzolites and two (00-112, 01-286) by Gnt harzburgites. Deformation degree of rocks is reflected in the structure and texture peculiarities, as well as in quantity of relic clasts of initial olivine in mosaic matrix from small olivine neoblasts. In compliance with early distinguished types [4], deformed peridotites are subdivided into coarse-porphyric and fine-porphyric varieties. Large garnet porphyroclasts (1-6 mm) and enrichment in garnet and clinopyroxene (to 25-30% of each mineral), along with relatively narrow (0.1-0.5 mm) kelyphitic rims on garnets are typical of coarse-porphyric type. Semi-cut small crystals of Cpx, Ol and sulfide globules are infrequently enclosed in garnet. Regular, partially cut, garnet grains and sulfide globules occur in clinopyroxene too. Garnet in coarse-porphyric type are brown-red and orange-red; their color is similar to that of megacrysts. Specific rocks enclosing

high-deformed Gnt, Cpx, Opx, Ol and Il megacrystals and representing, to all appearance, deformed megacryst cumulates [2] there are in this type composition. In studied collection two xenoliths belong to megacrystalline ones (00-92 – with 2 Gnt megacrystals 0.5 an 1.0 cm in size; 02-49 with 2 Gnt megacrystals 0.5-1.5 cm in size). Fine-porphyric garnet lherzolites are sharply depleted in Gnt ( usually less than 10%) and Cpx (less than 3-5%), besides that Gnt has wide (0.5-2 mm) opaque kelyphitic rims. Gnt in fine-porphyric type is cherry-red, winy-red and violet-red with specific metallic luster. Development of whitish porous rims on clinopyroxene grains occupying from 10 to 70% of their areas is typical of these rocks. Two megacrystalline lherzolites (00-92, 02-49), coarse-porphyric lherzolite (00-99) and fine-porphyric lherzolites and harzburgites (01-297a, 00-112, 01-286) are studied in present paper.

Microprobe studies shown that Gnt and Opx are homogenous within grains and exhibit zoning only in narrow edges (100-150 µ). Cpx changes distinctly its composition in whitish muddy rims and Ol shows, in isolated cases, differences in compositions of relic clasts and matrix neoblasts. Usually, TiO<sub>2</sub> and FeO compositions increase significantly and Cr<sub>2</sub>O<sub>3</sub> and mg# value decrease in narrow garnet rims. Cr<sub>2</sub>O<sub>3</sub> composition in Gnt sharply increases (from minimum 1.6-1.7% to 11.5% in Gnt harzburgite 00-112) and CaO composition increase notably when changing from megacrystalline and coarse-porphyric Gnt lherzolites to fineporphyric lherzolite 01-297a and Gnt harzburgites. Behavior of TiO2, FeO and Mg# value is less clear. For example, fine-porphyric Gnt lherzolite 01-297a and harzburgite 01-286 contain garnets with 0.1-0.2 % of TiO<sub>2</sub>, whereas TiO<sub>2</sub> content in high-chromium Gnt from Gnt harzburgite 00-112 reaches 1.2%. All garnets studied belong to lherzolotic paragenesis. As a rule, notable composition change of almost all oxides occur in marginal altered Cpx zones, but it is not of directional disposition. When discussing the compositional variation of central parts of Cpx grains successively from mega-coarse-crystalline to fine-porphyric Gnt lherzolite 01-297a one can note the decrease TiO<sub>2</sub>, Al<sub>2</sub>O<sub>3</sub>, FeO, Na<sub>2</sub>O contents and increase mg# value (in the range of 86.6-92.5). Behavior of the most part of major oxides when passing from central homogenous parts of Opx grains to narrow marginal zones does not exhibits distinct regularities. Orthopyroxenes from central grain parts become more magnesian and less Al-bearing during passage from mega- and coarse-porphyric lherzolites to fine-porphyric lherzolites and harzburgites. Ol in series from mega-coarse-porphyric lherzolites to fine-porphyric lherzolites and harzburgites reveals a significant FeO decrease and increase of mg# with maximum difference 86.6-91.6.

P-T parameters calculated with method [9] displayed the ranges for central grain parts: T=1398-1206°C and P = 80.5-49 kbar and for marginal zones: T=1371-1038°C, P=84-46.5 kbar.

*Garnet megacrysts.* Three garnet megacrysts from the Udachnaya Zapadnaya pipe and one from the Udachnaya Vostochnaya pipe (UZ 77-857, UZ 91-71, UZ 76-910 – growth with ilmenite, UV - 91-81). All megacrysts are represented by
smooth round and ellipsoid nodules from 1 to 3.5 cm in size. Megacrysts have a coarse cleavage which is cut by fine jointing at oblique angle. Small cavities filled with submicroscopic polymineral aggregate and representing evidently the former melt inclusions [5] occur usually in megacrysts. According to major oxides the Gnt megacrysts belong to low-chromium association that is common for world kimberlites. They have low  $Cr_2O_3$  (0.33-1.63%), high TiO<sub>2</sub> (0.56-1.05%) contents and rather wide range of mg# (67-86). Gnt megacrysts are chemically more similar to Gnt from deformed lherzolites 00-92, 00-99.



## Fig. 1. Spidergrams for REE+HFSE and REE in garnet from megacrystalline garnet lherzolites.

Microelement concentration in mineral is relative to C1 chondrite [24]. Sample numbers are indicated in plots. Hereinafter: solid symbols – central grain zones, open symbols – narrow marginal grain zone. Different symbols indicates different grains. Analysis points in marginal zone is located in 60  $\mu$ (large open circle) in 150  $\mu$  from kelyphitic rim (small open circle) in Gnt 00-92 grain.

*Distribution of rare elements in garnet from deformed peridotites.* Data on distribution of studied incompatible rare elements in Gnt from deformed peridotites are given on spidergrams (Fig. 1, 2). Contents of rare elements in large Gnt grain (0.5 cm) from Gnt lherzolite with quasi-angular Gnt megacrystals (00-92) are determined in two fresh blocks from relatively homogenous central parts and in two

points from narrow marginal zone near kelyfitic rim (60 and 150  $\mu$  from grain edge) – Fig. 1. Low relative LREE contents (La, Ce, Nd) and its consistent increasing to HREE (Yb) are characteristic of distribution of rare elements in two central blocks. Distribution lines have very characteristic stable peaks HFS elements – Nb, Zr+Hf, Ti that was previously marked for Gnt from megacrystalline Gnt lherzolite from the Udachnaya pipe [2].



#### Fig. 2. Above - spidergrams for REE+HFSE and REE in garnet from coarse-porphyric (00-99) and fine-porphyric (01-297a) lherzolites.

Microelement concentration in mineral is relative to C1 chondrite [24]. Analysis points in marginal zone is located in 60  $\mu$ (large open circle) in 350  $\mu$  from kelyphitic rim (small open circle) in Gnt 00-99 grain.

Below - spidergrams for REE+HFSE and REE in garnet from Gnt harzburgites: 00-112 – squares and circles; 01-286 – stars.

Visible distinctions for different blocks of the grain central part are established only for the most incompatible elements – Nb, La, Ce. Compositional lines of the grain marginal zones are shifted upwards in left part of plots especially for the point located directly on grain boundary. At the same time peaks for HFSE become distinctly smooth. The same distribution regularities of incompatible rare elements are shown in plots for two small Gnt grains from megacrystalline Gnt lherzolite with Cpx megacrystals (02-49). Different Gnt grains from this rock show noticeable differences in La and Ce contents and Nb, La, Ce contents significantly increase, Nb content remarkably decrease and max Zr+Hf are smoothed out in marginal part of one grain.

Fig. 2 represents relative contents of incompatible rare elements in garnet grains from coarse-porphyric lherzolite 00-99 and from fine-porphyric lherzolite 01-297a. Lines of rare elements distribution in Gnt fom sample 00-99 are completely similar to lines for central parts of Gnt from deformed Gnt lherzolites of megacrystalline type (Fig. 1). Line for narrow marginal zone of Gnt 00-99 shows significant increase of relative contents of rare elements in series from Nd to Yb and its feasible absence for Nb, La and Ce.

Fundamentally different distribution type is revealed in Gnt from poorly deformed fine-porphyric Gnt lherzolite 01-297a, namely: sinusoidal distribution character of REE, weak minima for Zr+Hf and Ti, level lowering of almost all elements from Nd to Yb (Fig. 2). Grain marginal zone does not show change in rare element composition as compared to central one. Distribution type of rare elements in Gnt from Gnt harzburgites 00-112 and 01-286 is in general similar to that in Gnt from fine-porphyric Gnt lherzolite 01-297 a (Fig. 2). Gnt from poorly deformed rock 00-286 does not reveal essential distinctions between two different grains. Content level of rare elements and its distribution character in Gnt from this rock are very similar to those for Gnt from poorly deformed fine-porphyric Gnt lherzolie 01-297 a. Gnt from highly deformed Gnt harzburgite 00-112 has a significantly higher level of rare elements contents in series from Nd to Yb, at the same time keeping sinusoidal form of curve for REE. Narrow marginal zones of two Gnt grains in this sample show essential content increase of elements in series from Zr to Yb and some approach to distributions typical of coarse-porphyric and megacrystalline deformed Gnt lherzolites.

*Distribution of rare elements in garnet megacrysts.* Distribution of rare elements in four Gnt megacrysts (Fig. 3) both as to content level and curve type are similar to Gnt from deformed megacrystalline and coarse-porphyric Gnt lherzolites. There is certain difference in absence of max Ti for Gnt megacrysts except for Gnt UV 91-81. On line of Gnt UZ 76-910 (garnet growth with ilmenite) there is a distinct Ti minimum. Gnt megacrysts exhibit great similarity with zoning in Gnt from samples 00-92 and 02-49 for the most incompatible elements among Nb-Eu. As distinct from zoning Gnt megacrysts reveal significantly more wide range as to HREE (Gd-Yb) contents. At the same time for Gnt megacrysts there is a clear

correlation between HREE content and ferruginosity of mineral. Line of Gnt UZ 76-910 traverse a parallel line system of remaining megacrysts that apparently indicates the affect of Gnt co-crystallization with ilmenite.



Fig. 3. Spidergrams for REE+HFSE and REE in garnet megacrysts.

Microelement concentration in mineral is relative to C1 chondrite [24]. Megacrysts: UZ 77-857 (mg# =0.86) – solid circle; UV 91-81 (mg# = 0.825) – solid rhomb; UZ – 91-71 (mg# = 0.74) – open square; UZ 76-910 (mg# =0.67) – solid square.

*Clinopyroxene from deformed lherzolites*. Sufficiently similar distribution type is revealed for all Cpx studied: min for Nb, Zr+Hf and Ti and convex curves in the region of LREE (Fig. 4, 5). Cpx curves from fine-porphyric lherzolite 01-297a 01-297a have a more sharp slope from Nd to Er on REE plot and the most deep Zr+Hf trough. Rare element Cpx composition from pure large megacrystal block and the small grain center in sample 02-49 is identical. Content of all rare elements especially the most incompatible (Nb, La, Ce, Nd) and the least compatible (Dy-Yb) increases in decolorized marginal zone of small Cpx grain (02-49). Nb and La contents slightly increase and those of Er, Yb decrease in marginal zone of Cpx from megacrystalline Gnt lherzolite 00-92.

Calculated melts equilibrated with Gnt megacrysts and Gnt and Cpx from deformed peridotites. Melt compositions equilibrated with Gnt and Cpx were calculated for evaluation of possible compositions and sources of melts that gave rise to megacryst associations and transformed the substance at lithosphereasthenosphere boundary during kimberlite-forming cycle. Accordingly to conceptions [6] it is suggested that conditions of local mineral/melt equilibration realized in asthenospheric part and lithospheric plate basement during this period. Calculation was made by formula: CL+ Cmin/K, where CL – element concentration in melt; Cmin - element concentration in mineral; K – element distribution coefficient between mineral and melt.

The following distribution coefficients were taken for calculations – KGnt/L: for Nb – 0.005; La – 0.0011; Ce – 0.008; Nd – 0.049; Zr – 0.25; Hf – 0.23; Sm – 0.21; Eu – 0.345; Ti – 0.1; Gd – 0.54; Dy –1.592; Y – 2.5; Er – 2.702; Yb – 4. KGnt/L for Nb, Zr, Hf, Ti, Yb by [26]; for La, Ce, Nd, Sm, Eu, Dy, Er [ 10 at 1300°C]; for Gd by [27]; for Y by [13].



## Fig. 4. Spidergrams for REE+HFSE and REE in clinopyroxene from megacrystalline garnet lherzolites.

Microelement concentration in mineral is relative to C1 chondrite [24]. In Cpx 00-92 grain, composition is measured in narrow marginal zone (50-80  $\mu$ ) – large open circle and in pure transparent block inside wide colorless rim – small open circle. In sample 02-49, Cpx composition is measured in large pure block from megacrystal (solid circle), in unaltered central part of small grain (solid circle) and in turbid marginal zone – in 100  $\mu$  from this grain boundary (open circle).

KCpx/L: for Nb – 0.05; La – 0.061; Ce – 0.092; Nb – 0.199; Zr – 0.233; Hf – 0.2; Sm – 0.276; Eu – 0.31; Ti – 0.384; Gd – 0.44; Dy – 0.386; Y – 0.467; Er – 0.344; Yb – 0.43. KCpx/L for Nb, Zr, Hf by [26]; for La, Ce, Nd, Sm, Eu, Dy, Er, Yb by [10]; for Gd by [13]; for Ti, Y, by [14].

Rare elements contents relative to chondrite in calculated melts for different zones in Gnt grains from deformed megacrystalline Gnt lherzolites 00-92 and 02-49 are given in Fig. 6. Melt compositions equilibrated with megacrysts as well as with central zones of Gnt grains from all studied deformed peridotites are shown here too. L calculated contents for zonal Gnt 00-92 and 02-49 show enrichment of external narrow zones in the most incompatible rare elements, especially in Nb, La, Ce sharply expressed in the most external zone of Gnt 00-92. This tendency is also well traced for Gnt megacrysts. Gnt megacrysts enrichment in all rare elements is an essential difference L compositions equilibrated with megacrysts. This enrichment of L Gnt megacrysts increases with Gnt ferrusinosity increasing that may correspond to fractional crystallization trends and decreasing of temperature.



#### Fig. 5. Spidergrams for REE+HFSE and REE in clinopyroxene from coarse-porphyric (00-99) and fine-porphyric (01-297a) garnet lherzolites.

Microelement concentration in mineral is relative to C1 chondrite [24]. Cpx – circles; Cpx 01-297a – triangles. In Cpx 00-99 the composition is measured in wide colorless turbid rim – in 70-100  $\mu$  from grain edge (large open circle). In Cpx 01-297 a the composition is measured in wide colorless turbid rim – in 50-70  $\mu$  from grain edge (large open triangle).

L Gnt lines of the most ferruginous Gnt megacrysts UZ 76-910 traverse L Gnt lines of other megacrysts that may be related to effect of complimentary factor besides of mineral fractionating. Maxima for HFSE (Nb, Zr+Hf, Ti) on distribution curves of incompatible rare elements are common feature for all melts equilibrated with Gnt megacrysts and Gnt from megacrystalline and coarse-porphyric lherzolites. These maxima are practically not occur for Nb, Zr+Hf on lines of melt equilibrated with narrow marginal zones of Gnt 00-92 and 02-49. Distinct Zr+Hf minima occur on Gnt melt lines from central grain parts of Gnt harzburgites 00-112 and 01-286 and

the lines themselves show significantly greater slope from Nb to Yb. Line of melt equilibrated with Gnt 00-297a (fine-porphyric lherzolite) except for Ti peak is similar to L lines from Gnt harzburgites.

Contents of L melts equilibrated with Cpx from deformed Gnt lherzolites are given in Fig. 7. Troughs for HFSE the least pronounced for Ti are typical of all



#### Fig. 6. REE+HFSE and REE spidergrams for calculated melt compositions (L Gnt) equilibrated with garnet from deformed peridotites and megacrysts (vd. explanations in text).

Microelement concentration in mineral is relative to C1 chondrite [24]. For L Gnt 00-92 and L Gnt 02-49 - 1 legends are in Fig. 3. Symbols of melts equilibrated with central zone of garnet from deformed peridotites (L Gnt-c): 00-92 - 100-92 -

lines. This feature was previously recorded for Cpx from megacrystalline Gnt lherzolite from the Udachnaya pipe [2]. There is no Nb trough and that of Zr+Hf is shown significantly slighter on L line of marginal zone of Cpx 02-49 grain.

Increase of Nb, La, Ce and Dy-Yb concentrations is also recorded for this melt.



#### Fig. 7. **REE+HFSE and REE spidergrams for calculated melt compositions equilibrated** with clinopyroxene from deformed lherzolites (L Cpx).

Microelement concentration in mineral is relative to C1 chondrite [24]. All melts is calculated for central unaltered part of Cpx grains (solid symbols) except for marginal grain zone Cpx 02-49 (open square). L Cpx symbols for central grain parts: 00-92 – circle; 02-49 –square; 00-99 – rhomb; 01-297a – star.

Zr+Hf trough is sharply expressed and fractioning of rare elements from La to Y is shown stronger on L line of Cpx 01-297a (fine-porphyric Gnt lherzolite).

#### **GRANULAR PERIDOTITES**

A special series of granular peridotites with different olivine color was selected in order to verify the hypothesis for "washing" of lithospheric mantle by reduced fluids flows [5] deriving from asthenospheric melts during kimberlite-forming cycle. The Mössbauer study of polychrome olivine from granular xenoliths is shown that orange, brown and yellow olivines contain noticeable Fe<sup>+++</sup> admixture, whereas it is almost completely absent in pale-green varieties [22]. This

phenomenon was explained by irregular reworking of the mantle lithosphere substance by reduced fluids not long before the capture of xenoliths by kimberlite melt. Xenoliths with orange, brownish and yellow olivines are extremely rare in population of granular peridotites from the Udachnaya pipe. Xenoliths with polychrome olivine occur more often, whereas olivine is green, pale-green or colorless in sharply predominant xenoliths quantities. Garnet color of peridotite changes with olivine chromacity: garnet reddish-violet, pink-violet and violet in xenoliths with orange and polychrome olivine and garnet is of bluish, lilac and ink tints.



## Fig. 8. HFSE (Nb, Zr, Hf, Ti), Y+REE and REE compositions relative to chondrite [24] in garnet and clinopyroxene from granular garnet peridotites with different olivine color.

Symbols for samples: 1 - 343/87; 2 - 545/80; 3 - 45/82; 4 - 50/82; 5 - 325/87; 6 - 544/80; 7 - 42/82; 8 - 43/82; 9 - 00/212. Thick lines – samples of group1, dashed lines – group 2, and thin lines – group 3.

Three control groups of granular garnet peridotites (lherzolites and harzburgites) from the Udachnaya pipe were studied: 1) xenoliths with 50-90% of orange, brownish-pink and yellow olivine from its total bulk (343/87, 545/80, 45/82); 2) with polychrome olivine in which orange, brownish-pink nuclei and

grains occupy 10-25% and remaining body of mineral is represented by green and colorless varieties (50/82, 327/87); 3) with green, pale-green and colorless olivine among which there are sometimes up to 3-10% of yellowish-gray olivines (544/80, 42/82, 43/82, 00/212, 03/108). Three xenoliths with  $\text{Fe}^{+++}$  contents in olivine calculated using the Mössbauer method of spectroscopy entered into series composition [22 Sol98].

Studied garnet lherzolites and harzburgites from the Udachnaya pipe are granular peridotites of common type and have no any features of cataclastic structures typical of deformed peridotite xenoliths from kimberlites. Model mineral composition for major paragenesis of minerals, percentage compositions of orange, brownish and yellow olivines from total olivine quantity in rock and character of its secondary changes:

343/87 – garnet lherzolite: Phl I – single plate, 10.7% Gr, 3% Cpx, 20% Opx, 69% Ol (90% Ol is represented by pinkish-brown and yellow-gray varieties). By data [8] Fe<sup>+3</sup>/Fe<sub>tot</sub> is equal to 0.025 in pinkish-brown olivine. Xenolith is serpentinized by 5-7%.

 $545/80^*$  - garnet harzburgite: 3-7% Gr, 5-10% Opx, 85% Ol (approximately 50% of olivine is represented by orange and gray-yellow varieties). Fe<sup>+3</sup>/Fe<sub>tot</sub> is equal to 0.034 [8] in orange olivine. Xenolith is serpentinized by 59-60%.

45/82 - garnet lherzolite: 1, 6% Gr, 8% Cpx, 24.2% Opx, 66.2% Ol (approximately 50% of orange and greenish-yellow varieties). Xenolith contains no more than 1-3% of secondary minerals of late metasomatic paragenesis (kelyphite, phlogopite II, emphibole, calcite, ore mineral).

 $50/82^*$  - garnet lherzolite: 13% Gr, 10% Cpx, 22% Opx, 55% Ol (10% of orange and 40% of yellowish-green varities), the rock contain no more than 1-3% of secondary minerals.

325/87 - garnet lherzolite: 18% Gr, 2% Cpx, 29% Opx, 51% Ol (25% of orange, pink-brown varieties). Fe<sup>+3</sup>/Fe<sub>tot</sub> is equal to 0.020 in orange and 0.008% in green olivine. Rock is serpentinized by 25%.

544/80\* - garnet lherzolite: 8% Gr, < 1% Cpx, 15% Opx, 75% Ol (total olivine is represented by colorless variety); Rock is serpentinized by 50-60%.

42/82 - garnet lherzolite: 6.5% Gr, 2.5% Cpx, 23% Opx, 68% Ol (10% of pale-orange grains, remaining olivine – pale-green and colorless), rock contains no more than 1-3% of secondary minerals.

43/82 - garnet lherzolite: 3.4% Gr, 3% Cpx, 18.9% Opx, 74.7% Ol (total Ol is represented by pale-green and colorless varieties), rock contains no more than 1-3% of secondary minerals.

00/112 - garnet lherzolite: 6.8% Gr, 0.3% Cpx, 25% Opx, 67.9% Ol (5-10% of yellowish grains, the rest of olivine is pale-green and colorless), rock is serpentinized by 10-15%.

03/108\* - garnet harzburgite: 12% Gr, 18% Opx, 70% Ol (olivine is green, pale-green and colorless).

Mineral ratios in rocks in samples with\* were determined approximately by lines in large sample and in the rest of samples they are calculated by grain ratios in crushed samples. Single strongly deformed orthopyroxene megacrystals from 0.7 to 2 cm in size are present in samples 343/87 and 00/212. There is a negligible amount (< 0.5%) of irregular intergranular Al-Cr-spinel exsolutions frequently in myrmekitic growths with clinopyroxene in the most part of garnet lherzolites. From the mineral ratios one can say that Al-Cr-spinel developed after the main paragenesis of minerals and before metasomatic prekimberlite paragenesis, by [5].

Comparison of garnet chemistry in major oxides does not show sharp differences in three distinguished groups. It is recorded only increased FeO content in sample 45/82 (group 1) as well as systematically higher CaO content in garnets of  $3^{rd}$  group. Unlike garnet the clinopyroxene contains consistent increasing of Al<sub>2</sub>O<sub>3</sub>, Cr<sub>2</sub>O<sub>3</sub>, FeO and Na<sub>2</sub>O contents during transition from first to third xenoliths group that may be possibly due to Al-Cr-spinel development. P-T parameters calculated by method [9] are typical of low-temperature granular peridotites: 670-970°C and 33-45 kb.

Distribution of rare elements in garnet and clinopyroxene from granular peridotites. It is established that in garnets as well as in clinopyroxenes the contents of all incompatible rare elements studied decrease in series from rocks with orange and yellow olivines to rocks with pale-green and colorless olivines. Decrease of rare elements concentrations occur extremely irregularly: Nb and Yb content in garnet from different groups is practically remained at the same level whereas LREE content especially Nd and Sm change by two-three orders (Fig. 1). Reduction of all rare elements, maximum also for Nd and Sm (more than 10 000 times) occur in clinopyroxene. Curve forms on spidergrams change radically too instead of troughs for Nb, Zr+Hf and Ti, weak peaks of theses elements sometimes occur on clinopyroxene lines from rocks of group 1 and on lines of mineral from rocks of group 3. Troughs of Zr+Hf and Ti typical of garnets from group 1 rocks (with orange and yellow olivines) are usually absent in garnets from group 3 rocks ( with green and colorless olivine) and weak peaks of these elements occur in place of them. Lines of minerals from xenoliths of group occupy intermediate position between lines of minerals from groups 1 and 3, at that, distribution character of rare elements in Gr 50/82, Cpx 50/82, Cpx 325/87 is similar to group 1 minerals, while Gr 325-87 are closer to group 3 garnets by level and distribution type. Garnets of group 1 and Gr 50/82 of group 2 show sinusoidal curves typical of garnets from granular peridotites [8, 12] on REE plot.

#### DISCUSSION

Behavior characteristics of incompatible rare elements in Gnt and Cpx from deformed peridotites and in Gnt megacrysts from the Udachnaya pipe given in previous section are considered on basis of B. Harte et al. [16] mechanism: melt percolation through solid rock matrix with simultaneous fractional crystallization. Megacrystalline and coarse-porphyric lherzolites represented apparently high-pressure megacryst cumulates before deformation, as indicated by preservation of deformed Gnt, Cpx and Ol megacrystals in rocks, as well as almost complete similarity in rare element distribution in Gnt to that in Gnt megacrysts (Fig. 1-3). Gnt fractioning results in accumulation in residual melt of the most incompatible elements including LREE and in its depletion in HREE [10]. Cpx, Opx and Ol fractioning results in gradual accumulation of all REE [10] and possibly of HFSE in magmatic liquid and ilmenite crystallization – in its sharp depletion in Nb and Ti and to a lesser degree in Zr and Hf [2].

In this context the accumulation of most incompatible elements in marginal zones of Gnt 00-92 and 02-49 may be resulted from crystallization and garnet removal from melt (Fig. 1). At the same time more compatible rare elements in series studied do not depleted marginal zones as would be expected during Gnt fractioning that may be explained in terms of example of parallel exsolution of other silicate minerals from melt. On the contrary, narrow marginal grain zone in Gnt 00-99 is distinctly enriched in elements in series from Nd to Yb, that possibly can be related to the fact that accumulation of these elements is predominantly due to crystallization of other silicates. Enrichment lack of marginal zone of this Gnt in the most incompatible rare elements in series of Nb-Nd could resulted neither from silicate fractioning including Gr nor chromatographic effect and possibly is due to previous distillation of fluids. Reverse correlation of rare element composition with Gnt mg# is revealed for Gnt megacrysts and melts equilibrated with them (Fig. 3, 6). Most ferriferous Gnt megacrysts and melts equilibrated with them have the highest HREE contents hence excepting Gnt fractioning role in melt evolution. Apparently, contamination with lithospheric material also effect on rare element composition of asthenospheric melts. From this point of view the melt equilibrated with the most magnesian megacryst UZ 77-857 (mg#=0.86) is the most contaminated. Depletion of melt equilibrated with most ferriferous megacryst of Gnt Uz 76-910, light REE might be most likely resulted from previous removal of LREE-rich volatile phase.

It is significant that melts equilibrated with garnets from deformed lherzolites of coarse-porphyric and megacrystalline type and with Gnt megacrysts are enriched in HFSE, which show maxima on distribution lines. In order to explain HFSE maxima against REE in asthenospheric liquids equilibrated with Gnt from coarseporphyric deformed lherzolites and megacrysts one can draw the ancient oceanic crust submerged in transition zone or at lower mantle-core boundary as a melt source [19]. Hypothesis for the presence of cratons within oceanic plate accreted continental lithosphere as a result of the ancient underplating beneath lithosphere is a variety of hypothesis for buried ancient oceanic plate [25]. Enrichment of asthenospheric melts in HFSE against REE in analyzed case, may be related to refractory minerals – concentrators of HFS elements which are preserved in ancient oceanic melt [24]. It should be saying, that in spite of all attractiveness of R.Walker's et al. idea, none authentic xenolith of deformed eclogites that can testify the presence of ancient oceanic crust beneath lithospheric basement of the Siberian craton is to present day found among deformed xenoliths.

Uniform "mantle" isotopic oxygen composition in group I kimberlites and their Gnt megacrysts suggests that oceanic crust modified by sea water was not present in magmatic source [23]. From data [20, 25] minerals from deformed peridotites have negative  $\gamma$ Os values while this value is highly positive in eclogites from the Udachnaya pipe (about from 500 to 7000). Thus, even a negligible amounts of eclogites that may represent the ancient oceanic crust added to source material will sharply displace  $\gamma$ Os value in asthenospheric melts to positive values. These all data unlikely suggest that source material for asthenospheric liquids gave rise to Cr-poor megacrysts associations and transformed the mantle material in lithospheric plate basement and in asthenospheric upper part was ancient oceanic crust buried in deep mantle.

From rare elements distribution in komatiites and basalts of different ages Q. Xie et al. [26] argue the presence of layer enriched in majorite and silicateperovskite in the mantle transition zone. It is the authors' opinion that this layer preserved in mantle after crystallization and of high-pressure silicates in ancient Pre-Archaean magmatic ocean. It is significant that these mineral accumulate HFSE predominantly with REE in themselves [19, 26]. Ancient cumulative material isolated in the mantle transition layer arises to continental lithosphere basement during kimberlite-forming cycle in the form of giant plume[10]. Maximum T of Cr-poor megacrysts (1470°C, [23]) meets exactly this layer.

Incompatible rare elements distribution in Gnt from fine-porphyric Gnt lherzolite 01-297a and Gnt harzburgies 00-112 and 01-286 (Fig. 2) is most authentically explained by chromatographic effect during asthenospheric liquids percolation upwards through thick solid rocks of lithosphere [10]. Garnets from the Siberian samples likewise to central parts of some garnet grains from deformed peridotites of Southern Africa [10] have sinusoidal REE distribution that is typical both of garnet from granular peridotites of Siberia and Southern Africa [8, 12]. One more characteristic feature is distinguished, namely garnets from poorly deformed rocks (01-297a and 01-286) have more lower level of rare elements contents in series for Nd-Yb in comparison with strongly deformed xenolith 00-112 and well expressed Zr+Hf and Ti troughs, too. Zr+Hf minima and more differential rare elements distribution are revealed for L Gnt lines. All these characteristics can be apparently related to preferential accumulation of the most incompatible rare elements during the melt transport when the rock solid matrix absorbs predominantly the most compatible elements [16]. And at the same time poorly deformed rocks essentially preserved the features of initial granular protolith. External narrow zones of two Gnt 00-112 grains show some approaching to distributions in Gnt megacrysts and megacrystalline coarse-porphyric lherzolites that suggests their approaching to equilibration with asthenospheric melts.

Distribution peculiarities of rare elements in Cpx from deformed Gnt lherzolites 00-92, 02-49, 00-99 (Fig. 4, 5) are similar to those for Cpx from megacrystalline Gnt lherzolite described by K. Egorov et al. [2]. Contrary to garnets, HFSE troughs the most pronounced for Zr+Hf are typical of Cpx. Melts equilibrated with Cpx of these xenoliths except for Cpx 02-49 marginal zone also show HFSE troughs especially sharp for Zr+Hf (Fig. 7). Moreover, LCpx in comparison with LGnt from appropriate samples are essentially depleted in LREE that is possibly due to considerable fluid loss by melts before mass clinopyroxene crystallization. It can be inferred that clinopyroxenes were in equilibrium with liquids from which the solid phase like ilmenite and zircon were exsoluted [2]. Marginal zone of Cpx 02-49 might be equilibrium with residual liquid enriched in the most incompatible elements. Cpx from fine-porphyric lherzolite 01-297a have a lesser level of all rare element composition except for Nb that it is also recorded for calculated melt (Fig. 5, 7). Furthermore, a raise from Y to Yb atypical of other xenoliths clinopyroxenes is recorded for Cpx and LCpx curves of this sample. Apparently, it can be explained by relic geochemical features of rare element composition of primary solid rock remained in mineral.

It can be assumed that plume action consists not only in supply of heat and volatiles invoking melting but in rather long-term mechanical effect – pressure on overlying mantle layers. Thick cataclastic processes and plastic rock flows in asthenosphere and in the lithospheric plate basement saturated with melts could arise owing to viscous friction when spreading of plume substance in asthenosphere.

As discussed earlier, composition level of rare elements in garnet and clinopyroxene from low-temperature granular peridotites and curve type in plots of their relative compositions are not consistent with amounts of these minerals or orthopyroxene in rocks but they are directly resulted from composition of orange, brown and yellow olivine. For instance, garnet harzburgites 545/80 (gr. 1) and 03/108 (gr. 3) similar to modal mineral composition differ distinctly in composition and rare element distribution in compliance with color varieties of olivine. It was indicated earlier [22] that orange, brown and yellow olivine has a significant admixture of oxidized iron and it was preserved as relic one when transporting through solid rocks of reduced fluids related to early stage of kimberlite-forming cycle. Formation of native graphite in mantle metasomatites of reaction type [6] is probably due exactly to these fluids. Here, it should be emphasized that on data of

some researchers [5, 7, 18], Cr-poor megacrysts and deformed peridotites were formed under substantially oxidized conditions (QFM buffer). There is good reason to believe that the development of asthenospheric melts beneath cratonal lithosphere and reworking of their overlying horizons by reduced fluids took place within single kimberlite-forming cycle and they are complementary. Reduced fluid character can be explained by a head front of hydrogene in asthenospheric melts by [5]. These fluids extracted intensively incompatible rare elements from rocks and minerals and apparently realized the reaction type metasomatism [5, 6] at geochemical oxidized barriers. Complementary essence of processes in asthenosphere and overlying lithosphere is shown both in different distribution character of incompatible rare elements in megacrysts garnets and mega- coarseporphyric lherzolites and in garnet from granular peridotites of group I poorly reworked by reduced fluids. In spite of the former, the later show HFSE troughs against REE on distribution lines. This can suggest that ancient high-temperature orthopyroxene and olivine cumulates were a protolith of lithospheric mantle substance. Parent melt was impoverished in HFSE at the expense of precipitation of high-pressure majorite garnet and Mg-silicate perovskite and strongly enriched in LREE rare elements. Accordingly, cumulate opthopyroxene of mixed type was enriched in LREE and depleted in HFSE. Subsequently, during removal of exsolution structures from cumulate orthopyroxene, these minerals redistributed rare elements according to DGr/Cpx value.

From this it is inferred that representatives of primary substance of lithospheric mantle of the Siberian Craton are not rocks and minerals depleted in incompatible rare elements but on the contrary, those enriched in them. This inference causes us to give a new glance at initial nature of lithospheric mantle of ancient cratons. If ancient lithospheric mantle was primarily enriched in incompatible rare elements, it is not considered to be a restite of basaltic and komatiitic liquids melting as it is believed by majority of researchers until now [8, 12 et al.]. In our example, ancient enrichment of the restite mantle in incompatible rare elements at the expanse of introduction of subduction silicic melts [17] is not consistent with absence of direct relation between content of rare elements in garnet and clinopyroxenes and quantity of these minerals and orthopyroxene in rock. Garnet and clinopyroxene from lherzolite 00/212 (3 gr.) are depleted in rare elements in direct conformity with pale-green and colorless olivine, whilst the xenolith is enriched in orthopyroxene and contains a few relic deformed megacrystals of this mineral. If the later were ancient porphyroblasts originated during effect of subduction acid melts, minerals of this rock would preserved the rare element enrichments and not the reverse.

#### CONCLUSIONS

Geochemical characteristics of Gnt and Cpx from deformed peridotites shown that among this rock group there are two stable types – mega- coarse-crystalline Gnt lherzolites and fine-porphyric Gnt peridotite (lherzolites, harzburgites). Garnets from deformed lherzolites of the first type are nearly completely identical to Gnt megacrysts from the Udachnaya pipe in geochemistry of rare elements studied. Gnt from the second type peridotites have residual Gnt geochemical features from granular Gnt lherzolites from this type and kimberlites of the Southern Africa.

The data obtained are the most logically consistent with hypothesis of asthenospheric melt percolation upwards through the asthenosphere and continental lithospheric plate basement [16] in HFSE and REE distribution character in Gnt, Cpx and equilibrated with them calculated melts. The mechanism of chromatographic enrichment of minerals and melts with the most incompatible elements is, apparently, the most effective during formation of Gnt and Cpx rare element composition in fine-porphyric type of deformed peridotites.

The source of asthenospheric liquids producing a specific magmatic metasomatism at the continental lithosphere-asthenosphere boundary, along with local substance might be matter of plume initiating these processes.

Plume substance has geochemical features indicating both its origin from transition mantle zone enriched in majorite and silicate-perovskite and feasible existence of magmatic ocean in early history of our planet.

Lithospheric mantle of the Siberian Craton in early period of the Middle Paleozoic kimberlite cycle was "washed" by reduced fluids deriving from asthenospheric melt chambers. Uplifting fluids extracted intensively incompatible rare elements from rocks and minerals, enriched in these minerals and left them on oxidized barriers in reaction type metasomatites. Metasomatic graphite and diamond might be a product of this reduced process.

Light olivine-orthopyroxene cumulates crystallized in ancient magmatic ocean and complementary to high-pressure majorite-Mg-silicate-perovskite cumulates buried in the mantle transition zone and enriched in HFSE, were a protolithic substance of lithospheric mantle. Lithospheric protolith was enriched in LREE and LIL and depleted in HFSE, respectively.

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#### UDK 551.22 +552.11 VARIATIONS OF THE OXYGEN CONDITIONS IN MANTLE COLUMN BENEATH SIBERIAN KIMBERLITE PIPES AND IT'S APPLICATION TO LITHOSPHERIC STRUCTURE OF FEEDING SYSTEMS

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TP- fO2 were determined with the monomineral thermobaro- and oxybarometry for ilmenites and Cr-spinels (Ashchepkov, Vishnykova , 2005), The ilmenite method give the conditions for the megcrystalline veins served as the feeding systems for the protokimberlite melts for metasomatites and pyroxenites. Chromite thermobarometry shows the conditions for peridotites. Every studied region or kimberlite field or even kimberlite pipe reveal distinct features of the construction magmatic systems and mantle column layering.

Long compositional trends with TP values tracing interval from 20 to 65 kbar are common type determined for Mir and Dachnaya pipes representing the magmatic fractionation without the essential interaction with the wall rocks in the long vertical fracture. Rapid rise of the oxygen fugacity refer to the crystallization without exchange. Intensive exchange with the wall rocks shows the high variation of fO2 and the Cr-Ni-V components in the oxide minerals.

For Daldynsky region (Udachnaya, Dolgozhdannaya, Aeromagnitnaya show relatively short intervals (65- 50 kbar) while for Zarnitza , Leningradskaya, Osennyaya and Ukrainskaya 65-40 kbar. Cr content rising due to contamination. Layered structure with the high variations of fO2 is determined for every mantle columns in Daldyn region.

In Alakite field ilmenite from most of the pipes display the long compositional ranges with stepwise rising Cr content and melt upwelling. Three major levels of fractionation are determined for Sytykanskaya, Aykhal and Komsomolskaya and more for Yubileynaya. t P- fO2 trends found for Cr spinels show two branches in each horizons. One with continuous fO2 growth joining with the ilmenite trend is formed due to anatexic melt percolation while the other one is referring the conditions of the pervasive metasomatism.

For the MaloBotuobinsky region long compositional trend without rising of Cr content means tracing by crystallizing protokimberlites in rising fracture. The others including Internationalnaya, Taezhnaya, Amakinskaya mantle columns shows the stop in developing of feeding system at 40 kbar with further creation of metasomatites in veinlets. Peridotitic chromites also show the sharp divide at the 40 kbar with wide fO2 ranges. While rising of fO2 below this pyroxenites lens suggests high degree interaction with small relic trend.

Long continuous ilmenite trends are not characteristic for the Verhne –Munsky region – only discrete relics of the trend are found in concentrated from Poiskovaya, Debyut, Malaya, and InterKosmos. Only Deimos and Komsomolskaya- Magnitnaya pipes reveal long compositional ranges with three separate PT levels and two stages of contamination similar to the Alakite pipe. But chromite highly enriched in the ulvospinel show the straight rising of fO2 wit the decreasing pressure 2-3 such lines what possibly means the direct migration of the protokimberlite melts.

In Nakyn field only Nyurbinskaya pipe show crystallization of protokimberlite melt within 60-50 kbar and rising of fO2 conditions for peridotites at the same interval while upper their values sharply decrees.

In Prianabarie cold branch at the basement of the mantle columns reflect the developing of the metasomatites with clinopyroxene 70-60 kbar. The maim level of the magma fractionation close to 45-60 kbar seems to be continuous. Synchronous trend fO2 rise fro spinel seems to be the result of the high degree interaction also.

General tendency is that the multistage pipes have the more steps of the melt protokimberlite melt rising in the mantle. Long compositional and TP trends are more common for lithospheric mantle with continental signature and highly metasomatic mantle columns. Oxygen fugacity for ilmenite trends are partly within or close to the diamond stability field and at least some diamonds may grow up in megacrytstalline bodies or directly in the contacts

#### INTRODUCTION METHODS

Chromite (+Ol) and Ilmenite (+Ol) oxygen barometers [33] in monomineral versions [2,4] were used for the estimation of the oxidizing conditions of the lithospheric mantle beneath the kimberlite pipes in Yakutia The Ilmenite method [2] give the conditions for the megcrystalline veins served as the feeding systems for the protokimberlite melts for metasomatites and pyroxenites. Chromite [4] shows the conditions for peridotites. Usually the peridotites the oxygen conditions are lower the QMF in some unites; in most cases reducing is growing in depth. In opposite the ilmenite trends are close to the QMF slightly lower and the oxidation state is rising with the reducing pressure which also accords to the differentiation degree of the ilmenites. For the Kaapwaal craton the abrupt reduction of  $fO_2$  was detected at 150 km [19]. Melts in mantle are mostly more oxidized then peridotites due to the fast diffusion of the hydrogen from the magma. The values determined for the peridotites and ilmenites megacrystalls are compared with those found for the inclusions in the diamonds [28,20,29-31] dashed at the diagrams.

Usage of the monomineral versions give quite similar results as the bimineral versions (see fig.1) This means that the oxygen values are regulated by the Fe3 content of the spinels and ilmenites.



Fig. 1 Co- variations of the  $fO_2$  values determined by the Ol –Ilm oxygen barometer and values calculated with the monomineral versions.

# Variations of the oxygen conditions in mantle column beneath Siberian kimberlite pipes.

**Daldynsky region** (Udachnaya, Dolgozhdannaya, Aeromagnitnaya show relatively short P(kbar) intervals (65- 50 kbar) while for Zarnitza, Leningradskaya, Osennyaya and Ukrainskaya the main ilmenite fractionation interval is within 65-40 kbar. The Cr content for later pipes is rising together with the decreasing of the pressure and means the step by step contamination when melt intrude the next upper level. For the first three pipes the Cr is rising only at the end of the trend during the final melt solidification. Layered structure with the high variations of  $fO_2$  is determined for every mantle columns in Daldyn region.

On the diagrams for the P(kbar)  $-fO_2$  (Fig.2) for the mantle columns of the Daldynsky region the wide  $fO_2$  arrays determined with chromite with nearly equal P values were determined at 35-45 kbar and corresponds to the pyroxenite layer for Zarnitza, Udachnaya, Aeromagnitnaya and the other high variation interval  $fO_2$  corresponds to the diamond stability field. Practically in all pipes the relatively high number of the Cr-spinel TP values are plotted in the diamond stability in 50- 60 kbar interval but lower 60 kbars this became relatively rare and only diamond inclusions [22, 34-35] estimates are represented here. For the Udachnaya pipe each the 6 horizons detected with the spinel thermobarometry give there own range of the  $fO_2$  values. The deepest is reduced ~  $-4 -\Delta \log fO_2$ .

#### Table 1.

Element	TiO <sub>2</sub>	Al <sub>2</sub> O <sub>3</sub>	Cr <sub>2</sub> O <sub>3</sub>	FeO	MnO	MgO	NiO	$V_2O_5$	Total		
1	2	3	4	5	6	7	8	9	10		
				DALD	YN FIEL	D					
Aeromagnitnava pipe											
AerI21	54,51	1,01	0,98	30,24	0,25	14,28	0,11	-	101,28		
AerI23	53,11	0,60	3,93	27,47	0,43	15,24	0,11	-	100,49		
AerI26	52,70	0,45	4,81	24,15	0,53	15,52	0,13	-	97,80		
AerI97	48,15	0,49	0,90	39,75	0,30	9,05	0,01	-	98,56		
AerI99	47,74	0,45	0,84	41,67	0,32	8,20	0,06	0,17	99,37		
AerI100	47,30	0,67	7,89	34,89	0,21	10,36	0,18	-	100,71		
OSENNYAY PIPE											
OSII02	42,87	0,31	1,38	47,10	0,23	6,87	0,07	0,36	99,21		
OSI103	44,29	0,53	1,36	39,81	0,25	12,22	0,13	0,48	99,29		
OSI104	44,69	0,49	2,25	41,95	0,23	10,15	0,12	0,35	100,30		
OSI109	46,73	0,45	1,93	41,74	0,29	9,88	0,12	0,13	101,27		
OSII10	46,78	0,44	1,21	39,07	0,89	10,58	0,08	0,18	99,31		
OSII14	47,08	0,50	2,09	38,90	0,30	9,77	0,12	0,39	99,18		
UKRAINSK	AYA PIPE	3	•	1	•	•	1				
UkrI05	46,28	0,51	1,25	42,17	0,23	8,78	0,03	0,23	99,47		
UkrI08	46,59	0,45	1,19	41,83	0,29	9,48	0,09	0,24	100,15		
UkrI09	46,60	0,52	1,28	41,11	0,24	10,23	0,09	0,23	100,31		
UkrI11	46,66	0,62	1,27	39,90	0,24	11,55	0,08	0,18	100,49		
UkrI13	46,76	0,49	1,22	40,73	0,29	10,24	0,12	0,27	100,11		
UkrI14	46,81	0,53	1,26	42,58	0,29	8,60	0,10	0,15	100,31		
UDACHNA	YA PIPE		1		1			1	1		
Udch05	41,79	0,21	0,22	51,63	0,14	5,55	0,01	0,35	99,90		
Udch06	47,01	0,72	0,74	42,05	0,17	9,40	0,06	0,15	100,30		
Udch08	46,86	0,59	1,16	42,06	0,19	8,78	0,08	0,25	99,96		
Udch14	47,45	0,51	0,84	41,86	0,24	9,23	0,07	0,18	100,37		
Udch17	48,18	0,52	0,75	40,79	0,32	9,71	0,08	0,07	100,42		
Udch18	46,39	0,55	2,21	41,94	0,22	8,71	0,10	0,07	100,17		
MALOBOTUOBINSKY REGION											
Mir pipe	50.24	0.56	0.77	2675	0.21	10.22	0.10		00.14		
MirI01	30,34 49.61	0,30	0,77	20.00	0,51	10,52	0,10	-	99,14		
MirI02	40,01	0,48	0,80	24.50	0,34	9,12	0,10	-	98,39		
MirIOO	53 23	0,30	0,33	34,39	0,32	11,42	0,09	-	98,04		
MirI109	50.75	0,37	1,12	32,04	0,23	0.03	0,10	-	99,83		
MirI13	32 30	0,49	3.68	56.64	0,29	3.00	0,09	0.12	99,40		
MirI1/	36.46	0,74	0.66	56.08	0.14	1 30	0,48	0,12	98,15		
MirI16	48.05	0,07	0,00	41 38	0.21	<del>4</del> ,57 8,69	0,33	0.08	99.24		
DACHNAY	A PIPE	0,02	0,10	71,50	0,21	0,07	0,11	0,00	<i>)),2</i> +		
DacI01	52.49	0.72	0.81	31 33	0.49	13.48	0.17	_	99.49		
DacI04	52,49	0.56	4 19	28 31	0.54	13,40	0.15	_	99.47		
DacI05	51 77	0.76	0.68	33.64	0.46	12,00	0.13	0.09	100.24		
DacI07	51.72	0.51	0.57	35.25	0.33	11.08	0.10	-	99.56		
DacI11	51.55	0.39	4.07	28.13	0.53	15.24	0.10	-	100.01		
DacI12	51,52	0,90	1,28	32,95	0,40	12,58	0,20	-	99,82		

Representative analyses of ilmenites from Yukutia kimberlites

Table 1. Continued

1	2	3	4	5	6	7	8	9	10		
INTERNAT	IONAL'NA	YA PIPE									
IntI01	53,62	0,48	0,68	32,14	0,34	12,35	0,13	0,20	99,94		
IntI04	50,08	0,81	0,24	37,74	0,29	10,33	0,20	0,07	99,77		
IntI06	44,81	0,63	0,61	45,10	0,23	7,07	0,09	0,20	98,74		
IntI07	49,61	0,75	0,09	39,01	0,30	9,75	0,09	0,13	99,73		
IntI10	50,08	0,81	0,24	37,74	0,29	10,33	0,20	0,07	99,77		
IntI15	50,21	0,69	3,71	33,80	0,33	10,51	0,22	0,20	99,67		
AMAKINS	KAYA PIPI	E									
AmI159	51,14	0,20	3,91	32,90	0,51	12,25	0,09	0,38	101,38		
AmIl62	49,15	0,67	0,15	40,99	0,21	9,81	0,08	0,55	101,61		
AmIl65	42,47	0,80	0,12	47,92	0,15	7,82	0,11	0,67	100,04		
AmIl67	34,79	0,83	3,82	55,21	0,10	4,48	0,11		99,34		
AmIl70	44,94	0,59	0,16	46,10	0,16	6,99	0,04	0,27	99,26		
ALAKITE FIELD											
Aykhal pipe											
5AykS	50,34	0,56	0,77	36,75	0,31	10,32	0,10	-	99,14		
14AykS	49,34	0,49	0,79	38,90	0,33	9,34	0,08	0,14	99,41		
15AykS	51,16	0,50	0,55	34,59	0,32	11,42	0,09	-	98,64		
24AykS	53,23	0,57	1,12	32,04	0,25	12,48	0,16	-	99,85		
25AykS	50,75	0,49	0,73	37,12	0,29	9,93	0,09	0,07	99,48		
27AykS	48,38	0,25	2,07	39,47	0,24	9,93	0,11	0,29	99,02		
Ozernaya pi	be			,	,	,		,	,		
OZeI120	52,49	0,72	0,81	31,33	0,49	13,48	0,17	-	99,49		
OZeI122	52,21	0,70	0,73	32,44	0,36	12,50	0,21	-	99,15		
OZeI123	52,09	0,56	4,19	28,31	0,54	13,63	0,15	-	99,47		
OZeI124	51,77	0,76	0,68	33,64	0,46	12,71	0,13	0,09	100,24		
OZeI129	51,72	0,51	0,57	35,25	0,33	11,08	0,10	-	99,56		
OZeI135	51,55	0,39	4,07	28,13	0,53	15,24	0,10	-	100,01		
OZeI136	51,52	0,90	1,28	32,95	0,40	12,58	0,20	-	99,82		
YUBILEYN	AYA PIPE	2									
YbI09	52,58	0,54	0,64	35,05	0,23	11,56	0,08	0,09	100,77		
YbI10	50,11	0,14	3,98	35,31	0,32	9,70	0,16	0,47	100,20		
YbI11	50,10	0,50	1,68	37,49	0,26	10,12	0,11	0,18	100,43		
YbI15	44,55	0,34	4,12	40,93	0,31	7,77	0,10	0,22	98,35		
YbI16	51,08	0,47	0,71	41,05	0,27	6,95	0,04	0,34	100,92		
KOMSOMC	DL'SKAYA	PIPE									
KmI314	51,23	0,48	0,63	35,65	0,20	11,57	0,14	0,03	99,93		
KmI316	49,99	0,39	1,62	36,74	0,27	10,70	0,08	0,07	99,87		
KmI318	48,59	0,33	1,67	39,06	0,27	9,31	0,07	0,08	99,38		
KmI323	50,72	0,49	2,55	34,25	0,23	12,00	0,14	0,12	100,50		
KmI325	51,00	0,51	2,01	34,57	0,26	11,22	0,15	0,03	99,75		
KmI327	51,50	0,18	4,23	30,74	0,34	12,66	0,22	0,03	99,89		
SYTYKANS	SKAYA PI	PE									
SytI111	50,73	0,54	0,34	38,46	0,22	10,07	0,07	0,11	100,55		
SytI113	51,55	0,89	1,02	33,93	0,28	12,42	0,15	0,34	100,58		
SytI096	46,58	0,25	2,71	41,84	0,27	8,10	0,09	0,19	100,03		
SytI005	46,53	0,25	2,64	41,91	0,31	7,99	0,07	0,11	99,82		
SytI006	50,87	0,67	1,68	34,56	0,21	11,64	0,17	0,17	99,97		
SytI007	48,67	0,47	2,18	38,71	0,27	9,71	0,13	0,11	100,25		

Table 1. Continued

1	2	3	4	5	6	7	8	9	10	
SytI028	46,56	0,04	3,45	40,12	0,30	9,06	0,12	0,12	99,78	
SytI029	43,88	0,16	3,37	44,37	0,26	7,23	0,07	0,32	99,66	
SytI030	51,58	0,55	0,46	36,71	0,24	10,72	0,11	0,14	100,50	
SytI031	50,05	0,63	0,46	38,56	0,25	9,87	0,10	0,13	100,04	
SytI081	46,02	0,09	4,23	40,54	0,36	8,32	0,12	0,24	99,91	
				UPPER M	IUNA FI	ELD				
								Dei	mos pipe	
DeI001	52,70	0,63	1,76	30,83	0,24	13,71	0,22	0,32	100,40	
DeI002	51,30	0,51	0,73	35,43	0,26	11,70	0,12	0,36	100,41	
DeI012	50,29	0,47	2,04	36,27	0,29	10,17	0,12	0,46	100,11	
DeI013	48,93	0,44	1,85	39,84	0,79	7,74	0,21	0,60	100,40	
DeI015	51,93	0,53	0,65	35,14	0,24	11,22	0,12	0,24	100,07	
DeI016	47,42	0,25	2,24	40,83	0,30	8,18	0,07	0,51	99,80	
DeI018	52,44	0,68	0,54	33,61	0,28	12,32	0,15	0,34	100,38	
Novinka pipe										
NovIl02	47,97	0,72	0,22	40,12	0,21	9,03	0,10	0,15	98,53	
NovIl06	47,56	0,68	0,13	42,63	0,22	8,95	0,04	0,06	100,27	
NovIl07	51,90	0,38	1,10	36,79	0,29	10,00	0,08	-	100,54	
NovIl08	46,08	0,63	0,13	44,96	0,22	7,81	0,04	0,35	100,21	
NovIl10	48,75	0,60	0,10	41,61	0,30	8,94	0,05	-	100,36	
NovII11	47,89	0,61	0,16	41,75	0,18	9,11	0,09	0,11	99,89	
Zapolyarnaya pipe										
ZpI104	45,41	0,43	0,85	45,63	0,21	6,94	0,08	-	99,55	
ZpIl19	54,21	0,23	3,03	27,82	0,28	14,48	0,25	0,18	100,49	
ZpI120	54,73	0,80	2,99	25,78	0,42	15,37	0,26	-	100,35	
ZpIl28	44,43	0,48	0,88	46,52	0,23	6,71	0,06	0,11	99,25	
ZpIl29	46,78	0,39	0,88	43,48	0,21	7,75	0,11	0,10	99,49	
ZpI130	48,82	0,35	1,60	37,61	0,39	10,64	0,09	-	99,41	
ZpII32	53,58	1,00	3,74	27,46	0,24	14,54	0,26	-	100,56	
Komsomol's	skaya mag	nitnaya p	pipe							
KmM3	47,85	0,45	0,81	43,02	0,19	7,02	0,02	-	99,33	
KmM4	48,46	0,45	0,79	44,33	0,23	6,93	0,04	-	101,19	
KmM5	47,96	0,47	0,78	42,83	0,18	7,08	0,05	-	99,30	
KmM6	54,96	0,70	4,06	26,70	0,24	13,76	0,08	-	100,42	
				NAKY	'N FIELI	)				
Nyurbinska	ya pipe	0.40	0.00	20.20	0.00	0.40	0.47	0.01	100 50	
NurbI199	49,51	0,48	0,89	39,28	0,28	9,63	0,45	0,06	100,59	
Nurbi186	49,91	0,49	0,91	38,04	0,26	11,08	0,47	0,11	101,27	
Nurbl240	49,96	0,52	0,87	38,42	0,25	10,67	0,58	0,07	101,34	
Nurbl209	50,02	0,59	0,82	38,37	0,28	10,73	0,38	0,03	101,22	
Nurbl214	50,03	0,50	0,97	38,28	0,32	10,27	0,41	0,06	100,83	
Nurbl229	50,13	0,47	1,12	37,81	0,25	11,19	0,36	0,12	101,44	
Nurbl241	50,60	0,64	1,00	35,86	0,27	12,58	0,40	0,08	101,42	
Nurb1191	51.04	0,53	0,79	37,09	0,32	11,14	0,53	0,10	101,28	
Nurbi200	51,94	0,58	0,52	35,20	0,29	12,37	0,42	0,10	101,41	
Botuobinska	aya pipe	0 5 1	0.95	27 75	0.25	11.00	0.42	0.02	101.04	
2B 4D	50,15	0,51	0,85	31,13	0,25	11,09	0,43	0,03	101,04	
4B	49,61	0,46	0,89	39,64	0,27	9,72	0,31	0,08	100,98	
18R	49,39	0,88	5,19	34,19	0,25	10,87	0,66	0,17	101,59	



Fig..2.  $\Delta \log f_{O_2}$ . -P (kbar) diagram for the oxygen conditions determined with the bi mineral (Ol-Sp) oxybarometer [33], monominerl correction from the pressures determined with spinel and ilmenite monomineral barometers for the [2-4]) for the concentrate from the kimberlite pipes of Prianabarie.

But the upper megacrystalline dunites [26] are oxidized then again reduced with the gradual lowering of reduction to 30kbar. For Zarnitsa the layer near 55 kbar is relatively oxidized with the range characteristic for the diamond inclusions. Upper in 30 kbar again reduced horizon. The same was found for the spinels of Aeromagnitnaya pipe. The increase of the  $fO_2$  the conditions determined with the ilmenites means the interaction with the protokimberlite magma crystallizing the ilmenites. Continuous rising of the oxygen fugacity values for the ilmenites means the relatively closed system. While the break of the trend with the reduction in the upper level suggest the vein. Metasomatism with the interaction with the peridotites.

Alakite. At the diagrams  $\Delta \log f_{O_2}$ .-P kbar for the Alakite region the long continuous trends with two branches for all 5 studied pipes probably means the two stages of the malt percolation in the lithospheric mantle. For Aykhal it is more continuous then for others where the step by step rising conditions for spinels

probably means the interaction of the evolved and oxidizing melts with the peridotitic mantle.



# Fig..3. $\Delta \log f_{O_2}$ . -P (kbar) diagram for the oxygen conditions determined with the bi mineral (Ol-Sp) oxybarometer [33], monominerl correction from the pressures determined with spinel and ilmenite monomineral barometers for the [2-4]) for the concentrate from the kimberlite pipes of Alakite field.

The ilmenite trends everywhere corresponds to the chromite – based in pressures. For the Yubileynaya – at the basement to the top the traces of the interaction of the proto kimberlite melts produces several arrays with the negative inclination of the sub trends at the lower part due to interaction with peridotites of relatively small portion of the protokimberlite magma and to normal and then rapid fractionation at the upper twp level and then going to the metasomatic associations. Ozernaya represent at least 6 intervals here the lower is more oxidized then the upper what probably means the two stages of the formation megacrystalline ilmenite bearing veins. Komsomolskaya show quite continuous ilmenite trend of the fractionation to 50 kbar and then interaction at the upper levels.

*Malo Botuobinsky region.* The oxidizing conditions in the mantle columns for the chromites beneath the Malo Botuobinsky region are forming the wide inclined trends that are characteristic for the metasomatites. They partly coincide with the fields of the diamond stability. In mantle beneath the Internationalnaya pipe the wide  $fO_2$ . Dispersion near level ~40 kbars corresponds to the pyroxenite lens. Extended ilmenite trends for the Mir and Dachnaya pipes show relatively high inclination of the ilmenite PT path due to the fractionation in the closed system. For Mir at the basement to 55 kbars the sub vertical trend reflect the interaction with the peridotites. In mantle beneath the other pipes they reflect the stop of rising near 50 kbars and then probably developing of the metasomatic vein system upper.

Malo-Botuobinsky field



Fig..4.  $\Delta \log f_{O_2}$ . -P (kbar) diagram for the oxygen conditions determined with the bi mineral (Ol-Sp) oxybarometer [33], monominerl correction from the pressures determined with spinel and ilmenite monomineral barometers for the [2-4]) for the concentrate from the kimberlite pipes of Malo-Bouobinsky field.

#### Table 2.

Representative and yses of enformed from 1 arada rindernes
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Element	TiO <sub>2</sub>	Al <sub>2</sub> O <sub>3</sub>	$Cr_2O_3$	FeO	MnO	MgO	NiO	$V_2O_5$	Total
				DALDY	'N FIELI	)			
Aeromagnitn	aya pipe	-				-			-
AerC07	0,05	17,43	51,91	16,66	0,31	12,96	0,07	0,46	99,84
AerC09	0,22	8,85	60,92	18,16	0,28	11,40	0,06	0,38	100,28
AerC10	0,38	12,54	54,08	18,76	0,32	13,38	0,16	0,31	99,93
AerC11	0,23	5,44	62,71	19,01	0,31	12,05	0,11	0,20	100,05
AerC18	2,84	5,74	54,16	23,01	0,35	12,36	0,20	0,26	98,92
AerC20	1,16	4,05	57,68	26,17	0,39	9,85	0,13	0,33	99,76
Osennyay pi	<u>pe</u>		•			1			
OSCr019	0,56	5,24	58,98	23,03	0,19	11,94	0,10	0,17	100,24
OSCr022	0,74	4,82	59,39	21,49	0,19	12,64	0,20	0,20	99,67
OSCr024	1,81	6,52	55,14	24,10	0,30	12,35	0,15	0,24	100,63
OSCr028	3,03	6,96	49,25	28,33	0,28	10,92	0,24	0,31	99,34
OSCr031	3,75	12,94	41,03	27,29	0,12	13,28	0,33	0,29	99,02
OSCr032	4,71	5,02	36,28	43,08	0,23	9,96	0,25	0,39	99,94
Leningradska	aya pipe		1	1		1	1	1	
LEnS2	0,15	3,47	63,14	21,96	0,23	10,77	0,10	0,19	100,00
lenS02	0,03	48,51	20,77	11,93	0,11	19,16	0,20	0,11	100,81
lenS03	0,06	28,11	38,68	14,82	0,08	16,51	0,19	0,24	98,69
lenS04	0,07	23,63	46,42	15,44	0,16	13,92	0,08	0,25	99,96
lenS07	0,14	5,05	63,30	18,91	0,20	11,46	0,09	0,39	99,53
lenS09	0,16	4,99	64,51	18,42	0,27	10,84	0,10	0,42	99,70
Zarnitsa pipe				<u> </u>			<u> </u>		
Z55	0,38	9,15	59,07	17,94	0,32	11,46	0,08	0,39	98,79
Z67	3,06	2,35	48,36	36,51	0,48	7,67	0,15	0,38	98,95
Z18	-	17,60	48,19	21,59	0,36	10,89	0,05	0,29	98,97
Z51	4,98	1,78	38,57	44,94	0,44	6,92	0,20	0,59	98,42
Z19	0,02	39,09	29,25	14,71	0,17	16,57	0,15	0,12	100,07
Z44	5,58	1,69	34,71	48,88	0,45	6,37	0,28	0,55	98,51
			MAL	O-BOLU	OBINSKY	Y FIELS			
Mir pipe	0.00	47.10	00.41	10.04	0.25	01.04	0.00	0.10	00.40
MirC1052S	0,03	4/,18	20,41	10,04	0,35	21,06	0,29	0,12	99,48
MirC1053S	0,02	47,49	20,59	10,00	0,38	20,98	0,26	0,14	99,86
MirC1054S	0,65	26,58	35,04	20,18	0,55	16,31	0,23	0,31	99,85
MirC10555	2,47	0,07	43,89	35,72	0,99	8,84	0,21	0,22	99,01
MirC1057S	0,26	14,//	55,87	18,51	0,81	11,5/	0,07	0,44	100,30
MirC1058S	0,10	10,35	55,/1	22,23	0,96	10,41	0,07	0,31	100,14
MirC10595	1,29	8,54	30,78	21,40	0,94	10,01	0,15	0,30	100,07
Dachnaya pi	<i>2 60</i>	051	17 16	27.02	0.16	1151	0.15	0.20	09.91
DacS2	2,09	8,34 20.72	47,40	27,92	0,10	11,51	0,13	0,39	98,81
Dac55	2.09	7.91	47,34	10,00	0,12	14,92	0,14	0,34	99,89
Dacso	2,00	7.05	54.00	21,31	0,24	10,20	0,12	0,55	70,0/
DacSo	0,01	12.80	54,00	18.24	0,21	10,42	0,18	0,36	90,77
DacS12	0,20	6.02	61.28	10,24	0,13	11,40	0,00	0,35	00 20
International	U,11	0,02	01,28	19,23	0,17	11,91	0,10	0,39	99,20
Introlational	1 <b>7</b> 4	5 66	56.04	24.07	0.30	10.07	0.15	0.40	00.21
IntSO2	1,74	31.02	36.26	16.00	0,50	16.59	0,15	0.24	100.60
IntS03	0,14	27 54	12 10	15.09	0,10	14.04	0,19	0.24	100,09
111604	0,27	27,50	+∠,1U	10,00	0,24	14,24	0,10	0,24	100,39

Table 2. Continued

1	2	3	4	5	6	7	8	9	10		
IntS05	0.02	846	57 17	22.53	0.26	10.34	0.03	0.19	99.00		
IntS10	0.12	25.62	41 43	17.07	0.12	15.65	0.16	0.29	100.45		
IntS13	0.28	5 68	59 38	22.46	0.12	11 18	0.12	0.31	99 59		
IntS19	1.93	6.82	52.61	26.98	0.27	11,10	0.12	0.25	100.20		
IntS1	0.67	8.51	54.87	25,05	0.29	10.22	0.13	0.23	99.97		
Amakinskava	a nine	0,51	54,07	23,03	0,27	10,22	0,15	0,24	,,,,		
AmS159	0.27	7 48	59 34	20.92	0.29	10.48	0.10	0.48	99.36		
AmS168	2 00	3 15	53 58	30.32	0.44	9.26	0.18	0.40	99.33		
AmS170	0.55	11.90	55,50	18.86	0.27	12 51	0.12	0.42	100.19		
AmS172	2.88	4 88	46 69	34.89	0.37	9.20	0.26	0.41	99 58		
AmS172	0.12	22 10	46,05	17 34	0.16	12 77	0.07	0.36	99.87		
AmS169	0,12	10.30	54 63	21 30	0.18	11.55	0.11	0.48	99.23		
	0,00	10,50	0 1,00	ALAKI	TE FIELI	)	0,11	0,10	,,25		
Δykhal nine											
A-1-0C06	0.12	5.01	(2)((	17.24	0.26	11.71	0.04				
AykSC06	0,12	5,84	63,00	17,24	0,26	11,/1	0,04	0,27	99,14		
AykSC08	0,03	0,00	61,90 54.06	18,87	0,20	11,05	0,08	0,55	99,30		
AykSC20	0,31	11,02	54,00	21,74	0,35	11,01	0,09	0,28	99,40		
AykSC25	0,03	18,17	49,92	17,55	0,25	12,34	0,02	0,294	99,04		
AykSC20	0,23	10,05	48,20	20,52	0,25	0.52	0,10	0,27	100,24		
AyKSC27	4,81	4,50	43,08	32,78	0,51	9,32	0,18	0,28	99,04		
AykSC28	0,39	24,09	41,87	15,05	0,18	10,38	0,10	0,21	99,07		
Ozernaya pip	be	5 20	64.24	10.06	0.22	11.72	0.00	0.27	100.26		
OZeCr09	0,25	5,50	64,54	18,00	0,23	11,/5	0,09	0,37	100,30		
OZeCr12	0,81	4,03	01,42	20,45	0,24	11,70	0,12	0,20	99,03		
OZeCr3	1.93	10,75	55.20	21.90	0,23	12,54 8 50	0,08	0,20	99,91 100.60		
OZeCr39	1,01	2,24 4 30	52.27	26.87	0,40	0,30 10.77	0,19	0,38	00.10		
OZeCr31	4,21	4,30	50.06	20,67	0,39	10,77	0,11	0,20	99,19 100.21		
OZeC134	3,23	4,70	J0,90 45 58	30,34	0,39	9,00	0,19	0,34	00.04		
Vubilornovo	4,1/	1,05	45,56	39,31	0,42	/,41	0,17	0,55	99,04		
Tubleyllaya VubS02	0.02	18 87	11.08	17.21	0.10	20.82	0.47	0.18	100 55		
Yub\$03	0,92	40,07	27.57	10.28	0,10	18.00	0,47	0,10	00.01		
YubS04	0.17	35.02	31.49	14.62	0,00	17.95	0,15	0.22	00.81		
Yub\$05	0.37	15.02	38.81	28 70	0,13	15.54	0.320	0,22	100.19		
YubS08	3.12	12 54	45 69	23,70	0.12	13,54	0.27	0.33	99.26		
Yub\$09	0.15	11/13	55.05	19.67	0,17	11.67	0.05	0,33	99.45		
Komsomol'sk	ava nine	11,75	55,75	17,07	0,20	11,07	0,05	0,55	<i>уу</i> ,ч <i>3</i>		
KmsS1059	0.04	5.28	64 19	16 75	0.16	12 67	0.10	0.29	99 49		
KmsSl048	0.44	10.75	61 11	16,75	0.18	11.08	0.13	0.29	100.18		
KmsSl054	0.27	12.96	56.86	16.84	0.17	11,00	0.07	0.27	99 38		
KmsSl021	1.28	8 16	54 29	26.35	0.78	9 51	0.14	0.35	100.86		
KmsSl046	2.64	12.23	46.61	23,99	0.17	14 01	0.20	0.25	100,00		
KmsSl052	0.34	29.91	39.24	14 13	0.12	15 35	0.13	0.35	99 55		
Sytykanskava	a pipe	,,_	,	,		-2,55		5,00	,00		
Svt3	0.26	8.01	59.21	21.08	0.28	10.60			99.44		
Svt4	0.61	8.21	58.28	21.58	0.28	10.83			99.78		
Svt5	0.47	9.72	55.03	23.74	0.32	10.14			99.43		
Svt7	1.83	3.69	51.92	32.28	0.44	8.91			99.07		
Syt8	2.56	14.32	43.88	26.41	0.27	11.32			98.75		
Syt9	0.22	8.59	58.51	21.23	0.32	10.72			99.58		
· ·	/	,	7 -	, -	,-	7 .	1	1	/		

								Table 2.	Continued	
1	2	3	4	5	6	7	8	9	10	
			UP	PER MUN	IA FIELD	)				
								Dei	mos pipe	
DeiS22	0,03	4,97	65,56	17,15	0,19	11,66	0,06	0,32	99,93	
DeS025	2,69	1,55	56,61	29,58	0,30	8,14	0,20	0,32	99,40	
DeS026	0,02	48,03	20,50	12,22	0,12	18,70	0,18	0,10	99,88	
DeS032	4,91	8,00	47,69	26,42	0,32	11,73	0,22	0,29	99,59	
DeS034	1,35	3,09	57,31	28,66	0,34	8,42	0,118	0,33	99,60	
DeS037	0,37	26,82	39,47	19,54	0,21	13,12	0,15	0,16	99,04	
DeS038	0,403	7,28	56,14	26,72	0,29	8,70	0,11	0,23	99,67	
Novinka pipe										
OZeCr06	0,08	19,62	41,86	21,71	0,531	14,01	0,23	0,19	98,23	
OZeCr31	0,77	9,58	50,72	28,11	0,723	8,95	0,1	0,32	99,27	
OZeCr08	0,04	20,47	46,02	16,21	0,534	15,01	0,117	0,32	98,72	
OZeCr34	0,11	32,76	35,73	14,66	0,489	15,19	0,076	0,29	99,30	
OZeCr21	0,70	20,24	40,66	21,1	0,541	15,79	0,219	0,17	99,42	
OZeCr30	0,10	40,42	27,23	12,02	0,348	19,39	0,214	0,19	99,92	
Zapolyarnaya pipe										
ZpS11	0,04	5,02	63,72	17,13	0,85	12,32	0,06	0,26	99,40	
ZpS13	0,04	5,41	64,64	16,36	0,76	12,66	0,08	0,38	100,33	
ZpS14	0,04	4,29	62,01	21,20	0,73	10,68	0,11	0,24	99,31	
ZpS16	0,07	12,48	55,19	18,33	0,67	12,62	0,10	0,28	99,73	
ZpS18	0,08	4,45	62,55	17,91	0,69	13,33	0,10	0,32	99,42	
ZpS19	0,08	9,10	59,17	18,17	0,69	13,05	0,07	0,22	100,56	
ZpS21	0,09	12,85	55,24	17,19	0,66	13,40	0,08	0,25	99,76	
ZpS29	0,44	18,51	34,74	32,59	0,59	12,04	0,09	0,43	99,43	
Komsomol's	kaya magi	nitnaya p	ipe	1	1	1	1	1		
KmMS01	2,72	1,13	57,60	29,21	0,51	8,32			99,49	
KmMS05	0,02	5,37	63,23	16,95	0,28	12,41			98,26	
KmMS07	1,75	5,55	58,98	19,67	0,33	11,68			97,96	
KmMS11	0,05	5,82	65,25	16,40	0,30	12,66			100,49	
KmMS14	0,91	14,22	51,21	19,78	0,26	13,59			99,96	
KmMS15	3,36	4,94	59,33	18,58	0,31	12,98			99,50	
KmMS17	0,22	8,59	58,51	21,23	0,32	10,72			99,58	
NAKYN FIELD										
							Ny	urbinsk	aya pipe	
"145m19 "	0.32	5.34	63.70	19.10	0.24	10.24	0.09	0.39	99.43	
"160m22 "	0,13	17.27	52,84	17,34	0.20	11,84	0.03	0.38	100,01	
"145m18 "	0,30	12,62	51,80	23,63	0.20	10,92	0,11	0.38	99,97	
"130m09 "	2,87	8,29	51,92	23,54	0.21	12,14	0.13	0,56	99,66	
"130m10 "	0,48	22,72	45,56	17,53	0,20	13,36	0,09	0,26	99,67	
"145m03 "	0,17	47,17	21,30	12,52	0,10	18,21	0,29	0,22	99,69	
"175m06 "	1,38	8,75	58,66	17,46	0,26	12,04	0,10	0,33	99,94	
Botuobinska	ya pipe		· ·			· · · ·	I		· · ·	
5B	0,91	17,81	46,60	18,67	0,54	14,21	0,167	0,279	99,18	
7B	0,28	18,81	42,38	24,44	0,58	12,82	0,150	0,202	99,66	
11B	0,67	26,13	35,40	22,01	0,42	14,79	0,218	0,094	99,73	
12B	0,20	5,19	62,98	19,16	0,74	10,97	0,025	0,233	99,49	
13B	1,09	13,87	51,27	21,71	0,61	11,25	0,093	0,346	100,24	
15B	0,22	5,68	63,16	18,01	0,81	11,33	0,043	0,168	99,42	
22B	0,22	13,41	51,22	21,54	0,66	11,85	0,076	0,246	99,21	

Upper -Muna field .The Cr- spinels from the pipes of this field in general give several continuous trends  $\Delta \log fO_2$ -P with the increasing of the oxygen fugacity in the lower part of the mantle section. This means the percolation of the Upper-Muna field



Fig..5.  $\Delta \log f_{O_2}$ . -P (kbar) diagram for the oxygen conditions determined with the bi mineral (Ol-Sp) oxybarometer [33], monominerl correction from the pressures determined with spinel and ilmenite monomineral barometers for the [2-4]) for the concentrate from the kimberlite pipes of Upper Muna field.

reduced melts from the bottom which was not associated with those produced ilmenite megacrystalls. For pipes Deimos and Zapolyarnaya it starts from 60 kbar and join with the ilmenite trend at QMF buffer near the pyroxenite lens. The other trends started from 50 kbar probably means the presence of intermediate melting levels and percolation upward what corresponds to rather enriched composition of the mantle beneath Upper Muna field. Interesting feature that chromite rising trends are often finished with the ilmenite associations. In mantle beneath Malaya, Inter-Kosmos, Debyut there are not less then 5 intermediate trends with different  $fO_2$  starting values. For the upper part (less then 30 kbar) relatively reduced sub vertical trends probably mark the way of the mantle rising diapir. Ilmenite trends

here at the deep levels are presented only in Deimos Zapolyarnaya and Novinka pipes that are of the most practical use.

*Nakyn field.* Oxygen conditions for the mantle columns beneath the Nakyn field are determined more detail for Nyurbinskaya pipe. It shows the two branches at the deep level from 60 to 40 kbars. And more reduced conditions are determined at 50-40kbars. High variations are typical at 40 kbar. Ilmenite trend developed at the basement 65-50 kbar is of typical magmatic type and rather oxidized. Some deviations to reduced conditions refer to the interaction with the peridotite mantle.

#### Nakynsky field



Fig..6.  $\Delta \log f_{O_2}$ . -P (kbar) diagram for the oxygen conditions determined with the bi mineral (Ol-Sp) oxybarometer [33], monominerl correction from the pressures determined with spinel and ilmenite monomineral barometers for the [2-4]) for the concentrate from the kimberlite pipes of Nakyn field.

**Prianabarie.** Variations of the peridotite mantle in Prianabarie are represented mainly for the Vympel pipe. The common values for mantle peridotites near -4-2 log units are more complicated at the 50-55 kbar due to interaction with the ilmenite parental melts which also demonstrate the reduction trends. The interaction is extended up to pyroxenite lens.

Deep seated magmatic trends for the ilmenites are typical for the mantle columns beneath the Trudovaya, Hardah, and Anomaly 17.3. For the later ilmenite and spinel trends practically coincide what probably means the high degree of the interaction in the deep part of mantle columns beneath the Dyuken. This is

emphasized by the high ulvospinel content of chromites there. Sub vertical trends are separated at 3-5 parts probably due to the mantle layering. Ilmenites from Hardah pipe show several sub parallel trends probably due to the several pulses of the melt percolation what coincides with the zonation in the grained ilmenite aggregates. The break at the upper part of the mantle columns and formation of the separate trend with the more inclination probably means the direct percolations of the ilmenite parental melt in to relatively small veins systems.



Fig..7.  $\Delta \log f_{O_2}$ . -P (kbar) diagram for the oxygen conditions determined with the bi mineral (Ol-Sp) oxybarometer [33], monominerl correction from the pressures determined with spinel and ilmenite monomineral barometers for the [2-4]) for the concentrate from the kimberlite pipes of Prianabarie.

#### Discussion

# **Reconstruction of the vein-feeding deep seated magmatic systems based on the ilmenites.**

Ilmenite is one of the most wide spread mineral in the upper mantle conditions and is developed in the different Magmatic associations in mantle lithosphere. It is typical for the protokimberlite or carbonatite melts that were forming the feeding channels for the further eruptions. In this megacrystalline associations ilmenite are forming the intergrowths with the olivine(Ol), garnet (Gar), and contain the inclusions of rutiles (Rut), apatite (Ap), zircons (Zr) In the polycrystalline nodules the contacts wit the Ti- enriched peridotites which include the Cr- rich garnets (to 9 %). Monomineral ilmenite thermobarometry give the positions in the mantle sections of the magmatic bodies and metasomatites. The calculate fractionation models in general refer the precipitation of the ilmenite and olivine but the fluctuations of Cr, Ni, V, Al also suggest the contamination and participation of the garnets and sometimes pyroxenes in the crystallization processes. Simple schemes of the formations of the Magmatic systems in the mantle it is possible to make basing on the thermobarometry and deciphering of the trends of the major components.

The intervals of the pressure found for the ilmenites are in coincidence with the estimates based on the other minerals- Cr- spinels, garnets pyroxenes. Ilmenite trend that was analyzed from the concentrate from Zarnitsa pipe consist from the 7 intervals and is characterized by the fluctuations of the  $Cr_2O_3$  in each level and probable characterize the polybaric systems with the local stops at the boundaries between the layers. Of the primary subduction horizons. Ilmenites from he Osennyaya pipes are characterized by the irregular Cr concentration within the mantle section what probably refer to the high temperature interaction of the varying in scale magmatic systems in the peridotite wall rocks at the different levels. Ilmenite trend from Udachnaya is just close to the PT values of the convective branch (the same for Dolgozhdannaya) at the basement and became hotter in the 50-40 kbar interval. Ilmenites from Aeromagnitnaya probably trace the hot metasomatic system at the basement.

Lithospheric mantle columns beneath the Mir and Dachnaya [7] pipes traced by the ilmenite trends from bottom to the Moho with some intervals according to the layering are close. The TP gradients (see the paper in this volume) for the ilmenite are close to adiabatic or at least correspond to the advective geothems. In the bottom The TP values are close to the hot peridotitic branch and upper the ilmenite bearing rocks became more H-T<sup>o</sup> as well as the spinel peridotites. Upper the pyroxenite lens the different rapidly growth what probably means that this magma conduit serve for the transference of the large Magmatic masses or very rapid intrusion of the magma in the open fractures. Ilmenite Pt values for the Amakinskaya and Taezhnaya pipes are restricted in the pressure to 50 kbar and refer to the conductive TP gradients that are calculated for the garnets and clinopyroxens.

In Alakite mantle domain [8] the PT values of ilmenite trends reflect the formation the mantle magma feeding systems and various metasomatic rocks in the different levels in the mantle. For the Aykhal pipe this is the continuous system with the gradual evolution judging by the smooth TP pass. There were at least four separate levels of the protokimberlite systems beneath the Aykhal pipe. The pipe Yubileynaya show gradual rising of the magmatic system through the several separate levels to pyroxenite lens. Upper it trace sub vertical advective geotherm ~950°C. Close located Ozernaya pipe also reveal several levels of the metasomatites which give the range of the T°C values it each horizons to pyroxenite branch. In the upper part of mantle column the TP gradient became

close to SEA geotherm forming together with the Cpx and Gar's TP estimates the adiabatic branch. The degree of the interaction judging on the Cr – content rapidly raises upward the mantle section. The amount of the melt in this mantle column was lower then beneath the Aykhal pipes where the Cr is nearly constant in ilmenites and raises only upper then 40 kbars. In the mantle columns beneath the Sytykanskaya and Komsomolskaya pipes the magmatic sources were developing only from the bottom to the interval 40kbars mainly with the high interaction within the mantle columns and formation of the metasomatites. The tracers of the veins beneath the Sytakanskaya are rising to the spinel facie.

Beneath the Upper Muna field [6] using TP values for ilmenites it possible to determine several stages of the creation of protokimberlite system and not less then five levels of the magma concentration levels within the main channel of the rising system. Ilmenite TP estimates traces 3 different geotherms and probably reflects TP passes of the different in the sizes magmatic systems or different stages of the evolution of Magmatic sources. Starting from the 55 kbars the interaction with peridotites became visible due to the Cr- rising. Ilmenite TP estimates for the mantle columns beneath the smaller pipes Poiskovaya, Malaya and Inter- Kosmos are not abundant and localized mainly upper 50 kbar are localized reflecting for the Inter-Kosmos the polybaric system rising to the Gar-Sp facie. For Malaya the main Magmatic source probably is locating near 50 kbar.

In Anabar region ilmenite TP values reflect the most deep seated associations in most cases located just on the geotherms and TP values obtained with the silicates. The TP conditions for ilmenites in most cases are the only points represented on the TP diagrams in the lower part of the sections due to the much depleted composition of the Anabar mantle. They represent the metasomatites that rarely contain Cr- bearing garnets and clinopyroxenes evolving in the vein system up to 40 kbars. Starting from this level the Mn bearing ilmenites that were formed in the eclogites slices where the presence of Mn - bearing sediments is probable. Ilmenites from the Dyuken field in general give more restricted interval of pressure 55-40 kbars which is closed to the characteristics of some small pipes of the Upper Muna field

#### CONCLUSIONS.

- 1. Monomineral thermo-and oxybarometry for pycroilmenites and chromites allow to estimate the variation of the redox conditions within the mantle columns beneath pipes showing mainly the process of the melt percolation
- 2. Basement of the lithosphere show high scale variations of the oxidation state which are traced mainly by diamond inclusions of chromites. The upper levels are marked by the rising of  $fO_2$  to the 40kbar (pyroxenite lens) for peridotites.

3. Ilmenite shows the sub vertical of the rising feeding system at the preeruption stage with the more reduced conditions for the contaminated portions and for metasomatites.

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# Intensity of asthenospheric influence as a crucial factor for the diversity of kimberlites

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Detailed and comprehensive examination of all the late Devonian volcanic rock occurrences known in the Zimni Bereg kimberlite district (more than 60 igneous rock bodies) revealed a regular, symmetric distribution of variations in kimberlite composition within this area. The wide variety of local kimberlites may be due to polygenetic and polyfacial structure of the mantle, which, in turn, may be related to asthenospheric matter intrusion (diapiric or local) into the old, "homogeneous" lithospheric mantle in the Middle to late Devonian. This intrusion might have resulted in the formation of a peculiar mantle substrate in the zone of kimberlite magma generation, this substrate being complex both in structure and in composition due to participation of newly formed asthenospheric rocks and to varying intensity of metasomatic effect exerted by the asthenospheric intrusion on old lithospheric rocks. The degree of asthenospheric influence on the mantle substrate, which manifests itself in kimberlite mineralogy (presence of picroilmenite), geochemistry (Ti, Nb and Ta contents) and Nd-Sr isotope parameters (close to BSE) would be one of the definitive factors when classifying kimberlites by composition. The "Ta-Sc" diagram, accounting for the type of mantle source of the rocks and degree of kimberlite melt differentiation at once, is most efficient and versatile for geochemical typification of different kimberlitic rocks.

#### INTRODUCTION

The kimberlitic rocks occurring in different regions worldwide vary in composition very widely, and their variety appears to widen with routine discoveries of new kimberlite occurrences and as the known occurrences become better studied. The great diversity of kimberlitic rocks does not fit in with the commonly accepted strict classification scheme which includes just two kimberlite groups (Group 1 and Group 2, [19]) and a single lamproite group. There is a need to develop a regular classification system accounting for the totality of kimberlite varieties, since a good rock classification scheme is a useful research tool for genetic derivations. To do this, it is necessary to elucidate the essential factors and processes governing the changes and variations in kimberlite composition.

Much useful information, which would help tackle this problem, could be gained from the analysis comprehensively explored and studied Zimni Bereg district. Varying quite widely in composition, kimberlites of this province have similar modes of occurrence, their bodies being therewith not only closely located but also regularly distributed, all of them having essentially the same (Late Devonian) age. In addition, almost all the local igneous rock bodies contain highpressure indicator minerals of the coesite subfacies [13] and (even though single) diamonds (according to sampling data available from mining companies), which may suggest a common magma chamber initiation depth for local kimberlites. This similarity of many characteristics at once shown by the Zimni Bereg volcanics, along with the wide variation in their composition, suggests that the formation of local kimberlitic rocks may be related to a single common magmatic event, which would be "convenient" for detailed analysis of this event and evolution sequels. We infer that the wide variety of Zimni Bereg kimberlites and their rocks is due to an interaction of the asthenospheric and lithospheric mantle sources, or plumemantle interaction [4,5]. The result of plume effect on mantle rocks and the ensuing development of local igneous rock varieties with differing composition are the subject matter of this study.

# Subject matter and methods of investigation

The Zimni Bereg kimberlites and kimberlite-allied rocks have been extensively and thoroughly investigated (including isotope studies) by many researchers [1,4,5,8,9,12 and many others].

The main objective of this study was to analyze the regularities in compositional variations of Zimni Bereg kimberlites and their allied volcanic rocks, which have been revealed based on generalization and interpretation of data on geology, petrography, geochemistry and high-pressure mineralogy for all (more than 60) known igneous rock bodies in this province (including about 4000 mantle xenoliths samples), along with Nd-Sr isotope data on 30 igneous rock bodies representing all volcanic rock varieties of the Zimni Bereg district.

The obvious spatiotemporal unity of the Zimni Bereg volcanics and their general similarity in geological structure suggest some consanguinity for all Early Hercynian volcanic rocks in this province, which, in turn, enables one to treat this volcanic area as a "spatiotemporal development for the mantle chamber evolution" [Vaganov, 1987]. Analysis of this evolution was performed through isoline mapping of different compositional characteristics for Zimni Bereg volcanics. Geochemical and isotope studies were conducted predominantly with autolithic inclusion. In isoline mapping, arithmetic average values of the parameters being plotted were used for each particular igneous rock body.

When plotting the model plans and sections for the upper mantle of the study area to illustrate the rock spatial arrangement of different mantle rocks, we assumed that each volcanic pipe is a "natural borehole" of a sort, such that the high-pressure material it contains must, in one way or another, have bearing on the sequence of "sampled" mantle rocks underlying this pipe [2]. The "arrangement" of the different mantle rocks in depth was performed having regard to the following inferred direct correlation: "forming pressure of mantle xenolith mineral suites = lithostatic load = rock occurrence depth" [2].

## MINERALOGICAL AND GEOCHEMICAL PECULIARITIES of Zimni Bereg volcanic rocks

Comprehensive examination of Zimni Bereg volcanic rock occurrences revealed that among local kimberlites rock are representatives of two distinct groups (series), drastically different from each other in most of the compositional characteristics, including geochemistry, isotope chemistry and high-pressure mineralogy: Al-series and Fe-Ti-series [8,9] (Figure 1). Within each of these series, kimberlitic rocks also differ from each other, however, only "quantitatively". The "qualitative" dissimilarity in rock composition is related to peculiarities of the initial mantle substrate, the "quantitative" distinctions are due to differentiation nuances of the products of initial mantle substrate transformation (extent of disintegration and partial melting).



Fig. 1. Distribution of volcanic rock on the Zimny Bereg kimberlite area. I-II – line of the section for the figure 7.

The Al-series kimberlitic rocks formed from an extremely geochemically "depleted", homogeneous "dunite" substrate, which is reflected in the very restricted set of mantle rock relicts occurring in kimberlites: chrome spinel dunite (sharply predominating), garnet dunite (rare), peridotite (very rare), pyroxenite and eclogite (few if any).

The Fe-Ti-series kimberlites formed from a heterogeneous "eclogiteperidotite" mantle substrate of "fertile" type, which manifests itself in the very rich spectrum of mantle rock relicts in these kimberlites: garnet peridotite and pyroxenite; ilmenite, garnet-ilmenite and garnet-rutile peridotite; various eclogite (Groups A, B and C) and grospydite rocks; clinopyroxene-phlogopite metasomatic rocks; lastly, varuous megacrists (garnet, olivine, orthopyroxene, clinopyroxene , phlogopite, ilmenite, rutile, omphacite-jadeite). In cases where the rocks contain no mantle xenoliths, mantle material products occur in Al-series kimberlites as single grains of chrome spinel and\or, more rarely, chrome-diopside and pyrope, while in Fe-Ti-series rocks they occur as single grains of picroilmenite, pyrope and\or, more rarely, chrome spinel, but most commonly as Ti-rich chrom spinel from the rock matrix.

The two kimberlitic series differ drastically in geochemistry. The Fe-Ti-series rocks are characterized by significantly higher than average concentration of incompatible elements, primarily and particularly Ti, Ta and Nb, and also Th and Hf. This agrees perfectly with the abundance of picroilmenite, the main high-pressure host mineral for Ti, Ta and Nb, in the parental mantle substrate of Fe-Ti-series kimberlitic rocks.

Within each of the two series, the rocks form regular sequences (rows), in order of decreasing ultramaficity, from diamondiferous kimberlites to melilitites and picrites, which shows as a decrease in high-pressure material (olivine macrocrystals and mantle nodules) content from 50 vol. % to trace amounts, along with an increase in relative abundance of kimberlite melt crystallization products (phenocrysts of olivine, melilite, nepheline, phlogopite, Ti-rich chrome spinel and glass). Along with this, concentration of compatible elements also show regular trends of decrease in Ni, Cr, Co contents and magnesian index and increase in Al, Ga, Y, Na, Fe, V and Sc contents. Diamond potential of the rocks in each of the two series depends on how close are their mineral modes and chemical composition to those of the initial diamondiferous ultramafics of the melting mantle substrate.

The Al-series rocks form a sequence (row) from diamondiferous kimberlites of the Zolotitsky pipe cluster (the Lomonosov deposit) to kimmelilitites (the term taken from [Milashev, 1974]) and divine melilitites of Verhotina, Suksoma, Chidviya and Izhma. The Fe-Ti-series rocks form a row from diamondiferous kimberlites of the V. Grib deposit and poorly diamondiferous kimberlites of Pachuga to kimpicrites [Milashev, 1974] and melilite picrites of Shocha, Kepina, Soyana, Pachuga and Megra. The carbonate-phlogopine picrites comprising the famous sill at the Mela River occupy a peculiar, "intermediate" position in the Zimni Bereg magmatic system, being just somewhat closer to Al-series kimberlites.

By a set of compositional characteristics, the Fe-Ti-series kimberlitic rocks of Zimni Bereg are similar to the petrological type reckoned as Group 1 South African kimberlites, for which an asthenospheric mantle source is inferred. The asthenospheric effect manifests itself in the mineralogical characteristics of the Fe-Ti-series rocks (presence of picroilmenite, "non-depleted" peridotite and megacrysts), as well as in their geochemistry (high Fe, Ti, Nb and Ta contents) and Nd-Sr isotope characteristics (close to BSE) [Lasko, 1979]. As regards Al-series rocks, they are in some respect similar to the petrological type known as Group 2 South African kimberlites, for which an old lithospheric mantle source is inferred. Along with this, the Al-series rocks appear to be a good candidate for the role of a "petrological type" for kimberlites of lithospheric origin (no matter what their name will be). Those rocks are stronger geochemically depleted and dissimilar in composition relative to Group 1 asthenospheric kimberlites (and to Fe-Ti-series Zimni Bereg kimberlitic rocks) than the South African Group 2 kimberlites, obviously being related to an extremely depleted mantle substrate. Alternatively, the role of petrotype for the geochemically "depleted" lithospheric mantle could be played by the homogeneous "dunite" substrate of the Zimni Bereg district [11], and its geochemical characteristics, when calculated as average composition parameters for the most abundant mantle rock xenoliths from kimberlites of the Zolotitsky pipe cluster, may be of great petrological importance, along with the average "primitive mantle" composition.

The "Ta-Sc" diagram is the most efficient and convenient discrimination diagram being used to group (typify) kimberlites by geochemistry and identify particular kimberlitic rock varieties within the previously identified series (Figure 2). This diagram enables one to take into account the geochemical type of mantle source and the extent of differentiation (and ultramaficity) of kimberlitic melts at once, since Ta content of the kimberlitic rock reflects the extent of asthenospheric influence on the mantle source, and its Sc content has bearing on the ultramaficity of this rock.

In kimberlite geochemistry studies, only autoliths and bulk of rock with xenogenic material content not higher than 5% are usually dealt with. The bulk rock composition is indicative of the abundance of depth-derived mantle material relicts (primarily, olivine macro crystals) in the examined rock, and the composition of autoliths has bearing on the character and extent of kimberlitic melt differentiation. It is precisely the pair "autolith-rock bulk free of xeniliths" that yields maximum geochemical information on a kimberlitic rock. As regards isotope chemistry studies, these are only worthwhile with autoliths or effusive-appearing rocks.

In the "Ta-Sc" diagram, the Fe-Ti-series rocks are prominent due to high Ta content of their autoliths (10-40 ppm), with no significant difference found in this respect between kimberlites, kimpicrites and picrites. The lowest Ta contents (8.6

ppm for autoliths and 3.3 ppm for kimberlitic tuffisites) are found in the diamondiferous V.Grib pipe, which is characterized by the highest ultramaficity among the Fe-Ti-series kimberlitic rocks.



Fig. 2. Diagram of Ta-Sc for Zimnii Bereg volcanic rocks.

Lines show of Al series and Fe-Ti series volcanic rock composition trends. Rhomb shows the «primitive mantle» compositions [15]; circle shows the «depleted mantle» compositions [11].

The Al-series rocks are characterized by a lower than average Ta content (below 5 ppm), with the following trend: in the row "kimberlite-kimmelilitite-pyroxene-free olivine melilitite", the Ta content consistently decreases from 3-5 ppm to 2-3 ppm and 0.5-1 ppm, respectively. Along with this, Sc content of the rocks increases from 8 to 21 ppm.

Compositional trends of all kimberlitic rock types, and even of the "geochemicaly enriched" Fe-Ti-series rocks, originate precisely from the field of the calculated average "depleted" mantle composition rather than from the field of average "primitive" (or "fertile") mantle composition. Quite close to each other are kimberlitic tuffisites from the Lomonosov and V.Grib deposits with a high proportion of crushed mantle material; these rocks are most similar in composition to mantle peridotite.

When analyzing the Nd-Sr isotope characteristics of the kimberlitic rocks (Figure 3), a situation similar to that described above is observed in the  $\epsilon$ Sr- $\epsilon$ Nd diagram: different Fe-Ti series rocks, regardless of their ultramaficity (be it

kimberlite, or kimpicrite, or picrite), are characterized by quite similar parameters, close to BSE (which is characteristic of asthenospheric mantle, as was revealed previously [4,5]). Alternatively, Al-series rocks form a regular row (according to the trend of change in Nd-Sr isotopic characteristics) within the compositional field of old, enriched lithospheric mantle: from diamondiferous kimberlites of the Lomonosov deposit ( $\epsilon$ Nd = -2.5 / -4.9;  $\epsilon$ Sr = -4.2 / -18.8 ) to melilitites of Izhma and Suksoma ( $\epsilon$ Nd = -8.6 / -10.2;  $\epsilon$ Sr = -29.6 / -60.5).



Fig. 3. Sr and Nd isotopic composition of the Zimni Bereg kimberlitic rocks (for a kimberlite age 370 Ma).

To summarize, based on the most indicative isotopic and geochemical characteristics, the Fe-Ti-series kimberlitic rocks form a quite homogeneous group (regardless of their degree of ultramaficity), whereas the Al-series rocks are characterized by regular and intercorrelated variations in composition, including the Ta-Sc relationships and Nd-Sr isotopic parameters.

Analysis of the spatial distribution of kimberlitic rocks with dissimilar compositional characteristics in the Zimni Bereg area allows us to make certain inferences as to one of the factors probably responsible for the actual diversity of local kimberlites and their allied rocks.

## ZIMNI BEREG KIMBERLITIC ROCKS: RELATIONSHIP BETWEEN COMPOSITIONAL PECULIARITIES AND SPATIAL localization

The regular distribution of different compositional characteristics of diversified volcanic rocks within the Zimni Bereg area appears to be evidence for some consanguinity of these volcanics [8,9]. The most regular pattern, with evident central symmetry, characterized the distribution of Ti and Ta concentration in volcanic rock autoliths (Figure 4). Ta is an indicator of the presence of ilmenite ultramafic rocks in the deep-seated substrate since picroilmenite is the main highpressure host mineral for Ta in kimberlites [3]. Ta isoconcentration lines form a clearly concentric pattern within the Zimni Bereg area, with Ta concentration peaks shown by the rocks located in its central part and low Ta content values characterizing the volcanics occurring in the marginal zones of this area. This Ta distribution pattern agrees well with the spatial distribution of mineralogical peculiarities characterizing the local volcanics; ilmenite-bearing Fe-Ti-series kimberlitic rocks occur in the central part of the area, whereas ilmenite-free Alseries rocks and basaltic occurrences are confined to its marginal zones. Moreover, Al-series rocks themselves also show a regular change in their Ti and Ta contents with distance from the central part of the area. In particular, when passing from kimberlites of the Zolotitsky pipe cluster through Chidviya kimmelilitites to Izhma melilitites, Ta content of the rocks regularly decreases from 3-5 ppm to 2-3 ppm and to 0.7-1 ppm, respectively.



Fig. 4. Isolines of TiO<sub>2</sub> content (through 0.2 wt. %) and of Ta content (through 1 ppm) in the autoliths of the Zimni Bereg volcanic rocks.

A similar pattern, with an obvious central symmetry, characterized the distribution of Nd isotopic characteristics of the volcanic rocks in the Zimni Bereg area. This holds both for  $\epsilon$ Nd and for model age of mantle source as determined relative to depleted mantle (T<sub>(ND)</sub>DM) and to uniform chondrite reservoir (T<sub>(ND)</sub>CHUR).

As regards  $\epsilon$ Nd (Figure 5), it shows a regular change, from the values typical of weakly depleted mantle and BSE in the central part of the study area, i.e., in the zone of development of Fe-Ti-series rocks (+0.1 to +2.5), to the values typical of wearly enriched mantle shown by Zolotitsky cluster Al-series kimberlites (-2.6 to -4.9) and Chidviya and Verkhotina kimmelilitites (-6.1 to -7.2) and, further, to values typical of enriched lithospheric mantle exhibited by Izhma and Suksoma melilitites (-8.6 to -10.2).



Fig. 5. Isolines of  $\mathbf{\mathcal{E}}_{Nd}^{t}$  and of  $\mathbf{\mathcal{E}}_{Sr}^{t}$  in the autoliths of the Zimni Bereg kimberlitic rocks.

The distribution of the  $\varepsilon$ Sr parameter in Zimni Bereg volcanics (Figure 5), while largely depending on the degree of alteration of the rocks being analyzed , yet proves to show a regular trend of change, from negative weakly positive values in the central part of the area, i.e., in the zone of Fe-Ti-series kimberlites (-9.7 to +15.8), to values typical of weakly enriched mantle show by Zolotitsky cluster Alseries kimberlites (+4.2 to +18.8) and Chidviya and Verkhotina kimmelilitites (+4.8 to +24.3) and, further, to values typical of enriched lithospheric mantle exhibited by Izhma and Suksoma melilitites (+29.6 to +60.5).

The model age of mantle source determined relative to depleted mantle  $(T_{(ND)}(DM))$  shows a similar regular trend of change (Figure 6). The youngest model age characterized the Fe-Ti-series kimberlitic rocks in the central part of the area (on the average, 850 Ma); older-dated mantle sources are inferred for Zolotitsky cluster Al-series kimberlites (1250 Ma) and Chidviya and Verkhotina

kimmelilitites (1520 Ma); lastly, the oldest model age would characterized the mantle source for the Izhma and Suksoma melilitites (2110 Ma).



Fig. 6. Isolines of  $T_{(ND)}DM$  and of  $T_{(ND)}CHUR$  (Ba) in the autoliths of the Zimni Bereg kimberlitic rocks.

The model age of mantle source determined relative to uniform chondrite reservoir ( $T_{(ND)}$ CHUR), being generally about 500 Ma younger than ( $T_{(ND)}$  (DM)), exhibits the same trend of change.

# **RESULTS AND DISCUSSION**

The distribution of different compositional characteristics of the diversified kimberlitic rocks in the Zimni Bereg district is very regular, symmetric and interrelated. The Fe-Ti series kimberlitic rocks, occurring in the central part of the area, show a set of mineralogical, geochemical and Nd-Sr isotopic characteristics suggesting an asthenospheric origin for these rocks (abundance of picroilmenite along with high Ti, Nb and Ta contents, and Nd-Sr isotopic Characteristics close to BSE). Diversified Al-series kimberlitic rocks, inferredly related to old, enriched lithospheric mantle [4,5], occur in the marginal zones of the Zimni Bereg area. These rocks show a regular change in composition when passing from the very marginal zones to the zones located closer to the centre of the area. They are characterized by the following "centripetal" trends: Ti, Nb and Ta contents of the rocks gradually increase, their Nd-Sr isotopic characteristics become closer to BSE, and the model enrichment age for the mantle source relative to chondrite reservoir becomes younger (from 1480 Ma in the very marginal zones to 650 Ma in the zones located closer to the centre of the area). Along with this, the intrusion age (determined from paleofloristic data, with scarce radiogeochronic data involved [7,9]) proves to be nearly the same for all know volcanic rocks of the Zimni Bereg, namely, Late Devonian (360-374 Ma), which is generally close to the

model enrichment age of mantle source determined relative to chondrite reservoir for Fe-Ti series Zimni Bereg kimberlitic rocks.

It is interesting that the intrusion age of the V.Grib pipe, be it determined based on paleofloristic data (Late Devonian: 360-374 Ma) or by isotope dating ( Rb-Sr: 372 +/- 8 Ma), is very close to the Nd-model enrichment age of mantle source for kimberlites comprising this pipe ( $T_{(ND)}$  (CHUR) = 351 Ma) and to the age of mantle metasomatic rock xenoliths from the same pipe (Rb-Sr: 372-385 Ma), as well as to the age of garnet-ilmenite clinopyroxenite from its "scarnoid" zone, i.e., zone of contact with garnet lherzolite (Rb-Sr: 361 Ma). This fact suggests that the processes of mantle source enrichment with LREE, formation of ultramafics and metasomatic rocks of the mantle substrate, and intrusion of the V.Grib pipe kimberlites were nearly synchronous (on the geological scale) and could be interrelated .It is probable that the formation of the V.Grib pipe and other Fe-Ti-series kimberlitic rock occurrences of the Zimni Bereg area had its origin in some asthenospheric material intrusion (diapiric or local) into the lithospheric mantle in the Middle to Late Devonian. This intrusion must have been quite local (several Tens of kilometers across) and probably, it had partially tectonic contacts, but, anyway, it was capable of forming vast zones of metasomatized (reworked to a variable extent) lithospheric mantle rocks around itself, from which the diversified local Al-series kimberlitic rocks could have originated. Gradual metasomatic transformation of mantle rocks, up to complete replacement of common garnet peridotite by garnet-ilmenite peridotite, is observed in mantle rock xenoliths of the V.Grib pipe.

The spatial trend of change in model enrichment age of mantle source for different Al-series Zimni Bereg kimberlitic rocks (the closer to the centre of the area, the younger the source enrichment age) may be due to gradual change in Nd isotopic composition under the action of the young asthenospheric diapir (intrusion) rather than to several enrichment events that could have taken place at different times. In line with this gradual change, the calculated model age of mantle source enrichment also tends to change gradually, which may be an artefact. Metasomatic "steaming" of surrounding lithospheric mantle rocks by the asthenospheric diapir could cause changes in their geochemical and isotopic characteristics (gradually weakening with distance from the diapir) which in turn, could give rise to changes in geochemical and isotopic characteristics of the kimberlites that formed as a result of melting of these mantle rocks. A model section of the upper mantle of the Zimni Bereg district is shown in Figure 7. It is probable that the intensity of asthenospheric effect on the old lithospheric mantle, which results in certain complication of structure and composition of mantle rocks in the kimberlitic magma generation zone, is one of the crucial factors responsible for the diversity of kimberlitic rocks occurring in the Zimni Bereg area (and, probably, this is true for other provinces worldwide).

In other Middle-Late Devonian melilitite-kimberlitic provinces of the northern part of the Russian platform, asthenospheric effect on the mantle substrate manifests itself quite variously and, in general, less clearly than in the Zimni Bereg district.

This effect is most prominent in the Umbinsky district, Middle Timan, where kimberlites and kimmelilitites are characterized by the presence of picroilmenite, high Ti, Nb, Ta and K contents, and Nd-Sr isotopic characteristics close to BSE.

The model age of rocks comprising the Srednenskaya pipe Tnd



Fig. 7. A model section on line I-II (see figure 1) of the upper mantle of the Zimni Bereg district.

A - asthenospheric intrusion, L - lithospheric mantle. Different shading - zones with varying intensity of metasomatic working of lithospheric mantle rocks formed under the influence of an asthenospheric intrusion. Oval - the magma generation zones for different volcanic rock types. Above - character of change of isotopic-geochemical characteristics of volcanic rocks of the Zimni Bereg area on a line I-II.

(CHUR) = 355 Ma, which is essentially the same as that of the V.Grib pipe kimberlites (351 Ma) and of other Fe-Ti-series kimberlitic rocks of Zimni Bereg. In kimberlites and mililitites of the Tersky Bereg district, Kola Peninsula, the asthenospheric effect is less pronounced: these rocks contain no picroilmenite, however, being characterized by higher than average Ti, Nb, Ta and K contents and Nd-Sr characteristics close to BSE (except for the higher than average proportion of radiogenic Sr in melilitites). In the Nyonoksa district, Onega Peninsula, no sign of asthenospheric influence on the parental mantle substrate of melilitites is evident. Being strongly impoverished (depleted) in incompatible elements (including HFSE, i.e., Ti, Ta and Nb ) and isotopically enriched, these rocks are similar in geochemistry and Nd-Sr isotopic characteristics to pyroxene--free melilitites of Izhma and Suksoma in the Zimni Bereg area.

# CONCLUSIONS

The heterogeneous "eclogite-peridotite" mantle substrate of the central part of the Zimni Bereg district is comprised of garnet peridotite and clinopyroxenite, ilmenite peridotite and clinopyroxene-phlogopite metasomatic rocks, i.e., it is geochemically non-depleted, or "fertile", which may be evidence of its asthenospheric origin. This substrate obviously generated the Fe-Ti-series kimberlitic rocks, which appear to be analogues to Group 1 South African asthenospheric kimberlites. The gomogeneous "dunite" substrate, which is characteristic of the marginal zones of the Zimni Bereg district, is comprised essentially of chrome spinel dunite. This substrate is extremely geochemically "depleted", which may be evidence of its relationship to old, geochemically depleted lithospheric mantle. The diversified Al-series kimberlitic rocks that formed from the "dunite" substrate show a regular change in their compositional characteristics depending on their location within the area. As the central zone with the heterogenic "eclogite-peridotite" substrate is approached, pyroxene-free olivine melilitites (Izhma, Suksoma and others) give way to kimmelilitites and phlogopite kimmelilitites (Chidviya, Verkhotina pipes), and then to kimberlites (Zolotitsky cluster pipes). In this succession, Al-series rocks show a regular increase in Ti, Nb, Ta and K contents along with an increase in phlogopite content, while their Nd-Sr isotope characteristics grade from the values typical of isotopically enriched lithosphere to BSE. This may be due to gradual strengthening of the influence of the asthenospheric intrusion on surrounding lithospheric mantle rocks as the contact zone is approached. The intensity of asthenospheric influence on the mantle substrate, which manifests itself in kimberlite mineralogy (presence of picroilmenite ), geochemistry (Ti, Nb and Ta contents) and Nd-Sr isotopic characteristics (closeness to BSE), appears to be one of the definitive classification criteria for typifying kimberlites by composition.

It is quite probable that the wide variety of kimberlitic rocks appearing somewhat "intermediate" between the two kimberlite petrotypes by isotope chemistry and geochemistry is related to varying intensity of some infiltration and /or diffusion (i.e., magmatic and/or metasomatic) effect of an "active" asthenospheric mantle source on a "passive" lithospheric mantle substrate rather than to simple intermixing of mantle matter from two (or more) different sources. This effect may result in the formation of a mantle substrate with very complex structure and composition, both due to participation of newly formed rocks of asthenospheric origin and because of varying intensity of metasomatic effect of the asthenospheric intrusion on old lithospheric rocks. The great diversity of kimberlitic rock varieties may be essentially due to heterogeneous structure of the mantle in the kimberlitic magma generation zone. This study is illustrative of the fact that analysis of all plausible characteristics of all known volcanic rock bodies within an igneous province enables one to reveal a lot of interesting regularities, often interrelated, sometimes unapparent and, not uncommonly, even unexpected.

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# UDK 550.4:552.323.5:551.732.2 Geochemistry of within-plate basaltic lavas from the SW Paleo-Asian Ocean: evolution of plume magmatism from the Late Vendian to the Early Cambrian

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A group of seamounts existed in the south-western Paleo-Asian Ocean and was produced in Late Vendian - Early Cambrian time due to oceanic hot spot magmatism. Major and trace element concentrations and Sr-Nd isotopic compositions of pillow lavas and flow samples in the Kurai and Katun' Paleoseamounts of Gorny Altai provide a 50 Ma record of the chemistry of the Hawaiian-type magmatism manifested during the evolution of the Paleo-Asian Ocean from 600 to 550 Ma. These data demonstrate that there were large variations in the composition of magmatism over this period. Tholeiitic basalts from the Kurai paleoseamount (600 Ma), southern Gorny Altai, have lower concentrations of incompatible trace elements, and higher <sup>147</sup>Sm/<sup>144</sup>Nd ratios, compared with younger tholeiitic and alkaline lavas from the Katun' paleoseamount (550 Ma) outcropped in northern Gorny Altai. The Katun' lavas have trace element compositions similar to those of lavas from the Hawaiian Islands. The <sup>147</sup>Sm/<sup>144</sup>Nd ratio extends to a lower value compared to the Kurai lavas. We suggest that the older Kurai paleoseamount was built upon thin lithosphere closer to a former spreading centre than the younger Katun' lavas. However, the trace-element and isotopic compositions of the older Kurai lavas appear to rule out plume-ridge interaction as an explanation for their depleted compositions. The observed temporal chemical and isotopic variations may instead be due to variations in the degree of melting of a heterogeneous mantle, resulting from differences in the thickness of the oceanic lithosphere upon which the Kurai and Katun' Paleoseamounts were constructed. During the Late Vendian, when a Paleo-Asian Ocean plume was situated beneath younger, thin lithosphere, the degree of melting within the plume was greater, and incompatible trace element depleted, refractory mantle components contributed more to melting. In the Early Cambrian, i.e. 50 Ma later, the plume "tapped" the thicker lithosphere resulting in lower melting degrees and more enriched components of lower mantle heterogeneities contributed to the basaltic melt.

KEY WORDS: Kurai paleoseamount; Katun' paleoseamount; oceanic plume; lava geochemistry; lithosphere thickness; mantle heterogeneity

#### INTRODUCTION

Hotspot (within-plate) magmatism is generally thought to be the surface expression of mantle plumes—columns of relatively hot material rising from deep mantle [15]. Geochemical studies of Precambrian-Paleozoic within-plate magmatism can therefore potentially give insights into the structure and composition of the Earth's deep mantle existed at that time. Previous studies of the Kurai and Katun' Paleoseamounts (Fig. 1) which occur in Gorny Altai orogenic

belts, SW Siberia, Russia - have shown that the chemical compositions of intraplate oceanic lavas are more diverse than those of lavas from mid-ocean spreading centers of the Paleo-Asian Ocean, which sample only the uppermost mantle (e.g. [4, 10, 25]). The origin of the chemical and isotopic heterogeneity in both young and old within-plate oceanic lavas is unclear, although subduction of oceanic crust and sediment is probably responsible for generating heterogeneity in the deeper mantle [10, 11].



Fig. 1. Location of Kurai and Katun accretionary wedges and regional geology of Gorny Altai (modified from [3])

Seamount chains, created when oceanic plates move over mantle plumes, record changes in the composition of lavas erupted above a single plume over time. The study of temporal changes in the chemistry of within-plate magmatism can give us insights into the influence of the oceanic lithosphere on the chemistry of within-plate lavas [1, 6]. In order to use the geochemistry of within-plate lavas to probe the composition of the deeper mantle, it is important to know the influence

of shallow-level processes such as these on the chemistry of oceanic island lavas. The existing data show that along some seamount chains, there are significant, systematic variations in lava chemistry, whereas other hotspots appear to have erupted lavas with very similar composition over long periods of time. For example, the composition of the lavas erupted above the Hawaiian hotspot in the southern Pacific has changed significantly over the past 80 Ma. The lavas from the Hawaiian Islands erupted between 80 Ma and the present and appear to show temporal changes in trace element and isotope chemistry. Geochemical variations among lavas are explained by variations in the distance from spreading centers and/or by their eruption over the oceanic lithosphere of variable age and therefore thickness [20, 21].

In this paper, we present the results of a geochemical study of lavas from the Kurai and Katun' paleoseamounts from Gorny Altay which were formed in the SW Paleo-Asian Ocean and provide a 50 Ma record (from 600 to 550 Ma) of the geochemistry of a paleooceanic mantle plume. The Paleo-Asian Ocean existed in Vendian time between Siberia and East Gondwana and could be up to 4000 km wide (e.g. [14, 26, 33]). Its NE branch was closed in Late Carboniferous-Permian time as a result of the collision of the Kazakhstan and Siberian continents. The fragments of the oceanic crust of the Paleo-Asian Ocean, including ophiolites and plume-types basalts (oceanic island/plateau basalts – OIB/OPB) have been preserved in Altai-Sayan area (ASA) – a Caledonian collisional zone at the western and southern margins of the Siberian continent [3, 22, 25].

Previous geochemical studies of these lavas have shown that the trace elements of the lavas from the oldest, Kurai paleoseamount differ significantly from those of the younger Katun' paleoseamount [22-25]. The existing data also show that the chemical variations are largely compared with most other seamount chains ([21] and the references cited therein). We therefore carried out a detailed geochemical and preliminary isotopic study of lavas from the Kurai and Katun' Paleoseamounts, to document in detail the temporal variations in oceanic plumerelated magmatism and estimate their origin.

## AGE OF BASALTS AND SAMPLE LOCATION

The geological position of the Kurai and Katun' accretionary wedges – structures hosting the Kurai and Katun' Paleoseamounts, respectively - has been previously discussed in many publications (e.g. [2, 3, 5, 22, 25) (Fig. 1). In both structures basalts are closely associated with three types of paleoseamount sedimentary rocks: carbonate cap, slope facies with con-sedimentation (Z-shaped) folding textures and foothill sediments. Here we would like to focus on the dating of basalt-sedimentary units. The age of the Kurai paleoseamount was estimated as Late Vendian by the Pb-Pb isochronal 598±25 Ma age of the 500-m thick massive carbonate cap conformly overlying the pillow lavas [2, 29]. The dating of the carbonate unit was performed by Uchio et al. [29] in the Tokyo Institute of

Technology. The basaltic units of the Katun' paleoseamount are overlapped by slope-facies seamount sedimentary rocks. The microfossil assemblages of microphytolites, calcareous algae and siliceous sponge spicules in the overlying sediments indicate that the age of the Katun' paleoseamount is at least 550 Ma (Lower Cambrian) as was determined by Terleev et al. [28].

The previous papers of Safonova and co-authors presented the detailed description of geochemical characteristics of 21 tholeiitic basalts of the Kurai seamount and 9 lava samples from the Katun' seamount [22-25]. The samples provide a ~50 Ma record of the geochemistry of Paleo-Asian Ocean within-plate magmatism between ~600 and 550 Ma. Kurai seamount basalts were sampled at the Karatyurgun' River, east of Kurai Vil., and Katun' basalts were sampled in the Akkaya Valley, SW of Chagan-Uzin Vil., and south of Ust'-Syoma Vil. [5, 25]. The paper presents 11 more trace-element and 5 new isotopic analyzes of paleoseamount basalts from both localities.

#### SAMPLING PROCEDURES AND ANALYTICAL METHODS

Samples for geochemical analyses were obtained from the least altered and deformed outcrops of pillow-lavas and flows in the Kurai and Katun' localities (Fig. 1). All the samples were powdered using an agate mill. Major element oxides were  $100\pm1$  wt%.

All analyses except for several XRF determinations were made at the Institute of Geology, Siberian Branch, Russian Academy of Science. Abundances of major and trace elements (Rb, Sr, Ba, V, Ni, Zn, Ga, Y, Zr, Nb) were determined by XRF using a "Nauchpribor" spectrometer (analytical procedure following the Russian state analytical standard OST-41-08-212-82 Mingeo SSSR; analyst – A.D. Kireev) and SR XRF (synchrotron radiation XRF, description of technique see in Phedorin et al. [17]). Abundances of Sc, Cr, Co, Hf, Ta, Th and rare earth elements were determined by instrumental neutron activation (analyst V.A. Bobrov) using Ge detectors for Y-rays higher that 30 KeV and below 2000 KeV. For verification of obtained INAA and XRF results 29 trace elements (REE, HFSE, LILE) were determined in 3 samples by inductively coupled plasma mass spectrometry ICP-MS) using a Finnigan mass-spectrometer in the Institute of Geology, SB RAS. The observed discrepancies between INAA/XRF and ICP MS methods respectively appeared to be less than 10% for REE and less than 15% for other trace elements.

Sr and Nd isotope analyses were carried out at the Vernadskiy Institute of Geochemistry and Analytical Chemistry RAS, Moscow under supervision of Dr. Yuri Kostitsyn. The accuracies achieved were <1% for Rb/Sr, <0,01% for <sup>87</sup>Sr/<sup>86</sup>Sr, <0,1 % for Sm/Nd, < 0,005% for <sup>143</sup>Nd/<sup>144</sup>Nd. Nd and Sr isotopic ration were determined by mass spectrometry using a TRITON instrument. Rb, Sr, Sm and Nd concentrations were determined by isotope dilution. The samples were digested in HF-HNO<sub>3</sub> in titanium autoclaves equipped with fluorplastic bushes at 200°C for 2 days, also using a mixed <sup>85</sup>Rb-<sup>84</sup>Sr tracer. Rb, Sr and the rare-earth elements (REE)

were separated using a chromatography cation exchange techniques on fluorplastic columns with 3,5 ml Dowex 50×8 resin using 2,3H HCl as eluant. Nd and Sm were separated on Eichrom Ln.spec columns using 0,5H and 0,75H HCl as eluants, respectively. The SRM-987 Sr and La Jolla Nd standards gave  ${}^{87}$ Sr/ ${}^{86}$ Sr=0,710256±18 (2 $\sigma$ , n=21) and  ${}^{143}$ Nd/ ${}^{144}$ Nd=0,511843±11 (2 $\sigma$ , n=19).

#### RESULTS

#### Major and trace element composition

Major and trace element data for the Kurai and Katun' Paleoseamounts lavas are illustrated in Figs 2-4. Alteration has modified the compositions of most of the samples analyzed. As a result, concentrations of mobile trace elements in many of the samples cannot be considered as primary. We consider most LILE as relatively mobile elements during post-magmatic alteration processes. Post-eruptive growth of secondary minerals (particularly calcite, quartz, chlorite) in vesicles, together with variable replacement of olivine, orthopyroxene and plagioclase by alteration products such as saussurite, calcite and Fe-hydroxides, have affected the concentrations of some major elements such as Ca, K, Na and P, given the observed abundance of these secondary minerals in the most altered samples.



Fig. 2. Major element compositions of Kurai and Katun paleoseamount lavas. The lower TiO<sub>2</sub> and P<sub>2</sub>O<sub>5</sub> contents of the Kurai tholeiites compared to Katun' and Emperor-Hawaii lavas should be noted.

Kurai Paleoseamount (~600 Ma)

The 19 selected samples from the Kurai locality have diverse petrographic textures, ranging from fine- to medium-grained amygdaloidal and porphyritic. The samples are highly altered clinopyroxene  $\pm$  plagioclase-phyric basalts, with Mg#=66-36. TiO<sub>2</sub> and P<sub>2</sub>O<sub>5</sub> contents ranging from 0.43 wt% to 2.3 wt % and from 0.1 wt% to 0.58 wt%, respectively. The Kurai Paleoseamount lavas have lower concentrations of TiO<sub>2</sub> and P<sub>2</sub>O<sub>5</sub> than tholeiites from the Katun' Paleoseamount and from the Hawaiian tholeiites (Fig. 2).

The Kurai tholeiites have lower concentrations of the highly incompatible trace elements, e.g. Th, Nb, La, Yb, than tholeiitic basalts from the Katun' Paleoseamount and Emperor-Hawaii Chain with similar MgO (Fig. 3). Concentrations of Th and Nb lie at the depleted end of the range of the Emperor-Hawaii seamount lavas. However, concentrations of the heavy rare earth elements (HREE) in the Kurai lavas are higher than those of Hawaiian lavas (Fig. 4a), and La/Yb ratios are intermediate between those of tholeiites from the Hawaiian Islands and mid-ocean ridge basalt (N-MORB) from Sun, McDonough [27] (Fig. 3).



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The 17 samples from the Katun' locality are tholeiitic and alkaline clinopyroxene and plagioclase basaltic porphyrites containing 3.5-8.7% MgO. The Katun samples have higher TiO<sub>2</sub> and P<sub>2</sub>O<sub>5</sub> contents compared with those of the Kurai lavas and overlap with the Hawaiian lavas in all bivariant major oxide versus MgO plots (Fig. 2).



Fig. 4. Multi-component diagrams of Kurai and Katun' paleoseamount lavas (with MgO ranging from 3.9 to 7.9 and from 3.5 to 8.7 wt.%, respectively).

The bold dashed lines represent average OIB and N-MORB of Sun, McDonough [27] (upper and lower curves, respectively). The bold black lines represent typical Daikakuji (42 Ma) and Detroit (81 Ma) lavas of the Emperor-Hawaii seamount chain (upper and lower curves, respectively).

Compared with tholeiites from the Emperor-Hawaii Chain, the Katun tholeiites have similar concentrations of the highly incompatible trace elements (Figs  $\underline{3}$  and  $\underline{4}$ b). Concentrations of the most incompatible elements such as Th are lower than is classic OIB of Sun and McDonough [27] but similar to the Th concentrations in the Emperor-Hawaii seamount lavas [21] (Fig.  $\underline{3}$ ). The concentrations of the HREE (Er to Lu) are in average close to those in the Emperor-Hawaiian seamount tholeiites and their incompatible trace element patterns are similar as well (Fig.  $\underline{4}$ b). The trace element compositions of the Katun' tholeiites are like those of most other intra-plate tholeiites [30].

#### Sr-Nd isotope data

All Sr and Nd isotopic data for Kurai and Katun' paleoseamount lavas shown in the following figure have been corrected for radioactive decay since eruption;  $\varepsilon_{Sr}$  and  $\varepsilon_{Nd}$  values are reported relative to CHUR.

#### Kurai Paleoseamount

Initial  $\varepsilon_{Sr}$  values of the Kurai tholeiitic basalts are much higher than those of any tholeiites yet reported from the Emperor chain and Hawaiian Islands. The Kurai tholeiites plot in the "prohibited" right-upper quadrant in Fig. <u>5</u>. Mahoney et al. [13] observed a similar result for leached, old, highly altered basalts, and attributed this to incorporation of non-magmatic Sr into the crystal structure during replacement of plagioclase by secondary feldspar. Therefore, we interpret this as a result of post-magmatic alteration of sea-floor conditions. If we boldly project the plotted point to the vertical axes we will find them within the Hawaii field. For most samples the values of  $\varepsilon_{Sr}$  are positively correlated with K, Sr and Rb contents. This indirectly confirms that secondary alteration, in particular, kaolinization and



Fig. 5. Initial  $\varepsilon_{Sr}$  and  $\varepsilon_{Nd}$  ratios of lavas from the Kurai and Katun' Paleoseamounts, Russian Altai. Fields for tholeiitic and alkalic shield and post-shield lavas from the Hawaiian Islands (GEOROC geochemical database: http://georoc.mpchmainz.gwdg.de) are shown for comparison.

feldspathization, was possibly responsible for such abnormally high values of  $\varepsilon_{Sr}$ . Sr and Nd isotopic compositions of similar rocks from the Altai-Sayan folded area analyzed by Yarmolyuk and Kovalenko [32] as well as of Aruba oceanic plateau basalts analyzed by White et al. [31] also fall within the right-upper-quadrant range of our samples. Thus, if ignoring the post-magmatic alteration processes, the data for Kurai lavas overlap with the field defined by Hawaiian lavas in terms of their Sr–Nd isotopic compositions (Fig. 5).Field for MORB and average values for Detroit (81 Ma; triangle) and Daukakuji (42 Ma; rhomb) lavas are from [21]. The only Katun's date is within the both Hawaii fields. The Kurai field lies to the right from the mantle trend assuming a high degree of post-magmatic alteration in seafloor metamorphism conditions. If projected to the vertical axis (thus ignoring the post-eruption Sr input) the values would plot within the Hawaii field as well. Symbols see in Fig. 2.

#### Katun' Paleoseamount

The only analyzed Katun' tholeiitic sample has a higher initial  $\varepsilon_{Sr}$  and lower  $\varepsilon_{Nd}$  than all other Hawaiian–Emperor lavas from [21] (Fig. 5). On the other hand, the Sr and Nd isotopic composition of this lava sample overlaps with the field of the Hawaiian Islands in Fig. 5.



Fig. 6. Age of Kurai (600 Ma) and Katun' (550 Ma) lavas versus La/Yb, Gd/Yb, Th and Nb. Symbols see in Fig. 2.

Causes of chemical variations in Kurai and **KATUN' LAVAS** 

Figure 6 compares the compositions of lavas from the Kurai (600 Ma) and Katun' (550 Ma) Paleoseamounts following the approach proposed by Regelous et al. [21] who compared older (85-43 Ma) lavas of the Emperor Seamounts and younger lavas of the Hawaiian Chain (<43 Ma). Similarly, we tried to find out if there are any temporal chemical and isotopic changes in Paleo-Asian Ocean plume magmatism. Tholeiites from the Kurai Paleoseamount (600 Ma) have lower

concentrations of incompatible elements, and depleted trace element ratios and higher  $\varepsilon_{Nd}$  isotope values, compared with younger Katun' tholeiites. The older Kurai lavas have in average lower La/Yb and Gd/Yb ratios and lower Th and Nb contents (Fig. 6). They are similar to the oldest lavas of the Emperor Seamounts - Meiji and Detroit Seamounts of 85 and 81 Ma age, respectively. Tholeiitic and alkalic lavas from the Katun' Paleoseamount of approximately 550 Ma age have trace element compositions that lie within the range of young lavas from the Hawaiian Islands and of the youngest Emperor Seamounts (Daikakuji) (Figs 4, 6).

As far as we confidently regard the older depleted Kurai lavas as those related to mantle-plume oceanic magmatism according to geological, lithological and chemical features [5, 22-25] we must explain such a diverse composition of plumerelated lavas of the SW Paleo-Asian Ocean from two localities in Russian Altai. The lack of a representative number of Sr-Nd isotope values and the absence of Pb isotope data make determination of mantle sources problematic. That is why we will make our conclusions based on the previous studies performed by many geologists in the Emperor Seamounts and Hawaiian Chain ([21] and the references cited therein).

By analogy with previously reported interpretations we must explain the temporal changes in Paleo-Asian Ocean magmatism to account for the more depleted trace element and Nd isotopic compositions of tholeiitic lavas from the older Kurai Paleoseamount compared to the younger Katun' Paleoseamount.

Keller et al. [12] suggested that beneath young, thin oceanic lithosphere, the melting column extends to shallower depths, and the ratio of depleted asthenosphere to enriched plume mantle that is melted may be larger. Regelous suggested that intra-plate lavas erupted onto younger, thinner lithosphere were produced by larger mean degrees of melting, at shallower average depth, than melts produced beneath thicker lithosphere. This is because the overlying lithosphere acts as a 'lid' that restricts the upper boundary of the melting column during decompression melting ([21] and the references cited therein). Therefore, both the degree and the depth of melting influence the chemistry of intra-plate magmas.



Fig. 7. Variation of Lu/Hf and La/Sm in Kurai and Katun' lavas. The curves are taken from [21] and show the effect of the equilibrium melting of spinel and garnet peridotites between 0.01% and 20%.

The source (star) has primitive mantle composition [27]. Most of the younger Katun' lavas are mixtures of spinel and garnet peridotite melts formed at low to high degrees of melting. The older Kurai lavas have low La/Sm and high Lu/Hf ratios which are indicative of even higher degrees of melting within the spinel stability field.

Variations in the depth of melting may also influence the trace element chemistry of intra-plate lavas, according to how much of the melting occurs within the stability field of garnet [7, 8]. The Kurai tholeiites have low La/Sm and high Lu/Hf ratios compared with Katun' lavas (Figs <u>4</u> and <u>7</u>). The trace element compositions of the Kurai tholeiites indicate that they are the product of relatively high degrees of mantle melting, and that much of the melt was generated at low pressure, within the stability field of spinel (Fig. <u>7</u>). Larger mean degrees of melting beneath thinner oceanic lithosphere may explain the low incompatible element concentrations of Kurai lavas.

Thus, the temporal variations from Kurai to Katun' lava compositions may result from variable degrees of disequilibium melting of the heterogeneous plume mantle supposedly consisting of low melting point, incompatible element enriched heterogeneities embedded in a more depleted, refractory matrix (e.g. [9, 16, 19]), which in turn could be a result of variations in the thickness of the overlying lithosphere (Fig. 8). With progressive melting of such a heterogeneous mantle, as a result of an increasingly thin lithosphere, the melts produced would have increasingly lower incompatible trace element contents and lower ratios of moreto less-incompatible elements [18]. We suggest that the depleted compositions of the Kurai lavas are the result of melting a relatively refractory depleted component contained within the ascending plume mantle (Fig.8). That depleted component did not contribute to younger Katun' lavas, which were formed by lesser degrees of melting beneath thicker lithosphere and the resulting melt was dominated by the substance of enriched mantle heterogeneities.



Fig. 8. Scheme for the 600 Ma depleted and 550 Ma enriched hot-spot magmatism of the SW Paleo-Asian Ocean (modified from [21]).

At 600 Ma the relatively high degrees of melting result in a larger portion of incompatible element deleted refractory material in the melt, whereas at 550 Ma, when the plume is located under a thicker but younger oceanic lithosphere, the average degree of melting is lower and the melt is more contributed by incompatible element enriched less refractory material of mantle heterogeneities.

Such a tendency was first proposed by Regelous et al. [21] who to some extent confirmed it at other seamount chains: 1) the Ninetyeast Ridge (~90 Ma to 38 Ma) and Kerguelen Archipelago (~45 Ma to present) in the Indian Ocean which record changes in the chemistry of magmatism above the Kerguelen plume; 2) the Easter Seamount Chain in the eastern Pacific which age of magmatism varies from 0 Ma at Easter Island and Salas y Gomes, to ~30 Ma at its eastern end; 3) the

Louisville Seamount Chain, in the SW Pacific, ranging in age from 66 Ma at the western end, to 0 Ma at the eastern end, etc. ([21] and the references cited therein).

In summary, the geochemical variations along other seamount chains are consistent with lithospheric thickness being an important control on the extent of melting of plume mantle, and hence the compositions of the lavas produced. The hypothesis of Regelous predicts that intra-plate lava chemistry will vary with the thickness of the lithosphere and that such an effect can occur only along those seamount chains where the age difference between the seamounts and the underlying oceanic lithosphere is less than ~70 Ma. The extrapolation made in the present paper confirmed the previously observed regularities in respect to much older Kurai (600 Ma) and Katun (550 Ma) Paleoseamounts of the SW Paleo-Asian Ocean presently outcropped in Russian Altai.

# CONCLUSIONS

1. Major and trace element, and Sr isotopic analyses of 36 samples of volcanic rocks from two seamounts of the SW Paleo-Asian Ocean provide a 50 Ma record (from 600 to 550 Ma) of the geochemistry of paleooceanic magmatism, and show that there were large temporal variations in trace element chemistry and isotopic composition of the magmatism over this period.

2. Lavas from the oldest Kurai Paleoseamount have depleted incompatible trace element and Sr–Nd isotopic compositions, compared with those of younger lavas from the Katun' Paleoseamount. Initial  $\varepsilon_{Sr}$  values of the Kurai tholeiitic basalts are much higher than those of any tholeiites yet reported from the Emperor chain and Hawaiian Islands. If ignoring the post-magmatic alteration processes, the data for Kurai lavas overlap with the field defined by Hawaiian lavas. The Katun' tholeiitic sample has a higher initial  $\varepsilon_{Sr}$  and lower  $\varepsilon_{Nd}$  than all other Hawaiian–Emperor lavas. Its Sr and Nd isotopic composition falls in the field of the Hawaiian Islands.

3. The trace element and isotope compositions of these lavas vary with the age of the underlying oceanic Paleo-Asian lithosphere at the time of seamount magmatism. The oldest Kurai Paleoseamount lavas, which were erupted onto relatively young lithosphere close to a former spreading centre, have relatively depleted incompatible trace element and isotope compositions. In contrast, younger Katun' lavas were erupted onto older lithosphere, and have more enriched compositions.

4. Major and trace element compositions of Kurai Paleoseamount tholeiites indicate that they were formed by relatively large degrees of mantle melting, at lower pressures, compared with younger Katun' tholeiitic lavas.

5. By analogy with the Emperor-Hawaii seamount chain we suggest that the temporal compositional changes in the Paleo-Asian Ocean magmatism can be explained by variable degrees of melting of a heterogeneous mantle and variable lithosphere thickness. When the plume was situated beneath thin lithosphere, the melting produced Kurai melts of relatively depleted trace element compositions

due to the contribution of more refractory depleted source material. In contrast, younger Katun' lavas built on thicker crust were produced by smaller degrees of melting, and so the melts were dominated by the contribution from incompatible-element-rich, easily melted materials of mantle heterogeneities.

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# Early paleoproterozoic vulcano-plutonic komatiitic association of Southeast Fennoscandia as Mantle plume "Windybelt" realization

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A comprehensive geologic, petrologic, geochemical and isotopic study was performed for Paleoproterozoic (2.45-2.41 Ga) komatiite basalts and layered intrusions in the Vetreny Belt paleorift. A representative sections of the paleorift, including Sinegorie lava plateau, Goletz, Shapochka, B. Levgora and Kozha was drawn using an original method. The volume of the erupted mantle material was estimated as at least 30000 of cubic kilometers for the structure. Isotopic research evidences the unit of magmatites isotopic system and formation age. The model of generation and evolution of the calculated primary komatiite melt suggests its differentiation from the "head" of the Early Paleoproterozoic Windy Belt plume. It was responsible for the formation of the Vetreny Belt paleorift with its thick sequence of komatiite basalt lavas and numerous comagmatic intrusive rocks, as well as other mafic-ultramafic intrusions that were emplaced in the paleorift sides, including the Burakovo layered massif, the largest one on the Fennoscandian Shield.

## **INTRODUCTION**

Large fanerozoic basalt provinces are formed by mantle plume activity. Such Precambrian provinces including volcanic and plutonic rocks are proposed to take place on the Fennoscandian territory [7, 12, 13]. In spite of their fragmentary safety komatiitic magmatic rocks play essential role in the Fennoscandian Shield construction. Mafic-ultramafic magmatic series (komatiitic, toleitic, picritic and so on) characteristics and role in the Earth history changed. High magnesia volcanic rocks (komatiites and komatiitic basalts) are the members of the specific komatiitic series, which plays an essential part in formation of the early Precambrian series. Such rock are not typical for Fanerozoic geotectonic settings. Complex geological, petrological and geochemical research of high magnesia objects of large stable Precambrian earth crust blocks can realize the evaluation of these series role in Paleoproterozoic intraplate magmatism. Creation of the model of the parental mafic-ultramafic melt evolution from the moment of its separation from the plume at the Earth crust base to the time of the magmatic objects formation is an important aim of the research. Besides it is necessary to solve problems of the volcano-plutonic association volume and order of the formation of bodies from different depths, intracrust chambers rise, processes of crust and parental melt interaction.

One of the problem in the komatiitic rocks research is their the place in magmatic classification. TAS-diagram ( $K_2O+Na_2O - SiO_2$ ) was not success, because the role of the main classification parameters (SiO<sub>2</sub> and alkalines) are insignificant. IUGS [8] recommended classification based on the TAS diagram, with TiO<sub>2</sub> and MgO value for komatiites, meimechites and boninites. The best separation of the high magnesia rocks was realized by the help of diagram ATM (MgO - 10\*TiO<sub>2</sub>-Al<sub>2</sub>O<sub>3</sub>, Al<sub>2</sub>O<sub>3</sub>/TiO<sub>2</sub>) [4, 9].



Fig. 1. Mafic-ultramafic intrusions (1. Burakovsky; 2. Monastyrsky; 3. Scheleiny Bor; 4. Kolozersky; Fig. 1. Geological scheme of the Vetreny Belt paleorift, Southeastern Fennoscandia.

Volcano-plutonic komatiitic association Vetreny Poyas is the most representative on the fennoscandian Shield thanks to the magmatic rocks volume and variations of komatiitic rock composition. Primary magmatic minerals and magmatic glass are presented in the rocks of some intrusions, lava covers and lava lakes, because of the low stages of metamorphism. Different igneous rocks facies (both differentiated covers and subvolcanic intrusions of paleorift and deep-seated massifs from the rift shoulder, such as Monastyrsky and Burakovsko-Aganozersky). Fig. 1.

#### PALEORIFT VETRENY POYAS STRUCTURE.

Linear paleorift Vetreny Poyas, which rocks build mountain range of the same name by chain of individual hills 200-350 meters high. The hills succession is the following (from NW to SE): Goletz, Chelpan, Shapochka, Shuigora, Olovgora, Karbasnikova, Big and Small Levgora, Ovinishnaya, Myandukha et al. Nowadays the primary structure of the paleorift is changed and fragmentary destroyed during Svekofenian orogeny (1.9 - 1.8 Ga). It is the result of thrust of Arhaean Belomorian metamorphic complex over the Karelian craton from the northern-east.



# Fig. 2. Geologic structure (A) and generalized section through the komatiite basalt lava plateau (B) of the Sinegorie region (by V.S. Kulikov, V.V. Kulikova, and A.I. Zudin).

A Paleoproterozoic (Sumian): 1. komatiite basalts of the Vetreny Belt Formation; 2. nonsubdivided mafic--ultramafic intrusions; 3 layered mafic--ultramafic intrusions; 4. sedimentary rocks of the Vilenga Formation; 5. sedimentary and volcanic rocks of the Kozhozero Formation; 6. volcanic rocks of the Kirichi Formation; 7. arkoses and quartzites of the Toksha Formation. 8. Boundaries of formations, 9. some faults within the lava plateau, 10. strike and dip of lava flows, 11. areas (profiles) studied in detail and their numbers (in circles): (I) Myandukha, (II) Unaksa, (III) Vazhozero, (IV) Somba, (V) Charosa, (VI) Ovinishnaya.

B Komatiite basalts: .a. high-magnesium (>14 wt % MgO), b. low-magnesium (<14 wt % MgO).

Preserved paleorift fragment is represented by one-wing sinclinorium, including sedimentary, volcanic-sedimentary and volcanic rocks, united into 6 series [5]: tokshinskaya (terrigenic-sedimentary), kirichskaya (volcanogenic), kozhozerskaya (sedimentary-volcanogenic), kalgachinskaya (terrigenic-sedimentary), vilengskaya (tufogenic- terrigenic-sedimentary) and Vetreny Poyas (lava) (Fig. 1).

High magnesia rocks of paleorift represent lava, funnel and hypabissal facies. Lava facies is the dominant. It is constructed by flows and covers sometimes over 4 km thick. Diatremes described by V.S.Kulikov on the hills Goletz, Myandukha, B.Levgora and Bezymyannaya ("Neccovaya") and hypabissal mafic-ultramafic dykes are concerned to funnel facies. Stratum-like mafic-ultramafic bodies, lying among volcanic-sedimentary rocks, underlying Vetreny Poyas series complete Hypabissal facies. The sections of Myandukha, Goletz, Shapochka B.Levgora are research best of all.

Komatiitic basalts plateau Synegorie corps out at the left coats of the Onega river at the southeast oh the paleorift. (Fig.2).

Its common section is build as a sum of 6 individual from different blocks (from south to the north): Myandukha, Unaksa, Vazhozero, Somba, Charosa, Ovinishnaya. This blocks are characterized as simply structured and monocline rock dip to the north or north-east with dip angles 5-30°. Common section thickness was controlled by the data of gravity and magnetic researches of the Kipozero-Telza profile, which is situated some km west. Geophysical parameters show that the lava plateau thickness is 4 km [6]. Vendian, Devonian and Carboniferous sediments of Russian Plate overlay lava plateau with angle unconformity at the southeast of Fennoskandian Shield.

Two types of cover are separated in the plateau structure: differentiated and undifferentiated. Differentiated cover are represented by layers with different structure, mineral and chemical composition. Undifferentiated covers include two zones – inner massif and upper pillow. Thickness of volcanic bodies change in the diapason 3-80 m. 62 lava covers were escaped in the komatiitic basalt lava plateau 3,15 km thick. (Fig.2).

The same researches were realized for base sections of Goletz, Shapochka, B.Levgora and Kozha. (Fig. 1). Vetreny Poyas series effusive rocks separates to ultramafic (komatiites - MgO>24%) and mafic (komatiitic basalts - MgO - 9-24%).

The major minerals of komatiites with the massif structure and porfiric texture with glass are olivine and clinopyroxene. This rocks always includes chrome-spinel and magnetide. Glassy mass fills about 50 % of the rock volume. It is crystallized. Serpentine, talc, chlorite, tremolite are the typical minerals of metamorphosed rocks.

Komatiitic basalts can be divided into high magnesia basalts (HMB) and low magnesia basalts (LMB). HMB (MgO = 14 - 24%) are characterized by porphiric and spinifex texture, and massif, amygdaloidal and steak structures. Major
rockbuilding minerals are olivine, clinopyroxene (augite, dyopside), chrom-spinel. Basic plagioclase is presented in crystallized mass. Serpentine, talc, chlorite, actinolite-tremolite are the typical secondary minerals. LMB (MgO or 9 - 14%) are the effusive rocks with porfiritic texture and massif structure. Course to mineral composition they can be divided into olivinitic, pyroxenitic and plagiopyroxenitic LMB. This rocks are typical for differentiated bodies. Spinifex structures can be formed by olivine, olivine and pyroxene and by pyroxene. Secondary minerals in metamorphosed rocks are chlorite, albite, quartz, actinolite-tremolite.

A lot of intrusions take the equal place with the effusive in the paleorift structure. Data, we have got last years, confirm their comagmatic nature and the common magamatic association membership.



Fig. 3. Geological structure of volcano-plutonic association Goletz-Guiga

1. Vilengskaya series sediments; 2. Komatiitic basalts Vetreny Poyas series with spinifex amygdaloidal textures; 3. pyroxenites; 4. websterites and verlites; 5. gabbro-norites; 6. gabbro; 7. boundaries; 8. tectonic fault; 9. sample situation.

<u>Ruiga intrusion</u> (Fig. 3). Lopolitic-form massif Ruiga is situated at the northwest end of the Vetreny Poyas riftogene structure in the mouth of the Ruiga river [4]. It outcrops around Goletz komatiitic basalt, forming half-ring.

Vilenga series sediments are Guiga intrusion including rocks. While Guiga intrusion is bad denuded the boundaries between different zones are not diagnosed yet. Rocks regular order and common structure prove, that all of this rocks belong to the same differentiated body. Guiga intrusion is the subvolcanic chamber, situated 2-3 km deep. Four zones can be separated: the lowest pyroxenitic (bottom), websterite-verlitic, gabbro-norite and gabbroic.

*Pyroxenitic zone* 100 m thick is represented by metamorphosed medium- and fine-grained rocks, including monocline pyroxene and amphibole pseudomorph. Sossurite and chlorite replace the intersticial plagioclase. In spite of deuteric alteration all major minerals can be diagnosed by microscope.

*Olivine websterite and verlite zone* is 350 thick and includes fragmentary alterated rocks. On the zone bottom a layer of the pegmatiod coarse-grained websterite with olivine is situated. Its thickness is about 70 m. Rock is presented by clinopyroxenes and orthopyroxenes. Orthopyroxene grains have a large size (1.5-2 sm). Olivine is fine. Its concentration is not more than 10%. Higher in the section orthopyroxene disappears and the rocks becomes verlites. Granularity and mineral-quantity vary. Thickness of the verlite section of the zone is 280 m.



#### Fig. 4. Geological structure Scheleiny Bor intrusion.

1. chlorite-tremolite rocks, 2. skarnoids, 3. serpentinuos verlites, 4. Tremolituos Serpentinites, 5. metagabbroids, 6. gabbro, 7. outcropping boundaries, 8. faults.

*Gabbro-norite and gabbroic zones* (200 and 160 m thick accordingly) are distinguished by cumulative plagioclase appearance. Gabbro-norite zone is constituted by medium-grained gabbro-norites. Gabbroic zone is constituted by leucocratic gabbro. Rocks of these zones are alterated.

Scheleiny Bor intrusion is situated 20 km southern from the central part of paleorift Vetreny Poyas (Kozhozero Lake). Its middle section break into blocks is denuded (Fig. 4). Gabbroids represent the nuclear zone, and ultramafic rocks represent marginal zones. Black peridotites build the East Hill and their typical feature is sheet joining. The boundaries between peridotites and gabbro are tectonic. Peridotites of the West Hill are replaced by tremolitic serpentinite. Rocks of the Lower endocontact zone (30 m thick) are metamorphosed.



Fig. 5. Geological structure Burakovsky-Aganozersky intrusion [14]

5. Olovozersky; 6. Nyukhcherechensky; 7. Ruiga); 2. Komatiitic basalts Vetreny Poyas series; 3. Vilengskaya series sediments; 4. Preriftic paleoproterozoic and archean rocks; Svekofenian thrust; 6. Russian plate platform mantle boundary.



ПЕТРОХИМИЧЕСКАЯ КЛАССИФИКАЦИЯ МАГМАТИЧЕСКИХ ПОРОД

Fig. 6. Komatiitic series on the ATM diagram [9].

Some mafic-ultramafic intrusion bodies of the plutonic facies are situated on the paleorift shoulders Burakovsko-Aganozersky, Monastyrsky et al.). Burakovsko-Aganozersky intrusion is situated on the east branch of the Onega Lake. It is the European largest layered intrusion (Fig. 5). Its average composition corresponds to the Vetreny Poyas komatiitic basalt composition [7].

Researching association of volcanic and intrusive rocks are concerned to komatiitic series [9, 10, 11] (Fig. 6).

First Sm-Nd isotopic dating of komatiites from the southeast part of the Fennoscandian Shield was realized in 1997 in the whole rock samples and monomineral olivine, augite and pigonite fractions of volcanic rocks. It is a good agreement between Re-Os, Pb-Pb and Sm-Nd isochron ages, which represent 2.45-2.41  $\pm$  0.05 Ga [10, 11]. Stable isochron data made it possible to diagnose series Vetreny Poyas and underlying paleorift series as Sumian (Fig. 1): Goletz Hill – 2432 $\pm$  34 Ma; B.Levgora Hill – 2387  $\pm$  57 Ma, paleovolcano Kirich - 2437 $\pm$ 3 Ma (Fig.2). Zircon age (2449 $\pm$ 1.1 Ma) and Sm-Nd age (2355 $\pm$ 90 Mлн. лет) represented for Burakovsky layered intrusion and Avdeevskaya Dyke situated on the south-west paleorift shoulder correlate with the ages of volcanic rocks [1, 2]. Modern research of geological structure, geochemical characteristics and Rb-Sr and Sm-Nd isotopic relations begins (project RFBR 05-05-64788). Its aim is also to represent the age of different deep-seated intrusion bodies of the riftogene structure (Ruiga, Scheleiny Bor, Monastyrsky et.al).



Fig. 7. Relations <sup>147</sup>Sm/<sup>144</sup>Nd and <sup>143</sup>Nd/<sup>144</sup>Nd for minerals and WR of Ruiga intrusion.

Isotopic research of Ruiga intrusion (Rb-Sr and Sm-Nd systems). Probes of the whole rocks and monomineral olivine, pyroxenes and plagioclase fractions were researched from different rocks (laboratory of isotopic geochemistry and geochronology GEOCHI RAS). Data of isotopic relations is represented in the Fig. 7. Relations <sup>147</sup>Sm/<sup>144</sup>Nd and <sup>143</sup>Nd/<sup>144</sup>Nd for minerals and WR form linear dependence serving the age of 2360±110 Ma, (<sup>143</sup>Nd/<sup>144</sup>Nd)<sub>0</sub> = 0.50946±0.00015, MSWD=40. But this line can not be called isochron, because of considerable errors. So it is age approximate estimation, which correlate well with age data of Vetreny Poyas volcanic bodies. Relations <sup>87</sup>Sr/<sup>86</sup>Sr  $\mu$  <sup>87</sup>Rb/<sup>86</sup>Sr <sup>144</sup>Nd for minerals and WR form linear dependence serving the age of about 2 Ga. Considerable variations and errors (240 Ma) evidence isotopic system distribution. The Rb-Sr errochron "age" is significantly lower, than the crystallization age. It indicate, that the rocks were affected by the metamorfic Rb addition during late tectonic influence of Svekofennian orogeny (1.9 – 1.8 Ga).

The same researches were realized for Scheleiny Bor intrusion rocks. 4 whole rock probe were studied. Using data of the concentrations and relations we can see that in isochronic coordinates Sm-Nd form a compact group. These rocks have the similar Nd concentration, while the <sup>147</sup>Sm/<sup>144</sup>Nd variations are small. So it is not possible to build isochron. The Rb-Sr errochron "age" is 1457 ±810 Ma.

The data of  $\varepsilon_{Nd}$  show is nearly uniform -3 to 1, which can be explained by minor crustal contamination or asthenospheric material. Vetreny Poyas magmatites geological position is in the ancient rocks of Vodlozersky Block. They can become the contaminant for komatiitic magma.



Fig. 8. Relations <sup>147</sup>Sm/<sup>144</sup>Nd and <sup>143</sup>Nd/<sup>144</sup>Nd for minerals and WR of different volcanic and intrusive bodies of Vetreny Poyas paleorift.

So we can say about both comagmatism volcanic and intrusive objects and their uniform isotopic relations and age. Using data from [2] we build Sm-Nd

isochron diagram for Burakovsky Intrusion (as the most deep-seated body) and situate on this diagram data for Vetreny Poyas effusives and Ruiga and Scheleiny Bor intrusions. All points lie on the line (Fig. 8).

 $\epsilon_{Nd}$  points situation in the negative field of the diagram (Fig. 9) evidence the comtamination of asthenospheric material by crust and it is typical for all rocks of research volcano-plutonic association. So, in spite of volcanic belt and plutonic bodies are situated far from one another, we can say, that they were formed in the unit time and from the same lythospheric mantle source.



 $\label{eq:Fig. 9. end} \mbox{Fig. 9. $\epsilon_{Nd}(T)$ versus $\epsilon_{Sr}(T)$ diagram for different volcanic and intrusive bodies of Vetreny Poyas paleorift.}$ 

Contamination model primary poor magma from the astenospheric source was discussed in [10]. It was base on the quantity modeling using geochemical and isotopic data. Crust contaminanta must constitute about 4-15 mas.% to provide isotopic-geochemical variations in Vetreny Poyas magmatites.

It is very difficult to estimate the depth of the magmatic hearths. Fig.10 demonstrates one of the variants of reconstruction of intermediate chambers depth: I – plume "head"; II – Burakovsko-Aganozersky intrusion; III – Monastyrsky, Melnichny, Shardozersky massifs, may be Scheleiny Bor; IV – subvolcanic intrusions (Ruiga, Kolozersky, et. al).

The volume of the Sinegorie lava plateau is about 8190 km 3 at its present area of >2600 km and total thickness of lava sequence of >3.15 km, which is consistent with geophysical data [6]. Total volume of erupted material probably exceeds 10 000 km 3, if we also include volcanic probably eroded from the northern part of the structure and intrusive rocks. The latter values should be trebled to about 30 000 km<sup>3</sup> for the whole Vetreny Belt paleorift structure.

According to White and McKenzie (1995) [16] extensive magma production cannot be provided only by lithospheric melting and requires contribution of hightemperature material from deeper mantle horizons. According to Girnis et al. (1987), such a material could be represented by komatiitic melt with about 30 wt % MgO, which was generated by high-degree melting of mantle at P = 35-40 kbar, T = 1790-1860 C, and lg f O 2 > -7. The melt was in equilibrium with harzburgite restite. Ascending together with plume it partially lost crystallizing olivine, reached the crust boundary, and spread beneath it over an area of about 200 km in radius. The main processes of magma differentiation with fractionation of intratelluric olivine operated within this magma reservoir and its separate parts, "plume pillows". Our modeling demonstrated that differentiation of the Sinegorie komatiite melt could occurred at P = 8 kbar. The melts separated from the plume pillow corresponded in composition to komatiite basalts with 9-17 wt % MgO.



Fig. 10. Reconstruction of magma chambers situation during Windybelt Plume degradation.

Melts with lower MgO contents (<9 wt %) could be formed in shallower conditions (3-5 km), for instance, in a subvolcanic chamber, such as the Ruiga differentiated intrusion beneath Golets volcano or poorly studied Undozero, Iksa, and some other mafic-ultramafic massifs.

V.V.Kulikova and V.S.Kulikov propose a hypotheses of Vetreny Poyas volcano-plutonic association belonging to the unit riftogene system (2.45 - 2.39 Ga) of paleoproterozoic multicontinent (modern Europian and American). Probably its volcanic belts reached from northern-wets to southern-east through the

Fennoscandian territory: Tremse – Karasyok – Paanajarvi – Lehti-structure (uncluding Imandra-Varzuga) – Vetreny Poyas - ... – Kotlas (?). Their intrusive facies are represented by layered intrusions of North Finland and North Karelia, Kola peninsular and Belomorian mobile belt [15].

According to N.L.Dobretzov et. al. notion [3] the major plume "head" size must be more than 1000 km. Plumes, which "head" size is letter than 200 km, are secondary. So, Vetreny Poyas volcano-plutonic association (Burakovsko-Aganozersky intrusion – Vetrny Poyas series) is a result of hypothetical "secondary" plume WINDYBELT. Its formation can be compared with rise and evolution mechanism of Hawaiian-Impair Arc. Fennoscandian komatiitic volcanoplutonic associations research permits to propose secondary paleoproterozoic plumes moving toward the surface, such as: Windybelt, South-Laplandian, Central-Kola. derivative from Sumean thermochemical superplume of Fennoscandian Shield (may be Euro-American). Wide-scale high-temperature komatiitic magmatites formation in the same time can prove the planetary event.

## CONCLUDING REMARKS.

Using complex methods for research high-magnesia magmatites we examined different effusive and intrusive objects of Vetreny Poyas paleorift and came to the following conclusions:

1. Chemical compositions of volcanic and plutonic rocks of the paleorift Vetreny Poyas are similar and correspond to komatiites and komatiitic basalts due to IUGS reklassification [8] and ATM [9]. Age of paleorift formation and crystallization is 2.41-2.44 Ga and it is unit for all studied magmatites. Sm-Nd isotopic systems of effusives and intrusions are similar both in the axis of the paleorift and on its shoulders.

2. Magmatic rocks of the Southeast Fennoskandia are the members of the unit volcano-plutonic komatiitic association, which was formed due to the Windybelt plume activity. Magmatic material volume was about 30000 km<sup>3</sup>. So, extensive magma production cannot be provided only by lithospheric melting and requires contribution of high-temperature material from deeper mantle horizons. Isotopic researchers evidence contamination processes.

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# Geochemical and isotopic constraints for the formation model of the Kingash ultramafic and mafic complex, Eastern Sayan ridge, Central Siberia

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New geochemical and Nd- and Sr-isotopic data assumed a multi-stage model of forming for the Kingash mafic-ultramafic intrusive massif that locate among highly metamorphosed rocks of the Kansk greenstone belt in the Eastern Sayan ridge, SW Siberia. Sm-Nd isochrons for whole rocks and some minerals from main petrography associations (dunites, wehrlites and pyroxenites, metagabbro and veinlike albitites) yield three different temporal boundaries with ages of 141049 Ma for "ultrabasic" unit, 87338 Ma and 87479 Ma for "wehrlite-pyroxenite" and "basic" units and 49845 Ma for albitites. It is mainly reflected three autonomous stages of magmatic melt injection during the forming of studied complex. The Rb-Sr isotope system was significantly transformed by late metasomatic and metamorphic processes.

Initial isotopic ratios calculated for corresponding ages: NdT from -7.3 to 3.4, 87Sr/86Sr(T) from 0.7031 to 0.7061 and SrT from 4.0 to 48.2, are confirmed a mixing mater sources of parental magmas with varied degrees of combination between mantle substrates of PREMA- or DM-type and crustal material including pelagic sediments and basalts.

These isotopic features, as well as high Th, U, LILE and low Nb, Zr and HREE concentrations in ultrabasites and basites may be explained by subduction of sediments derived from oceanic crust into depleted mantle and subsequent melting of metasomatized peridotite. The comparison initial isotope parameters of rocks from "ultrabasic" and "basic" units is demonstrated a difference of geodynamic environments between these petrography associations connected with scales of back arc sea basins.

#### INTRODUCTION

The geochemistry of ultramafic and mafic igneous rocks is a clue to understand an origin of deep magmatism and to estimate a geodynamic regime of mantle activity during the geological evolution of the Earth. Compositional features of such rocks are mostly reflected a matter source of primary magmas as well as a character of mantle-crust interaction, especially, for complexes formed in intercontinental and subduction geodynamic environments. The Kingash massif including specific PGE-Ni-bearing sulfide mineralization is one of interesting natural objects in the EW folded frame of the Siberian craton. It was studied during more 40 years and constantly attracted scientist's attention so many questions about the genesis of this complex are open now. For example, some authors believe these rocks to classify as komatiites of ancient greenstone belts [13, 14]. Other researchers compare the Kingash massif with PGE-bearing layered intrusions like to the Sudbury and Monchegorsk [9, 10] or assume its multi-stage formation from different derivative melts [2, 8]. The probable reason for this situation can be a shortage of modern geochemical and isotopic data that allow constructing more precious petrologic models.

On the other hand the age of the Kingash complex is discussed too. According to the regional geological scheme, all "Kingash-type" ultrabasic and basic rocks are corresponding to Early Proterozoic magmatites of the Kansk greenstone belt with the age of 2.1 Ga [22]. The main argument for this interpretation is a location of ultramafites and mafites among oldest highly metamorphosed sediments of the Eastern Sayan region and a crossing of them by veins of plagiogranites like to Proterozoic granite intrusions. Nevertheless K-Ar and Ar-Ar isotopic dating of biotite from these plagiogranites is temporally correlated with an Ordovician stage of regional metamorphism[10, 19].

Our geochemical and isotopic data received for main petrography varieties of the Kingash deposit and their minerals assume a new point of view on the model of formation that suggests a presence of two chronologically autonomous rock associations formed in the complex geodynamic regime of suboceanic crust evolution during the Early and Late Riphean time. Moreover geochemical and isotopic parameters of studied rocks are demonstrated different scales of mantle– crust interaction in primary magma sources and some role of latest metamorphic alteration at the Early Ordovician epoch.

### GEOLOGY OF THE KINGASH DEPOSIT

There are numerous small bodies of ultramafic and mafic rocks in the Eastern Sayan ridge. They are formed a structure like to greenstone belts that can be interpretative as fragments of oceanic crust in the suture zone between the Kansk subcontinental terrain and SW margin of the Siberian Platform. The country stratigraphic sequence is present by highly metamorphosed sediments of the Late Archean Biryusa Formation including plagiogneisses and garnet-bearing schists with local layers of amphibolites and marbles. As a rule, magmatic massifs are tectonically bordered with host rocks. It is usually connected with dynamometamorphic events in the Paleozoic time [19]. Nevertheless there can be no doubt about the igneous origin of studied ultramafites so they have a typical fabric of magmatic cumulates with the obvious idiomorphism of olivine crystals relative to other minerals.

The Kingash pluton is one of largest in the Kansk greenstone belt and is situated in the upper part o the Kingash river valley. Its lens-like morphology

stretched in the NW direction is similar to stratified dunite-pyroxenite-gabbro lopolith, concordant with the enclosing metasedimentary sequence. The present subvertical position of the Kingash intrusion and host metasediments is the result of late multi-stage deformation during regional metamorphism. According to geophysical data and cross sections of drilled bore holes, the vertical order of main rock units is following: ultrabasites  $\rightarrow$  wehrlites and pyroxenites  $\rightarrow$  gabbro. However the total structure of studied massif can be also present as a combination



# Fig. 1. Geological map of the Kingash massif (Made by O.M. Glazunov using materials of the Kingash GMT and IG SB RAS)

1-2 – the Buryusa Formation (1 – gneisses and schists, 2 – marbles); 3 – "Ultrabasic" unit (dunites, pyroxene- or plagioclase-rich dunites with local layers of magnesium- high-metagabbro and vein-like albitites); 4 – the contact-reaction zone of wehrlites and pyroxenites; 5 – "Basic" unit (melanocratic and mesocratic metagabbro with local layers and veins of plagiogranites); 6 – layers of sulfide mineralization; 7 – major faults; 8 – minor faults; 9 – problematic geological boundary; 10- drilled bore holes where was checked samples for researches; 11 - samples from prospecting trenches and their numbers.

of two equal bodies of predominant dunite- and gabbro-type associations accordingly. There is only a narrow contact zone that consists of wehrlites and olivine-bearing clinopyroxenites (fig. 1). Obtained difference in volumes of main rock units is unusual for layered intrusions such as the Monchegorsk or Bushfield, but more similar to multi-stage complexes of the Ural- or Alaska-type.

The heterogeneity of ultrabasic body is manifested itself in variation of content of sulfides, pyroxene and other interstitial mineral phases as well as in small-scale horizons of amphibolites. Most authors regard these features to be elements of initial magmatic layering. However there isn't any regularity in the vertical sequence of petrography types that mostly present by pyroxene- and plagioclasebearing dunites or their serpentinized kinds. This fact is indirectly confirmed an autonomous differentiation in the "ultrabasic" unit. The textural anisotropy of gabbro is usually displayed in the gneissosity and banding with rhythmic interchange of light- and dark-colored or fine- and coarse-grained rock varieties as well as local horizons of sulfide mineralization and rare dunite xenoliths with reaction rims of wehrlites and pyroxenites. Often melanocratic gabbro is arranged for the bottom of intrusive chamber, but that can't be a direct evidence of magmatic layering. There is more regular zonality in the "wehrlite-pyroxenite" unit, where a content of olivine decreases as far as dunites change to gabbro. This tendency is corresponding to micro fabric that shows a replacement of initial olivine grains by granoblastic crystals of diopside in these rocks.

Geological features of the Kingash massif more answer to the model of its two-stage formation in conditions of successive intrusion of picritic and basaltic magmas accordingly. Their crystallization could be obviously discrete in the time so the reaction zone of wehrlite-pyroxenite composition was formed on the contact of them. The clear differentiation in main intrusive units is absent that may to demonstrate a relative quick consolidation of melts on small depths. This scenario assumes an injection of picritic melt in the "porridge"-like consistence with intratelluric crystals of olivine. Nevertheless, the crystallization of similar magma must to be certainly accompanied by a preservation of latest differentiates enriched by the interstitial matter. Residual portions of melt could be present by local horizons of plagioclase- and pyroxene-rich ultrabasites [2] or autonomous highmagnesium basites. On our opinion vein-like bodies of amphibolites inside the "ultrabasic" unit are corresponding to such rocks. Thus we can propose a presence two genetic types of basites in studied object.

Late metamorphic processes significantly complicated the initial magmatic structure of massif that made a difficult for interpretation. The progressive regional metamorphism was realized on the level of epidote-amphibolite facies and altered primary appearance of basites by means of their total amphibolization and local garnetization. This process could be accompanied with the intrusion of numerous veins of plagiogranites and albitites crossing ultrabasic and basic rocks from both magmatic units. The Ar-Ar isotopic age of such veins is changed from 480 to 590 Ma [19]. The secondary low-temperature transformation including a development of serpentine, actinolite, chlorite, clinozoisite and other hydrosilicates had rather place during more late regressive stage of metamorphism [25]. Tectonic destruction was manifested too. There are few NW- and NE-strike faults. Their kinematics isn't fully recognized, but we assume a horizontal and vertical displacement for these structures, especially for that with NE strike.

## SAMPLES AND ANALYTICAL PROCEDURE

More three hundreds samples from the Kingash deposit were selected to represent the compositional range of igneous, metasomatic and metamorphic rocks as well as sulfide ores. Low temperature metamorphic processes alter most of samples, but some of them are relatively preserved primary magmatic mineralogy. Samples denoted "Kc"-index was selected from drill holes with corresponding number ( $N_{\text{D}}$  1, 2, 3, 4, 8, 17) and on the determined depth in metros. Samples denoted "Ksh"- or "KH"-index were selected from outcrops or mining trenches (trench  $N_{\text{D}}$  6). Their localities are shown in Fig. 1. Only twelve samples were selected for geochemical and isotopic researches. Main chemical parameters of them are shown in the table 1. These rocks are corresponding to all magmatic associations including «ultrabasic» (Kc-1/272, Kc-4/77, Kc-4/164, Kc-8/83, Kc-8/85) and «basic» (Kc-3/173, Kc-3/209, Ksh-6/5, Ksh-6/10) units as well as «wehrlite-pyroxenite» reaction zone (Kc-17/129, KH-5/1) and latest veins of albitites (Kc-1/241).

Major element concentrations were determined by conventional X-ray fluorescence techniques (XRF analysis) at the United Institute of Geology, Geophysics and Mineralogy of Siberian Branch of Russian Science Academy (Novosibirsk, Russia). Trace element concentrations were measured using ICP MS techniques (at the Institute of Mineralogy, Geochemistry and Crystallography of Rare Elements RAS, Moscow) and X-ray spectrum analysis (at the Institute of Geochemistry SB RAS, Irkutsk). The Sm- and Sr-isotope data were received for whole rock samples and some mineral separates such as olivine, clinopyroxene and plagioclase from most fresh varieties. These analyses were conducted at three different science centers (Niigata University, Japan; Institute of Geochemistry and Analytical Chemistry RAS, Moscow, Russia; Institute of Geology and Geochronology of Precambrian RAS, St-Petersburg, Russia) using a Finnigan MAT-262 mass spectrometer equipped with 8-collector system following standard technique. The isotope dilution method with addition of <sup>149</sup>Sm, <sup>150</sup>Nd, <sup>84</sup>Sr, and <sup>85</sup>Rb tracers measured element concentrations. Results of measurements in samples were checked relative to isotopic ratios in BCR-1, La Jolla and SRM-987 standards.

Data regressions have been performed using the Isoplot program, version 2.0 (Ludwig, 1989) with precessions for <sup>143</sup>Nd/<sup>144</sup>Nd are no more  $\pm 0.005\%$  and external precession for <sup>147</sup>Sm/<sup>144</sup>Nd is  $\pm 0.5\%$ . Both internal and external precessions for <sup>87</sup>Sr/<sup>86</sup>Sr are  $\pm 0.01\%$  and external precessions for 87Rb/86Sr are  $\pm 1.0\%$ . Nd-isotopic compositions are given in the  $\epsilon$ Nd notation as proposed by DePaolo and

Table 1.

		(	Geochei	mical pa	arame	ters of s	studied	l samp	les			
Comple	Кс-	Кс-	10. 0/02	1/ 0/05	Кс-	Кс-	I/II <i>5/</i> 1	Ksh-	Ksh-	Кс-	Кс-	Кс-
Sample	4/77	4/164	КС-8/83	КС-0/05	1/272	17/129	КН-5/1	6/5	6/10	3/173	3/209	1/241
Rock	1	2	3	4	5	6	7	8	9	10	11	12
SiO <sub>2</sub>	38,82	42,21	37,85	44,65	41,44	44,89	37,89	48,59	46,87	47,94	43,86	63,78
TiO <sub>2</sub>	0,17	0,38	0,37	0,87	0,24	0,22	0,12	0,10	0,57	0,48	0,22	0,10
Al <sub>2</sub> O <sub>3</sub>	2,40	4,46	4,40	11,60	4,91	4,10	0,72	4,43	15,38	12,70	16,71	17,13
$Fe_2O_3^t$	13,81	14,03	13,85	14,42	12,54	9,80	14,36	7,66	10,58	8,41	9,95	2,85
MnO	0,17	0,16	н.о.	н.о.	0,18	0,12	0,12	0,25	0,16	0,12	0,11	0,05
MgO	37,50	33,35	30,66	13,50	36,22	25,23	32,38	25,90	9,47	9,85	12,33	2,50
CaO	1,40	2,83	2,89	10,60	2,20	12,65	1,91	4,47	14,89	17,70	13,70	3,57
Na <sub>2</sub> O	0,85	0,80	0,12	1,60	0,48	0,30	0,30	0,40	0,61	1,01	1,10	8,71
K <sub>2</sub> O	0,08	0,11	0,02	0,58	0,25	0,14	0,10	0,14	0,53	0,17	0,30	0,68
$P_2O_5$	0,03	0,07	0,01	0,08	< 0,03	<0,03	< 0,03	< 0,03	0,04	-	-	< 0,03
п.п.п.	3,79	2,53	8,82	2,48	0,98	1,99	10,89	7,45	1,01	1,67	2,88	0,26
Total	99,02	100,93	98,99	100,38	99,47	99,48	98,82	99,42	100,11	100,05	101,16	99,66
			N	ormativ	e miner	al compo	osition (	%)				
Fsp	10,9	16,1	14,1	31,4	16,5	13,3	4,0	15,5	47,0	40,2	52,3	84,6
CPx	3,6	4,3	2,6	23,9	0,4	42,9	8,2	10,5	28,7	46,5	22,7	11,6
OPx	-	6,0	12,8	-	3,9	-	12,2	57,6	9,0	-	-	-
Ol	80,2	72,7	69,8	28,9	78,7	41,5	75,3	16,2	14,1	10,7	17,1	0,4
TM	5,4	0,9	0,8	6,0	0,5	2,4	0,4	0,2	1,2	2,7	7,9	3,4
f(Ol,OPx)	10,6	17,5	18,2	33,1	14,7	18,2	18,8	14,0	54,4	68,0	9,7	0
An % (Pl)	37,0	67,8	95,4	62,4	77,8	83,6	13,6	74,3	87,9	85,2	81,0	10,8
				Ch	nemical	paramet	ers					
Kh	5,35	4,69	4,43	1,87	5,68	5,08	4,47	6,53	1,76	2,31	2,45	1,53
mg#	71,2	68,8	68,7	44,9	72,9	70,9	68,5	75,4	44,4	50,4	51,8	16,9
ID	8,31	7,64	1,16	18,0	5,86	3,62	3,71	4,72	8,80	9,91	11,6	79,5
CaO/Al <sub>2</sub> O <sub>3</sub>	0,58	0,63	0,66	0,91	0,45	3,09	2,65	1,01	0,97	1,39	0,82	0,21
Ka <sub>2</sub> O/Na <sub>2</sub> O	0,09	0,14	0,17	0,36	0,52	0,47	0,33	0,35	0,87	0,17	0,27	0,08
				Trace	element	contents	s (ppm)					
Cr	6400	4900	3200	116,7	4232	2200	1648	2887	454,8	408,9	182,5	28,7
Ni	4200	4600	2400	80,9	2917	1200	1218	3905	215,7	83,7	125,4	71,2
Со	160	200	180	54,3	121,9	230	265	129,0	60,3	73,0	93,7	12,1
Cu	1000	4600	970	95,1	332	570	458	99,9	57,9	188,4	72,4	29,7
Ti	1020	2100	2220	6554	1209	1360	655	706,9	3959	2068	1587	1230
V	79	260	100	374	68,5	260	67	38,6	194	211	136	47
Sc	19	-	-	77	8,6	41	_	5,9	29	42	25	-
Ti/Cr	0,2	0,4	0,7	56,2	0,3	0,6	0,4	0,2	8,7	5,1	8,7	42,9
Co/Ni	0,04	0,04	0,08	0,67	0,04	0,19	0,02	0,03	0,28	0,87	0,75	0,17
Sc/V	0,24	-	-	0,21	0,13	0,16	_	0,15	0,15	0,20	0,18	-

**Note:** Rocks: 1, 2, 5 – dunite; 3 – serpentinized pyroxene-rich dunite; 4 – fine-grained metagabbro from «ultrabasic» unit; 6 – olivine-bearing clinopyroxenite; 7 – wehrlite; 8 – melanocratic gabbro (plagioclase-bearing metawebsterite); 9-11 – metagabbro from «basic» unit; 12 – vein-like albitite.

Symbols for normative mineral composition: Fsp – field spar; Cpx – clinopyroxene; Opx – orthopyroxene; Ol – olivine;  $f_{(ol, Opx)}$  – ferrous index for olivine or orthopyroxene in molecular % content of Fa or Fs end members respectively; An % (Pl) – content of An end member in the plagioclase. Chemical parameters: Kh – magnesian index by Hess (Kh=MgO/FeO<sup>t</sup> in molecular %); #mg – crystallization index by Kuno (#mg=100\*MgO/ (MgO+FeO+Fe2O3+Na2O+K2O)); ID – differentiation index by Thornton and Tuttle (the sum of normative albite and orthoclase contents). Analysis of rock samples № 1-3 were conducted at the IG SB RAS (Irkutsk, Russia) using silicate chemical (analysist is T.V. Ozhogina) and atom absorption (analysist is O.A. Froidakova) methods. Samples № 4-12 were analyzed at the UIGGM SB RAS (Novosibirsk, Russia; analysist – N.M. Glukhova) and at the IMGRE RAS (Moscow, Russia; analysist is D.Z. Zhuravlev) using RFA and ICP MS techniques respectively.

Wasserburg [3] with values calculated using CHUR parameters of Faure [7]. Srisotopic compositions are given in the initial ratio for corresponding age and in the  $\varepsilon$ Sr notation with values calculated using the UR parameters after. De Paolo [4]. The model age T<sub>DM</sub> was calculated on the base of modern depleted mantle parameters <sup>143</sup>Nd/<sup>144</sup>Nd=0.51315 and <sup>147</sup>Sm/<sup>144</sup>Nd=0.2137 after Peucat et al.[20].

### **GEOCHEMISTRY OF STUDIED ROCKS**

The distribution of major and minor elements in rocks of the Kingash massif has been described by many workers [2, 10, 13, 17 and other]. Usually they studied patterns of siderophile, chalcophile and precious metals but not REE and HFSE. Most of authors show a convergence of rock compositions with products of komatilitic or melapicrite-basaltic series as well as an inheritance of chemical parameters between main petrography varieties. Nevertheless, two types were divided among gabbroides that different between themselves in the level of TiO<sub>2</sub>.



Fig. 2. Variation of some geochemical parameters versus a magnesian index by Kuno for main rock types of the Kingash massif

Symbols: 1 – basites of "ultrabasic" unit; 2 – basites of private gabbroid intrusion; 3 – compositional field of rock association (1 - dunites, 2 - wehrlites and pyroxenites, 3 - gabbro, 4 - albitites and tronhjemites); 4 – variety trends of basites.

The two-stage model of the Kingash massif formation is according to the bimodal distribution of Ti between final mafic differentiates of picritic magma and rocks of private gabbroid intrusion. We took attention on the variety of some geochemical parameters (CaO/Al<sub>2</sub>O<sub>3</sub>, TiO<sub>2</sub>, Ti/Cr. Sc/V) relative to a magnesian index after

Kuno (#mg=100MgO/ (MgO+FeO+Fe<sub>2</sub>O<sub>3</sub>+Na<sub>2</sub>O+K<sub>2</sub>O)). Two trends are actually obtained for different basic associations (fig. 2). Figurative points of "ultramafic" metagabbro are distributed along the first trend, but points of private gabbrioc intrusion are mostly showed a tendency to stretch along the second trend. However this regularity isn't directly obvious so some samples of basites from the gabbroid body are located near the first variation line. It may be connected with chemical heterogeneity depending on a presence of local xenolith-like blocks of "ultrabasitic" series in the late gabbroic intrusion.

Our new geochemical data allow estimating peculiarities of fractional crystallization and geodynamic regime of formation for the Kingash massif. REE-patterns of studied rocks show a successive growth of content from dunites to pyroxenites and far to mafites (fig. 3). Some features of REE accumulation are for different types of gabbro. For example the spectrum line of metagabbro from the "ultrabasic" unit (sample Kc-8/85) opposite to "private" metagabbro (samples Kc-3/209, Ksh-6/10, Ksh-6/12) is characterized by a negative slope with a local depletion of LREE. This pattern is mostly conjugate to that of plagioclase-bearing dunite (sample Kc-1/272). It is partly confirmed a residual origin of similar basites from the parental picritic magma. Usually the degree of REE fractionation is low in Ti-rich rocks. This regularity was already remarked for basites of the Kansk greenstone belt as an indicator of different formation type or structural setting [17].



Fig. 3. Chondrite-normalized REE patterns for rocks of the Kingash massif. Chondrite REE concentrations are given after Evensen et al. (1978)

Symbols: 1 - rocks of ultrabasic unit; 2 - rocks of "wehrlite-pyroxenite" reaction zone; 3 rocks of gabbroic body; 4 - albitites of latest veins.

The distribution of REE in studied metagabbro and pyroxenites (sample Kc-17/129) hasn't Eu anomalies that refuse any role of plagioclase fractionation during crystallization of basic melts. The weakly negative anomaly of this element in some dunites (sample Kc-1/86) and wehrlites (sample KH-5/1) could reflect an early accumulation of olivine in the parental picritic magma. At the same time strongly positive Eu anomaly in albitites (sample Kc-1/241) and negative that in melanocratic gabbro (sample Ksh-6/5) may be to connect with metamorphic transformation of rocks such as secondary bedding or portioned melting.

Table 2.

		Kar	e elem	ent dat	a (ppm	i) for st	ualea r	OCKS			
Sampla	Kc-	Kc-	Kc-	Ksh-	Ksh-	Ksh-	Kc-	Kc-	Kc-	KH-	Kc-
Sample	1/86	1/272	8/85	6/5	6/10	6/12	3/209	3/173	17/129	5/1	1/241
Rock	1	1	2	4	5	5	5	5	3	3	6
Cs	0,227	0,153	0,155	0,165	0,098	0,142	0,512	0,052	0,207	0,086	0,103
Rb	3,461	4,573	35,82	1,07	11,91	19,68	18,10	3,43	3,402	1,091	6,504
Ba	24,154	34,973	744,8	133,01	147,07	143,76	166,43	40,67	39,82	8,548	431,9
Th	0,309	0,347	0,437	0,263	1,289	0,575	0,837	0,288	0,439	0,152	0,183
U	0,071	0,073	0,119	0,107	0,315	0,112	0,205	0,105	0,140	0,048	0,068
Sr	12,972	29,701	503,3	23,42	392,2	422,4	311,9	309,5	46,5	11,02	1002,5
Zr	11,853	10,251	49,09	4,722	41,99	32,48	22,63	17,35	14,23	9,007	3,618
Hf	0,333	0,306	1,89	0,158	1,306	1,169	0,639	0,603	0,506	0,261	0,117
Nb	0,84	0,919	2,145	1,359	3,244	6,025	7,164	0,921	0,815	0,881	15,645
Та	0,168	0,179	0,191	0,178	0,261	0,48	0,406	0,053	0,066	0,034	1,661
La	1,026	1,742	2,906	1,132	6,148	5,28	6,45	1,95	1,968	0,693	1,931
Ce	2,368	3,775	7,78	2,589	13,89	13,17	12,75	5,04	4,355	1,673	5,812
Pr	0,301	0,469	1,275	0,32	1,89	1,899	1,422	0,77	0,565	0,206	0,921
Nd	1,31	1,982	6,883	1,57	8,104	9,532	5,559	3,81	2,602	0,921	3,991
Sm	0,324	0,477	2,538	0,505	2,095	2,926	1,210	1,28	0,838	0,256	0,771
Eu	0,072	0,171	1,022	0,104	0,734	1,103	0,406	0,515	0,272	0,076	0,911
Gd	0,381	0,576	3,688	0,610	2,264	3,906	1,167	1,59	1,064	0,313	0,596
Tb	0,064	0,100	0,677	0,111	0,381	0,697	0,198	0,285	0,197	0,057	0,089
Dy	0,403	0,625	4,591	0,724	2,303	4,495	0,195	1,772	1,214	0,351	0,455
Но	0,089	0,143	1,012	0,163	0,495	1,028	0,241	0,361	0,248	0,077	0,088
Er	0,231	0,384	2,901	0,435	1,273	2,784	0,660	1,003	0,683	0,216	0,233
Tm	0,033	0,060	0,425	0,069	0,177	0,399	0,095	0,142	0,095	0,032	0,033
Yb	0,218	0,377	2,763	0,444	1,127	2,554	0,573	0,835	0,582	0,210	0,197
Lu	0,034	0,061	0,423	0,062	0,159	0,368	0,081	0,12	0,083	0,034	0,028
Y	2,401	3,88	23,28	4,41	13,68	27,64	7,09	10,69	7,37	2,294	2,72

Rare element data (ppm) for studied rocks

**Note**:Rocks: 1 – dunite; 2 – metgabbro from "ultrabsic" unit; 3 – wehrlite and pyroxenite of reaction zone; 4 – melanocratic metagabbro (plagioclase-bearing metawebsterite); 5 – metagabbro from "basic" unit; 6 – vein-like albitite. Element concentrations was determined at the Institute of Mineralogy, Geochemistry and Crystallochemistry of Rare Elements of RAS (IMGRE, Moscow, Russia) using ICP MS technique Finnigan mass spectrometer, analysist is D.Z. Zhuravlev.

The variety of rare element contents normalized to the primitive mantle (PM) and NMOR-type basalt is shown in fig. 4. The enrichment by "crust" components (Cs, Rb, Ba, Th, and U) as well as depletion by Nb, Zr and HREE assumes some

role of subduction zone during a generation of primary magmas. It is corresponding to the point of view about an initial formation of protolith for the Kansk greenstone belt in the regime of back-arc spreading [17]. The level of Zr, Hf, and Nb depletion is usually more strongly for rocks from the private gabbroidic body. We regard to make that depend on different scales of back-arc sea basin and velocity of spreading during generation of parental magmas for two main intrusive phases of the Kingash massif. The distribution of Sr is contrast in ultrabasites (lower to PM) and basites (higher to MORB) that can be explained as a relative mobility of this element in metamorphic processes.



Fig. 4. Spider-diagrams of trace element distribution for some samples of ultramafites (A) and mafites (B) from the Kingash deposit

### SM-ND AND RB-SR CHRONOLOGY

Sm-Nd and Rb-Sr isotopic data for whole rock samples and mineral separates are listed in table 3. The Sm-Nd isotopic system is more stable for metamorphic alterations so it can be more informative for dating of private magmatic events.



Fig. 5. The variety of Sm-Nd and Rb-Sr isotope composition in rocks of the Kingash ultramafic-mafic massif

A – Sm-Nd isochrones for whole rock samples from "ultrabasic" unit and for the vein-like albitite spatially associated with them; B – Rb-Sr isochron for dunites of "ultrabasic" unit; C – Sm-Nd isochrones for the "wehrlite-pyroxenite" reaction unit and for rocks of private gabbroid intrusion; D – a variation of Rb-Sr isotope parameters in rocks of "wehrlite-pyroxenite" and "basic" units.

Symbols: 1 – ultramafites (dunites, wehrlites, pyroxenites); 2 – basites; 3 – albitites.

			S	m-Nd	and Rl	b-Sr analyt	ical data and	mod	al par	ameter	S		E	ıble 3
Š	Sample	Rock	Тип пробы	Sm (ppm)	(uudd)	<sup>147</sup> Sm/ <sup>144</sup> Nd	<sup>143</sup> Nd/ <sup>144</sup> Nd	εNd	Rb (ppm)	Sr (ppm)	<sup>87</sup> Rb/ <sup>86</sup> Sr	<sup>87</sup> Sr/ <sup>86</sup> Sr <sub>H3M</sub>	$^{87}\mathrm{Sr}/^{86}\mathrm{Sr}_{(0)}^{*}$	ESr <sub>1</sub> *
	Kc- 4/77	Dunite	Wr	0,32	1,15	0,168223	0,512547±7	+3,4	0,91	18,15,	0,14504	$0,706043\pm10$	0,704232	+10,9
2*	Kc- 4/164	Plagioclase- bearing dunite	Wr	0,82	2,84	0,174556	0,512597±12	+3,3	2,78	46,37	0,17344	0,706472±8	0,704307	+11,9
3	Kc- 1/272	- // -	Wr	0,485	1,92	0,15260	0,512384±6	+3,1	4,28	35,10	0,35320	0,708710±15	0,704300	+11,8
*	Kc- 8/83	Pyroxene-rich dunite	Wr	0,79	2,54	0,188037	0,512715±12	+3,1	1,79	19,57	0,26463	0,707531±7	0,704227	+10,8
5*	Kc- 8/85	Metagabbro	Wr	2,53	6,92	0,221053	0,513021±7	+3,1	25,18	453,7	0,16057	0,706648±7	0,704643	+16,7
9	KH-5/1	Wehrlite	Wr	0,21	0,69	0,18200	$0,512600\pm9$	+0,9	0,60	5,9	0,29270	$0,708786\pm22$	0,705132	+23,7
2	Kc- 17/129	Olivine- bearing clinopyroxenite	Wr	0,666	1,938	0,20780	0,512736±9	+0,7	3,1	47	0,18760	0,707382±7	0,705040	+22,3
8	Kc- 17/129	- // -	CPx	0,595	1,521	0,23640	0,512900±7	+0,7	I	Ţ	ı	I	I	I.
6	Kc- 17/129	-//-	0	0,74	2,75	0,16210	0,512470±7	+0,6	ı	ı	ı	I	I	ı
10**	*Kc- 1/241	Vein-like albitite	Wr	0,200	1,297	0,09345	$0,511924\pm12$	-7,3	8,41	1529	0,0159	0,705767±10	0,705654	+24,7
11**	*Kc- 1/241	- //	CPx	0,346	1,423	0,14688	$0,512109\pm6$	-7,0	I	ŗ	I	I	I	ı
12	Kc- 1/241	- // -	PI	0,095	0,911	0,06280	0,511841±13	-7,0	ı		ı	I	I	
13**	*Ksh-6/5	Melanocratic gabbro (Plagioclase- bearing websterite)	Wr	0,527	1,639	0,19451	0,512629±7	+0,1	1,5	19,4	0,22370	0,707896±11	0,705103	+23,2
14	Ksh- 6/10	Metagabbro	Wr	1,59	5,46	0,17550	0,512511±5	0,0	11,16	548,0	0,05900	0,706861±	0,706124	+48,2
15*	Kc- 3/173	-//-	Wr	1,34	4,14	0,19568	0,512632±11	0,0	2,61	269,6	0,02800	0,706118±7	0,705768	+32,3
$16^{*}$	Kc- 3/209	- // -	Wr	0,68	1,94	0,211913	0,512717±7	-0,1	6,54	345,7	0,05473	0,706167±7	0,705484	+28,7
Not *** S anal valı anal 16) wer (for	e: * San Samples Samples Jytical d Les of N Lysis N <u>e</u> and to 7 e calcul vein-lik	mples were and were analyzed ata was receiv- id_r were calcu 1-5), to $T=875\Gamma=498 Ma (forated for the ag-ce albitite).$	llyzed at the ed at th lated f Ma (f albititk e of 87	at the GEO or the or rocl ss, ana 5 Ma	Niigat CHE F BP (St age cc ks of p llyses ] (for ul	a Universit AAS (Mosc Peterburg, prrespondin private gabl N <u>e</u> 11-13) tramafites a	y (Japan) af ow, analysis analysist is g to T=141( proid intrusi Initial Sr-isc and mafites)	ter T. st is A V.P. K O Ma on, an otopic and	Moril Moril Covacl (for "u nalyse tratios for the	kiyo. etrova hiltraba s $N_0 6$ - s and $\epsilon$ age o	). Other sic" unit, 9 and 14- Sr <sub>r</sub> values f 498 Ma			

In the Sm-Nd evolutionary diagram, data points of bulk compositions represented by dunites (Kc-1/272; Kc-4/77; Kc-4/164; Kc-8/83) and metagabbro (Kc-8/85) from the "ultrabasic" unit are carefully approximated (MSWD=1.01) by regression line with a slope corresponding to an age of 1410±49 Ma (fig. 5a). The  $\epsilon Nd_T$  of this line is +3.2. The revealed age can be regarded a final stage of consolidation of ultrabasic intrusive body so geochemistry of studied rocks is reflected features of picritic melt differentiation in the magmatic chamber. However the internal isochron constructed for data points of plagioclase, whole rock sample and augite from the vein-like albitite (Kc-1/241) is only shown the Early Ordovician metamorphic events with age of 498 $\pm$ 45 Ma (MSWD=0.99;  $\epsilon$ Nd<sub>T</sub>=-7.1). This isotopic dating is mostly corresponding to that received for biotites from host gneisses and vein-like trondhjemite (T=480-520 Ma) using the Ar-Ar isotopic method [19]. The distribution of Rb and Sr isotopes is less regular in studied rocks, especially in basites (fig. 5b). A regression line with parameters T=882±33 Ma, MSWD=1.75 and  ${}^{87}$ Sr/ ${}^{86}$ Sr<sub>(0)</sub>=0.704239 can approximate only bulk compositions of dunites. The deviation of figurative point of metagabbro (Kc-8/85) is obviously connected with the secondary amphibolization accompanied by extraction of <sup>87</sup>Sr from metasedimentary units of the Biryusa Formation.

The Sm-Nd isochron line calculated for four points of whole rocks and minerals from reaction zone (bulk composition of wehrlite denoted KH-5/1; olivine, diopside and whole rock composition of clinopyroxenite denoted Kc-17/129) is regained direct balance too (MSWD=0.89). Its slope corresponds to age of  $873\pm38$ Ma with  $\epsilon Nd_T = +0.7$  (fig. 5c). The variety of isotopic data in rocks of "private" gabbroid intrusion is less regular so a deviation of Sm/Nd ratio is very short in these samples. Nevertheless, the regression line calculated for whole rock compositions of this gabbro has a slope corresponding to similar age of 874±79 Ma with MSWD=0.64 and  $\epsilon Nd_T=0$ . The successive decrease of  $\epsilon Nd_T$  values from ultrabasites to pyroxenites and far to gabbro is confirmed a "contact-reaction" origin of total rock sequence in the Kingash massif. There is a significant transformation of Rb-Sr isotope system in listed samples also (fig. 5d). More weak deviation from regression line corresponding to age of 875 Ma is only obtained for figurative points of wehrlite (KH-5/1), olivine-bearing clinopiroxenite (Kc-17/129) and plagioclase-bearing metawebsterite (Ksh-6/5). At the same time basites are characterized by different scales of contamination by radiogenic strontium.

Our results of isotopic dating allow distinguishing three main stages in the evolutionary history of the Kingash massif. The first of them is corresponding to the Early Riphean (Early Neoproterozoic) time (T=1410 Ma) when the "ultrabasic" unit was formed. The equilibrium distribution of Sm and Nd isotopes in dunites and vein-like bodies of Ti-rich gabbro spatially associated with them confirm it. Following Late Riphean stage was remarked as intrusion and emplacement of

"private" gabbroid plutonic body accompanied by active thermal and metasomatic interaction with the early ultrabasites and forming wehrlite-pyroxenite reaction zone. Such scenario is according to the similarity of Sm-Nd dating for gabbro (T=874 Ma) and pyroxenites (T=873 Ma). Moreover these events was reflected in the transformation of Rb-Sr isotope system in dunites with near age of 882±32 Ma. The main mechanism of this reorganization could be a local metasomatic alteration with the development of secondary phlogopite crystals formed reaction rims around olivine and ore minerals or located in the interstitial space. The third stage was connected with injection albitite and trondhjemite veins that correspond to Early Ordovician epoch of regional metamorphism. Geological chronometers of these events are Sm-Nd and Ar-Ar isotope systems of leucocratic vein-like rocks in the Kingash massif.

# MATTER SOURCES AND GEODYNAMIC REGIME OF FORMATION

The high mobility of LIL-elements (Rb, Sr, Ba) during a secondary amphibolization [1] is confined to estimate matter sources of primary magmas for rocks from the Kingash massif using only initial <sup>143</sup>Nd/<sup>144</sup>Nd ratios. Values of  $\epsilon$ Nd<sub>T</sub> calculated for corresponding ages of different rock association are formed four independent groups that range to following intervals: a) from +3.1 to +3.4 in varieties of parental picritic magma; b) from +0.6 to +0.9 in rocks and minerals of wehrlite-pyroxenite "contact – reaction" unit; c) from –0.1 to +0.1 in "private" gabbroides; d) from –7.3 to –7.0 in bulk composition and minerals vein-like albitite (Kc-1/241). Such discrete variation of  $\epsilon$ Nd<sub>T</sub> supposes an obvious heterogeneity of mater sources for rocks in studied plutonic bodies. The decrease of this parameter in more young intrusive phases is shown a successive growth of participant role of crustal components in the generation of magmatic melts.

A lowest level of crustal contamination is characterized for ultrabasites. However,  $\epsilon Nd_T$  values obtained in them must to assume a mixing origin of initial matter source with Nd isotopic composition corresponding to the combination of depleted (DM– or PREMA-type) and enriched (EMI– or EMII-type) mantle substrates. Model Nd ages calculated for DM varies from 1.9 to 2.1 Ga in this rocks. It is a little younger relative to that for orthogneisses from the Kansk greenstone belt ( $T_{DM} = 2.4$  Ga). The dating of gneiss protolith using U-Pb isotope method shows an age of 2.3 Ga with  $\epsilon Nd_T =+3$  [18]. Similar Nd isotopic composition in ultrabasites of Kingash massif and metavolcanites of greenstone belts are demonstrated a multi-stage melting of uniform mantle substrate during Early Proterozoic and Early Riphean time.

The mixing model is necessary to explain features of Nd isotopic composition in rocks of "private" gabbroid intrusion too. The subduction-relative genesis of them is obviously remarked in geochemical and isotopic parameters. Moreover the emplacement gabbroic body is temporally correlated with main stages of tectonic evolution of Arzyba neighboring terrain that consist from juvenile crust complexes of oceanic arc type[21, 24]. Its arc-like forming is supposed at Grenville ocean on the boundary of 1000 - 1100 Ma, but its collision accretion to the Kansk terrain was later about 800 Ma [19]. Our chronological data for "private" gabbroid of the Kingash massif (T = 870 –900 Ma) is coordinated with a middle stage during the closing of this oceanic basin. According to such scenario, the basaltic magmatism could be connected with processes of back-arc spreading or local rifting on the continental margin and be accompanied with tectonic evolution of the Kansk terrain.

Slightly positive values of  $\varepsilon Nd_T$  in rocks of wehrlite-pyroxenite unit correspond to the model of their forming in conditions of contact interaction between solid ultrabasites and basic melt. Negative values of  $\varepsilon Nd_T$  in bulk composition and minerals of vein-like albitite are reflected increasing role of crust during Ordovician stage of regional metamorphism.

The Rb-Sr isotopic system is unstable during metamorphic alterations of rocks, but it could be more informative for reconstruction of geochemical evolution on latest stages of history. For example, the distribution of Rb and Sr isotopes in studied dunites was obviously transformed by influence of fluids from "private" gabbroid intrusion. We suppose to connect this changing with an addition of some <sup>87</sup>Rb content that could alter the initial distribution of Sr isotopes. In this case, samples with low Rb/Sr ratio must to mostly preserve Sr isotope composition similar to that in picritic parental magma. The of <sup>87</sup>Sr/<sup>86</sup>Sr<sub>(0)</sub> in sample Kc-4/77 calculated for age of 1410 Ma is 0.703113 ( $\epsilon$ Sr<sub>T</sub> =+4.0). At the same time Sr-isotopic composition of ultrabasites and pyroxenites calculated for age of 875 Ma could be more similar to that of parental basic magma with <sup>87</sup>Sr/<sup>86</sup>Sr<sub>(0)</sub> value ranging from 0.7064 to 0.7074. It is corresponding to variety of  $\epsilon$ Sr<sub>T</sub> from +11 to +22.

The Rb-Sr isotopic system was mostly transformed during regional metamorphism, when Sr-rich mineral phases (clinopiroxene and plagioclase) was undergoing to amphbolization and albitizatoin. The strong variety of <sup>87</sup>Sr/<sup>86</sup>Sr<sub>(0)</sub> value in basites is a regular argument for such interpretation. Nevertheless all rocks of the Kingash massif have relatively low values of  $\varepsilon$ Sr<sub>T</sub> (no more +50) that demonstrate a specify of isotope composition in host metamorphic units as well as origin of metamorphic fluids from the oceanic crust. The moderate  $\varepsilon$ Sr<sub>T</sub> = +24.7 in albitite similar to that in rocks of "wehrlite-pyroxenite" units may be shown on the forming of such veins by portion melting of ultramafite-mafite substrate, but not a relationship with the Ordovician granite magmatism in the Kansk terrain.

The composition of  $\varepsilon Nd_T$  and  $\varepsilon Sr_T$  values allow to distinguish two main regularities for variations of isotope composition in studied rocks (fig.6). The significant range of  $\varepsilon Sr_T$  (trend 2) is mostly connected with metamorphic alteration accompanied either with addition of <sup>87</sup>Rb from more late gabbroid intrusion (in case of dunites) or with alien <sup>87</sup>Sr supplement from host metasedimentary units during secondary amphibolization of mafites. The removal of such influence gives us to construct an evolutionary trend for private magmatic systems (trend 1). A slope of this trend corresponds to mantle array that reflect successive involving of crustal material in processes of melting in order from "ultrabasic" to "basic" unit and far to veins of albitites. Some <sup>87</sup>Sr enrichment in rocks of the Kingash massif relative to mantle-type varieties could be caused by action of slab-derived fluids with isotope composition similar to oceanic metasediments. Their participation in the magma generation is confirmed a subductional origin of main rock associations in studied massif.



Fig. 6. **ENd<sub>T</sub> vs. ESr<sub>T</sub> plot for the Kingash ultramafic and mafic rocks and vein-like albitites** 

Symbols: 1 – dunites of «ultrabasic» unit; 2 – metagabbro of «ultrabasic» unit; 3 – wehrlites and pyroxenites of contact-reaction zone; 4 – metagabbro of «basic» unit; 5 – hypothetical initial composition of rocks from «ultrabasic» unit; 6 – hypothetical initial composition of rocks from «basic» unit; 7 – isotopic composition of albitites from vein-like series; 8 – hypothetical trend of isotopic evolution for magma generated reservoirs; 9 – trend of metamorphic alteration; 10 – mantle reservoirs (DM – depleted mantle, HIMU – mantle substrate under oceanic islands enriched by Th and U relative to Pb, BSE – Bulk Silicate Earth, EMI and EMII – lithosphere mantle enriched by components of lower and upper crust respectively; 11 – field of middle ocean ridge basalts (MORB).

Location of mantle reservoirs and MORB are shown according to their modern parameters after A. Zindler and S. Hart [26].

### CONCLUSION

Results of our geochemical and isotopic researches allow to make following main inferences about geodynamic regime of formation and genesis of rock associations at the Kingash PGE-Cu-Ni-bearing deposit.

At the first, this object presents itself a multi-formation intrusive massif where is spatially combined different magmatic units. Its origin could be interpreted as successive injection of two autonomous magmas of picritic and basaltic composition at the time of 1410 and 875 Ma respectively. Obtained sequence of rock kinds (dunites  $\rightarrow$  pyroxenites  $\rightarrow$  gabbro) in the vertical cross section of massif is a result of contact interaction of these melts, but not a result of magmatic differentiation in intrusive chamber.

At the second, matter sources of total rock association are characterized by discrete range of isotopic parameters with four independent intervals of  $\epsilon Nd_T$  values corresponding to "ultrabasic", "wehrlite-pyroxenite", "basic" and "albitite-plagiogranite" units. This distribution is obviously reflected different degrees of mixing between prevalent plume-type (PREMA) or depleted (DM) mantle substrates and material of enriched lithosphere mantle (EMI) or oceanic crust (metasediments and basalts) during magma generation. The Rb-Sr isotopic system was significantly transformed by secondary metamorphic reorganization of latest history including a formation of final ore mineralization.

At the third, features of Sr and Nd isotope composition as well as patterns of HFS- and RE-elements in studied rocks are assumed an active participation of subductional processes in the geodynamic model of the Kingash ultramafite-mafite complex forming. The role of subduction is less for generation of parental picritic melt (rocks of "ultrabasic" unit) that could be connected with more wide back-arc sea basin in Early Riphean time. Rocks of "basic" unit are mostly correlated with products of island arc- or continental margin-type magmatism in rifting zones.

At the fourth, the Ordovician stage of regional metamorphism was enough significant for total history of the Kingash deposit. The vein-like complex of albitites and plagiogranites can be interpreted as evidence of local portion melting of ultramafites and mafites respectively. This conclusion is confined by relatively low  $\varepsilon Sr_T$  values and positive anomaly of Eu in albitite veins from "ultrabasic" unit that is contrast to that in melanocratic gabbro.

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# UDK 552.33 Metamorphism of the Veseloe carbonatites, North Transbaikalia (Russia)

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### INTRODUCTION

The carbonatite occurrences are which undergo the process of metamorphism are not wide spread.

This is carbonatite occurrence in Mount Grace (British Columbia), where carbonatites were metamorphized on the level of amphibolite facies. The bodyes of alkaline rocks were modify into gneisses. The carbonatites were recrystallized into coarse-grained granoblastic rocks [4]. Metamorphism of Gulinsky carbonatites (Russia) was obtained Zhabin [10]. According to Zhabin, the main process which occured during metamorphism was the recrystallization of the carbonatite matrix that was similar to the formation of marbles at the expense of limestones.

The discussion [8] about formation of Celigdar deposit (Russia) is probably related with it's participation in the process of metamorphism. It is likely that the metamorphic processes were caused by the of tremolites, talcs, metacrysts of apatites, the refined of minerals from admixtures and unusual  $\delta^{18}O\%$  and  $\delta^{13}C\%$  isotopic compositions.

As any other rocks, which were to metamorphism, the carbonatites there must here: a) the processes of recrystallization undegro of minerals with the change of structure and texture of rocks; b) the exchange of primary and appearance of new mineral species; c) the change of chemical compositions and some isotopic values of the minerals.

Metamorphism of carbonatites must have the convergent features with metamorphozed sedimentary apatite-bearing carbonate rocks. This is concerns only carbonatites matrix. High consentrations of P, Sr, Ba, Nb, TR must be due the appearance of new minerals in the carbonatites.

Metamorphosed Veseloe carbonatites are very interesting object of investigation. The age of the carbonatites has been determined as  $593\pm3.5$  Ma using the 207Pb/235U ratio from zircon. Rb-Sr dating of the metamorphism using phengite gave  $550\pm14$  Ma (MSWD – 1.2).

In the Veseloe carbonatites there must be clearly seen transformations of primary minerals, the appearance of new minerals that are not typical of carbonatites.

### ANALYTICAL METHODS

Whole rock chemistry was determined by atomic absorption, with trace element determinations by XRF and REE by ICP-AES with preliminary chemical concentration. For REEs certified standard apatite ore (AR) and apatite (AK) were used as quality control standards. CO2 was determined by titrameteric chemical decomposition. Strontium isotope values in non-rubidium minerals were carried out using an MI-1201 T mass-spectrometer. A Rb/Sr isochron was plotted using data from mica, K feldspar and whole rock. All these methods were carried out in the laboratories of the Geological Institute SB RAS (Ulan-Ude, Russia). The minerals were analysed using electron probe micro-analysis (EPMA). Most analyses were carried out at the Geological Institute SD RAS using a MAR-3 WDS (SV Kanakin) microprobe with an accelerating voltage of 20 kV, beam current of 40 nA, beam size of 3-4 µm and a 20 s counting time and a LEO-1430 scanning electron microscope with an IncaEnergy-300 energy-dispersive system (SEM-EDS) (NS Karmanov) 4 operated at 20 kV and 0.5 nA. Additional imaging and analyses were carried out at the Natural History Museum, London. Two instruments were used: a Jeol 5900LV SEM equipped with an Oxford INCA EDS system and operated at 15 kV and 1 nA specimen current measured on a Co standard, calibrated with various natural and synthetic materials and using a ZAF matrix correction and a Cameca SX50 WDS electron microprobe operated at 15 kV and 20 nA (carbonates) and 20 kV and 20 nA (oxides, silicates, phosphates). The  $\delta 13C$  and  $\delta 18O$  isotope values were analyzed in the isotope laboratory of the Analytical Center FESC RAS (Vladivostok, Russia) using a Finnigan MAT 252 sensitive mass-spectrometer. The analytical errors were not more than  $\pm 0.05$  for carbonates and  $\pm 0.5\%$  for magnetite. A concordia age was determined using the 207Pb/235U ratio in zircon by analysis on the 5- collector SHRIMP-II massspectrometer in VSEGEI (St. Petersburg, Russia).

# GEOLOGICAL DESCRIPTION OF THE VESELOE AREA

Veseloe carbonatites occur within the Baikal-Muya belt of Riphean islandarc area which is in southern part of Siberian craton. Rift zone was formed in Late Riphean-Vendian in the eastern part of this belt on the territory of Kelyano-Irokindinskaya area; this area was accompanied by typical complexes of volcanicsedimentary rocks and basic magmatism [1] (Fig. 1).

Veseloe area is located in the western part of North-Muya block. The block represents the part of Early Precambrian crystalline basement that consists of slates, gneiss, granite-gnaiss. The rock of basement was broken through Prevendian gabbroids and granites. Block presents Early Precambrian crystalline basemen consist from of slates, gneisses, granite-gneisses break through do vend gabbroid and granites. All rocks, including carbonatite as well as slates, gneisses, granite-gneisses and marbles were metamorphosed in greenschist conditions with a temperature up to 450°-470°C and a pressure from 6 to 8 Kbar (pressure was

calculated according to Massone et al. (1989) using mice). Rock of this area were metamorphosed at amphibolite, epidote-amphibolite and some times eclogitic conditions.

Crystalline slates occur on the west from block. The rock are regarded by some researchers to Late Precembrian granite-gneisses to processed crystalline basement [1]. Some geologists (S.V. Okorokov et al.,) repoted that rocks were formed metasomatic processes. To the west they are changed by volcanicsedimentary rocks of Muya series of Riphean age. Rocks of Muya series are represented by of basic and volcanic rocks that were metamorphosed it greenschist conditions. Built by chlorite, actinolite, epidote, albite and clear micas consist the rocks [1]. The composition of crystalline slates on the Veseloe area are micaceousquartz-feldspar. They are enriched by chlorite, actinolite, epidote, lenses of dolomite marbles.

Metamorphic rocks are broken through by bodies and dykes of gabbroes Riphean, ultrabaisites, Palaozoic granitoids. In Veseloe area the basic and ultrabasic rocks consist of the chains of bodies and dykes along tectonic zones to north-north-west (fig. 1).





1- drift, 2 – Cambrian limestone, 5 - Riphean dolomite marble, 3 – Late Riphean crystalline slates, 4 – Riphean crystalline mica-quartz-feldspar slates, 6 – Palaeozoic alkaline and biotite granites, 7 – Late Riphean gabbro: a) – bodies, b) –dykes, 8 – altered ultrabasic rocks (talc-carbonate rocks and listvenites), 9 – bodies of carbonatites, 10 – faults.

### GELOGICAL-PETROGRAPHIC DESCRIPTION OF THE VESELOE AREA

Micaceous-quartz-feldspar crystalline slates consist of the Veseloe area (fig. 1). These rocks are fine-grained, with banded textures, with thin bands of mica. In these rocks the zone of schistosity are widespread.

The feldspars are microcline, albite and oligoclase. They are sericitezed. Potash-feldspar contains 1.3-1.7% of BaO.

Layers of clear micas are orientated according to banded of rocks. Mica is phengite, it contains 2-7 wt.% of FeO and more than 1,5 wt.% of MgO (Table 1). Amount of Fe and Mg are up to 0,5 f. c. Minerals contain the barium (0,21-0,49 wt.%), fluorine (0,31-0,55 wt.%). Phengite has high content of silica (3,25-3,45  $\phi$ . e.), what indicate about high pressure at formation of mineral.

Table 1.

N⁰	SiO <sub>2</sub>	TiO <sub>2</sub>	AI <sub>2</sub> O <sub>3</sub>	FeO	MnO	MgO	K <sub>2</sub> O	BaO	F	Total
1	48,89	0,55	24,56	7,51	0,15	2,35	11,37	0,25	0,55	100,78
2	49,17	0,5	24,05	7,17	0,24	2,57	11,23	0,21	0,54	100,33
3	48,29	0,2	24,23	7,65	0,26	2,73	11,39	0,29	0,55	99,98
4	47,12	0,44	30,53	3,97	<0,06	1,25	11,56	0,38	0,28	100,13
5	47,54	0,58	29,64	4,32	0,07	1,57	11,36	0,49	0,37	100,58
6	47,67	0,52	29,44	4,29	0,09	1,57	11,23	0,48	0,31	100,23
7	49,25	0,33	27,52	3,18	0,07	3,13	11,44	0,31	0,55	100,55
8	49,82	0,06	28,52	2,39	0,09	2,86	11,66	0,35	0,33	100,56
9	48,95	0,39	27,86	3,03	0,08	2,93	11,32	0,37	0,52	100,15

Chemical composition of micas in rocks from slates the Veseloe area, wt.%

Note. Total result is given with theoretical water and accounting CaO, BaO,  $Na_2O$  contents considering which below detection limit of about wt.%.

Epidote consits isometric grain. Sometimes the mineral contains of the single inclusions of allanite. Large crystals of allanite (up to 2-3 cm) are present in potash feldspars veinlets of slates.

Biotite associated with epidote, quartz. They are orientated according to banded of rocks.

Magnetite is presented from ore minerals. It is also orientated according to banded of rocks.

Chemical composition of slates and admixture elements in them indicate that these slates to be formed as of metamorphism of acid rocks. On the composition the rocks (Na<sub>2</sub>O+K<sub>2</sub>O) are regarded to subalkaline series. Amount he silica varies from 69 - 76 wt.% of SiO<sub>2</sub>, 12-15,15 wt.% of Al<sub>2</sub>O<sub>3</sub>, of iron 0,25-2,39 wt.% of Fe<sub>2</sub>O<sub>3</sub> and 0,11-1,03. wt % of FeO. Amount of titanium is from 0.02 to 0.36 wt.% of TiO<sub>2</sub>. Amount of sodium, potassium and calcium vary from 3,68 - 6,57 wt.% of Na<sub>2</sub>O, 2,16 - 3,88 wt.% of K<sub>2</sub>O, 0,46-1,48 wt.% of CaO.

Dolomite bodes are presented in slates (fig. 1), and they were modified at metamorphism in marbles. They are lenticular bodies with thickness 10 metres.

Marbles contain the dissemination of phengite and veinlets of quartz. Phengite contains 0,16 wt.% of FeO and 5.53 wt.% of MgO (Table 2). Amount of silica is 3,41 f.c. in the mineral. In the dolomite marbles the carbon and oxygen isotopic compositions are similar to limestones (an average  $\delta^{18}O=13,7-16,1\%$ ,  $\delta^{13}C=+2,8-+3,8\%$ ). Table 2.

N⁰	SiO <sub>2</sub>	TiO <sub>2</sub>	AI <sub>2</sub> O <sub>3</sub>	FeO	MgO	K <sub>2</sub> O	Total
1	51,80	1,31	26,32	0,08	5,00	10,69	101,25
2	49,32	1,05	30,32	0,16	3,49	10,42	100,90
3	51,01	0,28	26,49	0,16	5,53	11,25	100,29

Chemical composition of phengite from dolomte marble, Veseloe area, wt.%.

The gabbroes and ultrabasic rock are presented among magmatic rocks [7]. They also undergo of metamorphism.

Ultrabasic rocks were metamorphosed with formation of quartz-talc-carbonate rocks. The rocks consists the lenticular bodies which thickness is up to several metres. Carbonate minerals are predominant in these rocks. Magnesite, dolomite, chlorite, talc, phengite, quartz, actinolite, phlogopite, chromite also consist these rocks.

Magnesite, dolomite and quartz are the main rocks-formation minerals. Dolomite contains less than 4 wt.% of FeO (Table. 3). Magnesite has amount of FeO (from 6,72 to 13,12 wt.% of FeO), it is similar to breunerite.

Table 3.

Chemical composition of carbonate minerals from dolomites marbles and talc-carbonates rocks the Veseloe area, wt.%.

N⁰	FeO	MnO	MgO	CaO	Total
1	<0,07	<0,06	21,69	29,73	98,45
2	0,22	<0,06	22,08	29,91	99,2
3	0,07	<0,06	22,33	30,19	99,59
4	0,09	<0,06	21,99	30,33	99,41
5	0,08	<0,06	21,79	29,79	98,67
6	7,09	0,24	42,24	0,04	102,00
7	6.72	0.09	42.56	0.04	99.44
8	8.94	0.18	40.33	0.56	100.10
9	13.12	0.36	38.72	0.05	99.67
10	7.55	0.14	42.12	0.06	100.17
11	2,50	0,25	19,79	29,12	99,11
12	3,98	0,17	18,26	29,88	100,34
13	3,55	0,43	18,33	30,24	100,02
14	2,20	0,46	21,10	28,76	99,85

**Note.** Samples 1-5 dolomite from marbles, 6-10 magnesite-breunerite and 11-14 dolomite from talc-carbonate rocks. Strontium was not identified. The total is given with account theoretical  $CO_2$ ,  $SiO_2$ ,  $TiO_2$ ,  $Cr_2O_3$ ,  $Na_2O$ ,  $AI_2O_3$ ,  $K_2O$  and BaO, Their amounts is not higher than ten fraction of the percent.

The oxygen and carbon isotopic compositions from quartz-talc-carbonate rocks are showing in table 4.

Table 4	
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Oxygen, carbon and strontium isotopic compositions in minerals from talc-carbonate
rocks of the Veseloe area

Nº	Samples	Mineral	δ <sup>18</sup> O‰SMOW	$\delta^{13}$ C PDB	<sup>87</sup> Sr/ <sup>86</sup> Sr
1	Sr-7	magnesite+	12,0	-1,9	
2	Sr-8	dolomite	12,2	-0,9	
3	4/04		11,7	-1,0	
4	8/04		10,4	-1,4	
5	EB-11	-//-	11,5	-1,6	
6	EB-12	-//-	14,2	-1,4	
7	131/05		14,1	-0,6	0.70426
8	22/05	-//-	12,6	-0,8	0.70527
9	19/05		13,3	-1,6	0.70583

**Note.** The analysis was carried out in Analytical Center of Far East Scientifie Centre RAS (Vladivostok, Russia).

Table 5.

#### Chemical composition of minerals from modified ultrabasic rocks, Veseloe area, wt.%.

N⁰	SiO <sub>2</sub>	TiO <sub>2</sub>	AI <sub>2</sub> O <sub>3</sub>	Cr <sub>2</sub> O <sub>3</sub>	CaO	FeO	MgO	Na <sub>2</sub> O	K <sub>2</sub> O	NiO	F	Total
1	30,76	0,05	17,40	0.52	-	7,41	30,30	-	-	-	0,64	99,18
2	31,19	0,05	17,83	0.52	-	7,78	29,36	-	-	-	0,45	99,24
3	29,36	0,05	16,23	0,53	-	7,22	28,07	-	-	-	0,78	94,58
4	30,64	0,05	17,63	0,93	-	6,83	29,14	-	-	-	0,65	97,95
5	32,38	0,05	16,44	0,56	-	7,62	29,78	-	-	-	0,74	100,19
6	31,23	0,05	15,16	2,11	-	6,50	30,18	-	-	-	0,75	86,09
7	48,41	0,11	25,72	4,68	-	0,85	3,01	-	10,70	-	-	98,93
8	50,20	0,09	24,13	3,76	-	0,83	5,03	-	10,74	-	-	100,44
9	47,89	0,10	25,39	5,90	-	0,89	3,14	-	10,74	-	-	99,18
10	50,49	0,12	22,59	3,28	-	2,59	4,66	-	10,82	-	-	100,51
11	48,36	0,06	26,67	4,22	-	0,92	3,00	-	10,81	-	-	99,32
12	47,77	0,10	24,51	5,79	-	1,09	3,45	-	10,79	-	-	99,10
13	47,91	0,11	25,27	5,71	-	1,01	3,16	-	10,50	-	-	98,41
14	56,54	-	0,81	-	10.95	7.11	19.85	1.25	-	-	-	98.9
15	56,07	-	1.09	-	11.36	7.31	19,99	1.24	-	-	-	97.43
16	54,82	-	0,93	1,17	11,03	5,61	21,15	1,09	-	-	-	98,80
17	43,10	-	10.47	0.94	-	2.75	26.76	-	10.84	1.03	5.35	101.59
18	42.77	-	10.86	0.77	-	3.13	26.68	-	11.51	0.95	5.02	101.51
19	42.49	-	10.77	0.84	-	3.12	26.57	-	10.99	1.24	5.25	101.27
20	42.59	-	11.41	1.38	-	3.23	26.29	-	10.90	0.48	5.01	101.29
21	43.24	-	11.11	1.55	-	3.48	26.86	-	11.09	-	5.08	102.41
22	42.70	-	10.96	4.75	-	4.20	24.26	-	10.10	0.88	4.53	102.38
23	43.14	-	10.99	1.98	-	3.05	24.98	-	10.44	1.01	5.81	101.41

Note. Sample: 1-6 chlorite, 7-13 phengite, 14-16 actinolite. 17-23 phlogopite. Total with theoretical water.

Chlorite is represented by leaf-like fibrous aggregates and occurs between grains of carbonate. Chlorites have high content of magnesium (from 28,07 to 31,23

wt.% of MgO) (table 5). Amount of aluminium is up to 17,83 wt.%  $AI_2O_3$ . Content of iron up to 8wt.% of FeO. The mineral contains up to 2,11 wt.%  $Cr_2O_3$ .

Talc (5-10%) forms colorless fine flakes with dolomite and magnesite. Mineral is associated with phengite.

Bright-green mica (1-2%) is chromium-bearing phengite. Mineral is growing together with quartz. Amount of MgO varies from 3,01 to 5,03 wt.% and FeO up to 2,59 wt.% (Table 5). As well as in others rocks the magnesium and iron concentration is 0,5 f. c. Amount of silica is 3,39-3,44 f.c. In phengite there is from 3,28 to 5,90 wt.% of  $Cr_2O_3$ .

Chromite is typical mineral and is represented as dispersed dissemination. Grains of chromite have isometric form. Mineral often is replaced by chromebearing phengite and chlorite.

In the talc-carbonate rocks presents as well phlogopite. It included together with quartz an chromite grains and it also occurs in talc-carbonate matrix. Mineral is similar to fluoroflogopite with high contents of chrome and nickel (Table 5).

Quartz (6-7%) forms aggregate accumulations, grains and veinlets. It is irregularly distributed in the rock.

Massifs of the basic rocks occur in the west and east part of the Veseloe area. This is hornblende gabbro transferring into the gabbro-diorite.

Gabbro dykes are presented on the Veseloe area. They are accumulated in the stripe of north-west direction with 5-6 km thickness (fig. 1). Thickness of dykes is up to several meters. Their contacts with host schists are sharp. These are massive fine-grained rocks of deep-green colour. In result of metamorphic processes the gabbroes were transformed in the aggregate of epidote, chlorite, amphibole, biotite, with few amount of plagioclase, titanite. There are veinlets and small assemblages of mica.

Chiefly, bulk of rocks are composed by epidote. It forms isometric grains. Large grains of epidote have sharp border and closely associate with chlorite.

Table 6

N₂	SiO <sub>2</sub>	Al <sub>2</sub> O <sub>3</sub>	FeO	MgO	CaO	Na <sub>2</sub> O	Ce <sub>2</sub> O <sub>3</sub>	La <sub>2</sub> O <sub>3</sub>	Pr <sub>2</sub> O <sub>3</sub>	Nd <sub>2</sub> O <sub>3</sub>	Total
1	38,04	27,09	9,11	0,10	22,98	-	0,16	0,16	0,27	0,28	99,41
2	37,77	26,33	10,37	0,05	23,33	-	0,16	0,16	0,27	0,28	100,01
3	38,41	27,39	9,27	0,06	23,12	-	0,16	0,16	0,26	0,28	100,38
4	37,39	24,01	13,24	0,03	23,14	-	0,16	0,16	0,27	0,28	100,12
5	37,26	27,21	9,56	0,6	23,25	-	0,16	0,16	0,26	0,28	99,35
6	67,82	19,89	0,14	-	0,09	11,4	-	-	-	-	99,34
7	67,33	19,82	0,33	-	0,06	11,4	-	-	-	-	99,01
8	67,75	19,97	0,17	-	0,05	11,4	-	-	-	-	99,91
9	67,32	20,33	0,14	_	0,46	11,2	_	_	_	_	99,61
10	48,70	7,66	13,35	14,22	11,25	-	-	_	-	-	100,01

Chemical composition of minerals from gabbro of the Veseloe area, wt.%.

Note. 1-5 epidote, 6-9 albite, 10 amphibole. Total with theoretical water.
Albite forms the isometric grains together with chlorite. By parthly albite is corroded by biotite. Amount of  $A1_2O_3$  is from 19,82 to 20,33 wt.%, amount Na<sub>2</sub>O is within the of limits 11,15-11,41 wt.% (Table 6).

Amphibole is presents a isometric grains, prisms and needles of bright green colour. It composition conforms tremolite with contents of CaO 11.25 wt.% (Table 6).

#### CARBONATITES

Carbonatite dykes and drift is occur on the area 1 km wide and about 4-5 km long. Carbonatites was traced on distance 600-1000 metres along of broadsides stream Veseloe and right-bank of stream Srednii. Thickness of carbonatites bodies are several metres.

The carbonatites are fine-grained rocks. They have banded textures, which conform to the country rock contacts. The texture is caused by common orientation of apatite, dolomite, magnetite, rutile and alkaline amphibole, as well as by the presence of zones and lenses rich in apatite, dolomite and calcite.

The contacts with host slates are tectonised and some of the slate has been pulled off into in carbonate matrix. Occasionally, thin phlogopite zones are preserved at the contacts and these are the only evidence of any fenitisation processes at Veseloe. The phlogopite is usually deformed and partially replaced by chlorite.

The Veseloe rocks are magnesiocarbonatites [6]. A distinct geochemical feature rocks is the high concentrations of P (up to 9.17 wt.%,  $P_2O_5$ ), Ti (up to 0.54 wt.% TiO<sub>2</sub>) and F (0,32-0,76 wt.%). Comparison with average magnesiocarbonatite [9] emphasized the higher Cr, Ni (at 3-4 time) at Veseloe and lower Nb (4-17 ppm), Zr (47-270 ppm), MnO (0,11-0,14 wt.%).

The carbonatites are LREE-enriched the chondrite - normalized plots are similar to other carbonatites worldwide.

Table 7

№ sample	Mineral	$\delta^{18}$ O‰ SMOW	$\delta^{13}$ C PDB	<sup>87</sup> Sr/ <sup>86</sup> Sr						
13/04	dolomite	10,7	-2,3	0.7037						
19б/04		11,1	-2,1	0.7038						
25a/04		10,1	-2,0							
8/04		10,4	-1,4							
72		8,9	-2,2							
8/03		10,1	-1,9							
8/03		10,2	-1,9							
124/03	calcite	9,8	-2,3							
8a/04		1,5								
196/04	magnetite	3,9								
215/04		4,0								
Bec-5	apatite	5,3								

 $\delta^{18}$ O‰ and  $\delta^{13}$ C values and initial  ${}^{87}$ Sr/ ${}^{86}$ Sr rations from the Veseloe carbonatites.

Oxygen isotopic composition in the carbonate minerals is similar to primary igneous carbonatites (Table 7).  $\delta^{18}$ O values in the magnetite and apatite are similar to mantle. Values initial  ${}^{87}$ Sr/ ${}^{86}$ Sr ration in dolomite from carbonatites also are similar to mantle – derived rocks.

The carbonatites consist of dolomite (75-80%), apatite (15-20%) and calcite (5-15%). Minor minerals consist of magnetite (1-2%), phlogopite (1%), and alkaline amphiboles. There are 3 associations that were formed at magmatic carbonatite process, metamorphic and hydrothermal stages. Mineral composition is given in the table 8.

Table 8.

Associations	Mineral composition				
Carbonatite	Main: dolomite, apatite, calcite Minor and accessory: phlogopite, magnetite, rutile, magnoriebeckite, richterite, zircon, molybdenite, monazite				
Metamorphozed	Phengite, quartz, talc, tremolite, actinolite, chlorite, anthophyllite, dolomite, calcite, barite, baritocelestine, monazite, allanite, strontianite, synchysite				
Hydrothermal not-related with carbonatites	Chlorite, quartz, pyrite, chalcopyrite, halenit, blende				

Mineral composition of carbonatites Veseloe area.

# FEATURE OF METAMORPHIC PROCESSES IN THE CARBONATITES.

The carbonatites, crystalline slates, basic and ultrabasic rocks, dolomite marbles have signs of metamorphic processes. There are recrystallisation, changed structure and texture of rocks, neoformation of minerals, and rafination of primary minerals from admixtures. Metamorphism of the Veseloe carbonatites was caused the partial recrystallization of carbonate matrix, rafination of apatite and dolomite from REEs and Sr. Recrystallization was accompaned by formation of secondary calcite and dolomite veinlets. The examples of these processes are given in figures 2, 3. Assembly of calcite, quartz, monazite, rutile, and ilmenite is present an the figure 2. Carbonate veinlet with allanite, monazite, barite-celestine is given in figure 3.

Introduce of Si, Al, Na and K an carbonatite from country rocks and interactive with calcite were caused the formation of talc, amphibole, phengite, chlorite, quartz, allanite. These minerals distrohited irregularity. It was caused by different degree of rock failure. Amount of these minerals are not more than 3%. In some causes up to 5-7 persent. Talc consists of the rare veinlets (with size up to 1-3 millimeter).

Phengite and amphibole form the disseminations between dolomite grains. Occuring of phengite in carbonatite is given in figure 4.



Fig. 2. Aggregate of quartz (9,10,12), calcite (7,8,11), monazite (2,3), rutile (1,4,5,6), titanite (13), which were formed at metamorphic processes.



Fig. 3. Veinlet of calcite (8) with allanite (2,4), monazite (5), barite-celestine (3), titanite consists of the main area (1). Magnetite and ilmenite are points (6 and 7). All these minerals were formed at metamorphic processes.

Phengite is most interesting from neogenic minerals. It consists of the mica flake and plates with size up to 0,2-1,5 millimeter (fig. 4). Composition of mica is similar to phengite from host slates. Amount of Mg and Fe in the mineral is 0,5 f. c. Phengite contains up to 1,72 wt.%  $Cr_2O_3$  (table. 9) in the Cr-rich parts of carbonatites (this is assemblage chrome-bearing of rutile, magnetite with chromite).



Fig. 4. Phengite (Phn) in the dolomite (Dol). White-rutile.

Table 9.

N⁰	SiO <sub>2</sub>	TiO <sub>2</sub>	AI <sub>2</sub> O <sub>3</sub>	Cr <sub>2</sub> O <sub>3</sub>	FeO	MgO	K <sub>2</sub> O	Total
1	46,43	0,27	30,55	1,12	2,31	1,87	11,95	99,21
2	47,29	0,42	29,64	1,13	2,39	1,79	11,54	98,94
3	49,25	0,34	27,57	-	3,18	3,13	11,45	100,55
4	49,82	0,06	28,52	-	2,4	2,86	11,67	100,56
5	48,95	0,39	27,86	-	3,04	2,93	11,31	100,15
6	48,58	0,41	28,77	-	3,22	2,75	11,36	100,8
7	48.71	0.79	26.44	1.72	3.72	2.56	11.96	95.90

Chemical composition of phengite from Veseloe carbonatite, wt.%.

Note. Total include the MnO,  $Na_2O$  and theoretical water.

The metamorphic amphibolies presented magnesia (anthophyllite) and calcium-magnesian (tremolite-actinolite) types (table 10, fig. 5), whereas the alkaline amphiboles (riebeckite, richterite) usually are in carbonatites.

Deep-seated magmatism, its sources and plumes

Table 10.

Nº	SiO <sub>2</sub>	TiO <sub>2</sub>	Al <sub>2</sub> O <sub>3</sub>	FeO	MnO	MgO	CaO	Na <sub>2</sub> O	K <sub>2</sub> O	Total
1	58,89	<0,05	0,09	4,03	0,07	21,92	12,22	0,46	0,13	100,81
2	59,18	<0,05	0,26	3,19	<0,06	22,62	12,36	0,50	0,95	100,63
3	59.00	<0,05	0,28	4,85	0,09	21,33	12,17	0,54	0,11	100,87
4	59,07	<0,05	0,24	4,06	<0,05	22,14	12,29	0,79	0,17	100,70
5	58,63	<0,04	0,29	4,74	0,08	21,66	12,07	0,62	0,14	100,47
6	58,83	<0,04	0,16	3,34	0,07	22,35	12,45	0,46	0,13	100,29
7	58,67	<0,05	0,36	4,51	0,06	21,82	12.00	0,70	0,15	100,64
8	58,67	<0,05	0,15	4,46	<0,09	21,88	12,33	0,43	0,09	100,34
9	62,79	< 0,04	0,08	5,40	<0,06	27,49	0,07	0,08	<0,03	98,50

Chemical composition of metamorphic minerals at the Veseloe area.wt.%.

**Note.** Samples: 1-8 tremolite-actinolite; 9 antophyllite, Total with  $Cr_2O_3$ , BaO, NiO,  $V_2O_3$ , their amounts are not more than fraction of the percent. Fluorine is not detected.



Fig. 5. Diagnostic diagram of metamorphic amphibolies from Veseloe carbonatites.

Talc was replaced of carbonatites and formation of qaurtz-carbonate veinlets were at the metamorphism. Talc replaced amphibole and dolomite. Sometimes it form the thin (up to 1-3 millimeters) veinlents. Talc usually occurs in silicification parts of carbonatites. Talc has high content of iron (6 wt.% of FeO).

Allanite also is in carbonatites (fig. 6). It compositions is in the table 11. Mineral have 13,16 wt.%  $Al_2O_3$ , 29,89-32,31 wt.%  $SiO_2$ , and 11,22 wt.% CaO. Allanite is enriched the light rare earth elements.

Table 11.

Chemical composition of allanite from Veseloe carbonatites wt.%.

N⁰	$Al_2O_3$	SiO <sub>2</sub>	CaO	TiO <sub>2</sub>	FeO	$La_2O_3$	$Ce_2O_3$	$Pr_2O_3$	$Nd_2O_3$	Total
1	11.62	30.94	10.58	1.10	17.25	4.44	12.54	1.83	5.73	96.03
2	13.16	31.31	11.22	0.53	17.59	6.38	11.01	0.86	3.55	95.61
3	10.91	32.21	10.67	-	15.53	6.79	11.39	-	2.01	92.81
4	10.89	29.89	9.99	-	15.48	6.73	11.71	0.74	2.34	90.99

Note. 3 and 4 samples have 1,22 wt. % BaO and 2,10 wt. % MgO.



Fig. 6.. Allanite (Alt) fill up the dolomites intergrowth space (Dol). Thin section 17/04. Nicol +. Increase ×70.

Rafination of minerals from admixtures (from REEs, Sr and Ba) was at metamorphic processes. Apatite and dolomite were the mains sources of these elements. The result of rafination was formation of minerals: monazite, allanite, strontianite, barite, celestine. These minerals formed microveinlets and disseminations. Small grains and microveinlets of monazite are usual for apatite (fig. 7).



Fig. 7. Monazite (Mnz) dissemination (white) in the apatite (Ap), dark-grey is dolomite (Dol).

Monazite is enriched of light rare earth elements (table 12). Amount of Ce is stabile, amount of La and Nd is varied. High amount of Nd is characterized for monazite inclusions from apatite. In addition the mineral contains strontium (from 0,5 to 4,23 wt.% SrO), thorium and admixture of calcium (1,8 wt. % CaO).

Table 12.

N⁰	P <sub>2</sub> O <sub>5</sub>	CaO	SO <sub>3</sub>	SrO	La <sub>2</sub> O <sub>3</sub>	Ce <sub>2</sub> O <sub>3</sub>	Pr <sub>2</sub> O <sub>3</sub>	Nd <sub>2</sub> O <sub>3</sub>	Sm <sub>2</sub> O <sub>3</sub>	ThO <sub>2</sub>	Total
1	30,34	1,08	0,46	-	16,48	33,81	3,88	12,63	-	-	98,68
2	30,87	1,37	-	0,82	14,74	31,43	4,14	15,66	1,56	0,9	101,49
3	32,18	0,69	n.d.	0,61	19,27	35,22	3,75	9,28	-	0,66	101,66
4	29,34	0,94	n.d.	0,76	15,44	32,12	3,88	15,02	-	-	97,50

Chemical composition of monazite from Veseloe carbonatites, wt.%.

Figure 8 show that dark parts within of apatite grains are poor of Sr and REEs.

Fig. 8. Rafination apatite from Sr and REE (dark sites), the main area is dolomite.

Strontianite consists of the dissemination of radial aggregates in the dolomite matrix. It size is not more than ten fraction of millimeter.

### DISCUSSION

Concerning the problem of metamorphism, recrystallisation of rocks is mainly mentioned in recent publications. The carbonatites of the Gulinsky deposit (Russia) which have relict sites with primary structures and textures are given as an example [10]. Similarly, there is a recrystallisation of carbonatites up to coarsegrained rocks in British Columbia (Carbonatites, 1989). An easy recristallisation of calcite is mentioned in the book [4]. As for other processes of carbonatite changes accompanying metamorphis, there is lack of information about it in literatures.

100µm

The Veseloe carbonatite of are of interest as they experienced metamorphic transformations like host crystalline slates, ultrabasites, dykes of gabbroids and bodies of dolomites. Besides the recrystallitation of rocks caused marked changes in the mineral composition. From one hand, the appearance of minerals not typical for carbonatites, but inherent to metamorphic processes is due to the changes in the mineral composition. Such minerals as talc, phengite, chlorite, tremolite-actinolite, ortite refer to them. Another consequence of metamorphism is rafinetion of minerals from ad mixtures with the form of new stable mineral phases. The main minerals among them are monazites, strontianites, barites, celestines.

Mineral parageneies of host rocks correspond to the faces of green slates (epidot-muscovite-chlorite subfaces). The present association of phengite, chlorite, epidote, tremolite, actinolite, albite witnesses about it. The paragenesis of metamorphic minerals in the carbonatites on the whole is similar to the paragenesis of host rocks. The composition of mica, which is a typomorphic mineral of carbonatites, metamorphosed ultrabasites, marbles, is on the whole similar to the composition of phengite of the host slates. Their important feature is the high contents of silica in the phengite in the tetrahedric site (3.25-3.45 f.c.) in all the cases, which show the high pressure conditions of rock and mineral formation. It the temperature of the formation of mineral association is to 400°-430°C the value of pressure must be 6-8 kbar (fig. 9) [3].



Fig. 9. The evaluation of metamorphic pressure of the rock by a phengite barometer (Massone et. al, 1989) in the Veseloe site.

The diagram of amphibole composition dependence versus their genetic nature [2] (fig. 10) shows that the amphiboles of metamorphic origin of the Veseloe site are clearly separated from the amphiboles of carbonatites and are located outside the area typical for these rocks.



Fig. 10. The diagram of amphibole composition from the Veseloe carbonatites with carbonatites of Ca-Mg-Fe contents.

1-ribecite, 2-richterite, 3-tremolite-actinolite. G – the area of amphiboles associated with carbonatites. The arrows show the evolution trends of chemical changes of amphiboles in granites and gneisses (A), amphibolites (B), and rocks Mg-rich (C).

The introduction of potassium, silica and aluminum from the host slates into carbonatites led to the replacement of some calcium and magnesium minerals (dolomite, calcite, alkaline amphiboles) by new minerals such as tremoliteactinolite, talc, phengite, chlorite.

One more consequence of the process of metamorphism is the process of rafination of the apatite, dolomite and calcite from admixtures. Originally they contained higher concentration of strontium and the apatite rare earth elements as well. As a result, the concentration of the mentioned above elements in the minerals sharply decreased, and newly formed monazite, stronzianite, sinchisite and allanite appeared round the edges of grains and cracks of minerals. The age of the monazite released from the apatites of the Pogranichnoe occurrence [5] coincided with the one of the Veseloe site located in north Transbaikalia and happened to be  $550\pm17$  mln years old.

It is likely that metamorphism had a certain effect on the oxygen and carbon isotope compositions in the minerals. The heterogeneity of carbon and oxygen values in carbonate minerals, which are easily subject to recrystallization is probably due to metamorphism. The enrichment of the dolomite and calcite by the heavy  $\delta^{13}$ C value in the carbonatites of the site is supposed to be connected with metamorphism as well. The heavy  $\delta^{13}$ C value is known to be also typical for the Seligdarsky deposit (Russia) the rocks of which underwent deep metamorphism. Such minerals as talc, tremolite and chloritized sites witness about it. It should be

pointed out, that increasing the carbon isotope composition in the carbonate rocks transformed in marbles is a rather spread phenomenon. From the other hand, magnetite and apatite are more stable to the processes of recrystallization which preserved the typical oxygen composition for carbonatites (5.3% in the apatite and 4.0% in magnetite).

## CONCLUSION

All research showed that in the carbonatites, being metamorphically changed, some alterations of structural textural peculiarities, of mineral composition, mineral recrystallization processes accompanying the admixture elements reduction and isotopic composition changes can exist. So it confirms the necessity of more careful geochemical and more detailed isotopic research. That will contribute to opening of discovering/finding some non typical and not always understandable processes of mineral formation in these rocks.

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