

# Geodynamic Interpretation of the Subsolidus Recrystallization of Mantle Spinel Peridotites: 1. Mid-Ocean Ridges

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**Abstract**—A kinetic diffusion model was developed for the estimation of peridotite cooling rates from the closure temperatures of exchange reactions (Ca–Mg between two pyroxenes and Fe–Mg between olivine and spinel). It was found that the peridotites of mid-ocean ridges (MOR) are characterized by a limited range of the closure temperatures of the two-pyroxene ( $995 \pm 34^\circ\text{C}$ ) and olivine–spinel ( $793 \pm 51^\circ\text{C}$ ) exchange reactions suggesting rapid (about  $0.01^\circ\text{C}/\text{year}$ ) continuous cooling of the rocks after melt segregation. A two-layer structure is advocated for the lithosphere beneath the axial parts of MOR (with respect to the prevailing mechanism of heat transfer). The upper layer is permeable for hydrothermal fluids. Its thickness is estimated from the maximum temperature of hydrothermal metamorphism from 9 to 21 km at an average value of 15 km. The thickness of the lower, impenetrable layer is estimated from the closure temperatures of exchange reactions. It varies from 0.8 to 19 km averaging at 4.4 km. The total thickness of the lithosphere in the axial MOR zone in the areas studied ranges from 11–33 km at an average value of 18.4 km. The pressure range of the separation of final melt fractions from peridotite is estimated as 3.9–10.6 kbar. A direct correlation was observed in normal MOR segments between the spreading rate and the degree of depletion of mantle peridotites, whereas the thickness of the lithosphere correlated negatively with these parameters. The peridotites under investigation do not show evidence of repeated heating or any isothermal subsolidus stage. This suggests the absence of large magma chambers beneath the axial parts of MOR within the areas investigated.

## INTRODUCTION

Mantle spinel peridotites are exposed in various geodynamic settings including intraplate areas (xenoliths in kimberlites and alkali basalts), constructive plate boundaries (mid-ocean ridges), and modern ocean–continent transition zones (ophiolitic peridotites in a broad sense of the term). Now, it is universally accepted that in a certain period of their history, these rocks were in equilibrium with mantle melts segregating from peridotites in various geodynamic environments. However, the compositions of primary peridotite minerals change usually in the course of rock recrystallization after melt separation.

A number of geothermometers were proposed on the basis of experimental calibration of temperature-dependent partition coefficients of various elements or their pairs between coexisting minerals. Their application to mantle peridotites demonstrates that temperatures estimated for various pairs of coexisting minerals from a single sample or for various element pairs from a single mineral pair are only rarely consistent with each other within the accuracy of the method.

This allows us to suggest that, in contrast to regional metamorphic rocks, subsolidus “metamorphic temperatures” estimated by mineral geothermometers in the majority of mantle spinel peridotites do not characterize a particular isothermal metamorphic event but are in fact closure temperatures of the respective

exchange reactions between minerals (Dodson, 1973). The calculated temperatures depend on the character of an exchange reaction, cooling rate, and grain size of the minerals. Thus, it is reasonable to consider the closure temperature of a particular exchange reaction for a given grain size of minerals participating in the reaction (Dodson, 1973). The subsolidus recrystallization of oceanic spinel peridotites occurred under decreasing temperature, which is suggested by systematic differences in the compositions of rims and cores of mineral grains and temperature estimates obtained by Ca–Mg two-pyroxene and Fe–Mg olivine–spinel geothermometers (Arai and Fujii, 1978; Fujii, 1990; etc.). Thus, the absolute temperatures obtained from the two exchange reactions can be used to estimate cooling rates and characterize mantle diapirs and the geodynamic settings of magmatism and metamorphism (Bonatti and Michael, 1987; Parkinson and Pearce, 1998; Bazylev and Kamenetsky, 1998).

This paper is an attempt at using this idea for the quantification of the rates of exchange reactions in peridotites and modeling possible character of changes in the temperature of mantle rocks during their cooling in mid-ocean ridges. The main goal was to determine the petrologic and physical parameters that controlled the closure temperatures of exchange reactions in peridotites and develop methods for the assessment of such parameters.

## METHODS AND OBJECTS

The traditional methods of petrologic geospeedometry are based on the investigation of diffusion profiles of element concentrations in minerals across the boundaries of touching grains, the scale of exsolution textures in mineral grains, or comparison of central zones of mineral grains of various sizes from a single rock (Ozawa, 1984). These methods cannot be applied to a considerable data set from the literature on the mineralogy of mantle peridotites, because the necessary mineralogical features were not addressed in the majority of studies. We believe that more promising may be an integrated approach based on the comparison (with certain assumptions) of the closure temperatures of a particular reaction in different objects (Ozawa, 1986; Bonatti and Michael, 1989; Parkinson and Pearce, 1998). The parallel investigation of two exchange reactions provides the most adequate realization of the advantages of such an approach, which eliminates much of the difficulties of traditional methods.

In this study, two reactions were explored: Ca–Mg exchange between two pyroxenes and Fe–Mg exchange between olivine and spinel. The temperature dependencies of interdiffusion coefficients were determined for these reactions. These coefficients are strongly different, and the respective geothermometers are reliably calibrated. The estimation of the closure temperature of the former reaction was carried out by the Wells (1977) method, which yields values at 900–1100°C consistent with more recent experiments at moderate pressures and provides better extrapolation to low temperatures compared to other expressions. In order to decrease the scatter of estimates, the calculations accounted for aluminum partition between crystallographic positions of pyroxenes estimated assuming charge balance and no ferric iron in pyroxenes. The estimate of the closure temperature of the latter reaction was carried out using the Ballhaus *et al.* (1991) calibration (error in their geothermometer expression was corrected after Portnyagin, 1997). To reduce the uncertainty of estimates, the calculations were carried out for 10 kbar ignoring the concentrations of titanium, aluminum, chromium, and sodium in olivine and calcium and sodium in spinel. Unusually high silica contents (more than 0.20 wt %) in spinel were also ignored, whereas lower SiO<sub>2</sub> contents were recalculated similar to titanium.

New data on the compositions of minerals from peridotites were obtained at the Vernadsky Institute of Geochemistry and Analytical Chemistry, Russian Academy of Sciences on an electron microprobe Camebax–Microbeam using standard procedures, analysts K.I. Ignatenko and N.N. Kononkova. Clinopyroxene compositions were determined in lamellae and inclusions in orthopyroxene at the contact of the two pyroxenes and in the rims of porphyroblasts and neoblasts, where clinopyroxene showed the most stable and reliable, usually also the highest, calcium contents.

Orthopyroxene was analyzed in the matrix at the contact with lamellae and inclusions or in the outer parts of porphyroblasts and neoblasts, where orthopyroxene showed normally the most stable and lowest calcium contents. Olivine was analyzed far (more than 0.5–1.0 mm) from spinel and pyroxene grains. Spinel was usually analyzed in the central parts of grains of the first generation (0.2–0.5 mm on average). In some cases, the average compositions of central and rim grain zones were used. The processing of published data was based, when possible, on the same criteria.

Spinel lherzolites and harzburgites, which are common in all geodynamic settings, were selected as object for this study. The choice of exchange reactions restricted the objects to those where all the four primary minerals were analyzed (pyroxenes, chrome spinel, and olivine). Cumulate rocks, rocks with unusually high iron mole fraction of silicates or anomalous high concentrations of titanium in spinel, as well as rocks with plagioclase and garnet were not used in this study. In order to provide higher reliability of estimates, average values were used for several samples with similar mineral compositions and similar closure temperatures of exchange reactions from a single deep-sea drill hole, a single dredge, or several spatially adjacent dredges.

In addition to the new data, available published and unpublished materials were included into the data set for oceanic peridotites (Table 1). Most of the data represent peridotites from the normal segments of slow-spreading ridges. In addition, peridotites from anomalous zones (15–20 and Azores) and moderately fast spreading ridges (Galapagos spreading center) were also studied.

## PHYSICOCHEMICAL MODELING OF EXCHANGE REACTIONS

The subsolidus recrystallization of mantle spinel peridotites occurs after melt separation from the rocks and continues until the closure of the exchange reaction owing to cooling. In general terms, the closure temperature of each exchange reaction depends on the initial rock temperature and cooling rate. The temperature of melt segregation can be considered as an estimate for the initial temperature of mantle spinel peridotite. This temperature is 1200–1300°C for the majority of mantle melts (Sobolev, 1997). Possible variations in the temperature of melt segregation affect the character of diffusion processes only within a narrow range (ca. 50°C) of the highest subsolidus temperatures and are not important for temperatures below 1150°C. Thus, the controlling factor is the rate of peridotite cooling.

Diffusion processes in minerals are governed by Fick's laws, which include gradients of atom concentrations as major parameters. One of the methods of the approximate solutions of Fick's equations results in an

Table 1. Average characteristics of spinel peridotites from mid-ocean ridges

Hole, dredge, fracture zone	Region	Latitude	Longitude	Cr#, SpI	T <sub>pp</sub> , °C	T <sub>os</sub> , °C	T <sub>pp</sub> -T <sub>os</sub> , °C	N	Source
Dredge 8, 43° N	MAR*	43°13' N	28°56' W	0.50	980	776	204	5	1
Dredge 41, Atlantis	MAR	30°03' N	42°09' W	0.28	1005	773	232	3	2, 3
Hole 920, Kane	MAR	23°20' N	45°01' W	0.29	976	773	203	7	4
Hole 670, Kane	MAR	23°10' N	45°02' W	0.23	1027	792	235	18	5, 6, 7
Hole 395, Kane	MAR	22°45' N	46°05' W	0.38	974	755	219	4	8, 9
Dredges 8, 56, 15-20	MAR*	15°37' N	46°35' W	0.46	990	832	158	3	3, 10
Dredge 77, 15-20	MAR*	15°06' N	44°58' W	0.55	992	828	164	10	3
Dredge 71, 15-20	MAR*	15°05' N	44°57' W	0.62	1010	868	142	8	3
Dredge 68, 15-20	MAR*	14°49' N	45°05' W	0.53	970	772	198	2	3
Dredges 65, 79, Romanche (east)	MAR	0°05' S	16°44' W	0.17	1002	886	116	2	11
Dredge AT196, Romanche (west)	MAR	1°01' S	24°16' W	0.10	954	684	270	2	12
Dredge IN18-20, Owen	AIR	12°36' N	58°13' E	0.20	1010	771	239	4	13
Station 23, Vernadsky	AIR	5°03' N	62°05' W	0.30	978	778	200	2	3, 14
Stations 5319, 28, Vityaz'	AIR	5°24' S	68°31' E	0.23	1049	756	293	4	3
Station 29, 2146, Vema	AIR	9°03' S	67°14' E	0.29	1025	757	268	5	3, 14, 15
Station 5324	SWIR	28°20' S	62°34' E	0.16	956	772	184	6	3, 14
Atlantis II	SWIR	32°01' S	57°09' E	0.16	932	780	152	2	16
Hole 895	GSC	2°17' N	101°27' E	0.53	1068	892	176	11	17, 18
Dredges 4312, 4318	MCSC	18°12' S	81°49' W	0.17	1012	831	181	4	3, 19

Note: Here and in Table 2: MAR, Mid-Atlantic Ridge (asterisk denotes areas within geochemically anomalous segments); AIR, Arabian-Indian Ridge; SWIR, Southwestern Indian Ridge; GSC, Galapagos Spreading Center; and MCSC, mid-Cayman Spreading Center. Cr# = Cr/(Cr + Al). T<sub>pp</sub> is the temperature estimate by the two-pyroxene geothermometer (Wells, 1977), T<sub>os</sub> is the temperature estimate by the olivine-spinel geothermometer (Ballhaus *et al.*, 1991), and N is the number of samples. Sources: 1, Shibata and Thompson, 1986; 2, Bazylev, 1992; 3, authors' unpublished data; 4, Cannat *et al.*, 1997; 5, Fujii, 1990; 6, Juteau *et al.*, 1990; 7, Komor *et al.*, 1990; 8, Arai and Fujii, 1978; 9, Sinton, 1978; 10, Cannat *et al.*, 1992; 11, Pushcharovskii *et al.*, 1995; 12, Bonatti *et al.*, 1993; 13, Hamlin and Bonatti, 1980; 14, L.V. Dmitriev, unpublished data (GEOKHI RAS); 15, Savel'eva, 1987; 16, Johnson and Dick, 1992; 17, Arai and Matsukage, 1996; 18, Dick and Natland, 1996; and 19, Palandzhyan *et al.*, 1990.

expression without concentration gradients (Putnis and MacConell, 1983):

$$Dt/x^2 = \text{const}, \quad (1)$$

where const is close to one, D is the temperature-dependent interdiffusion coefficient of the reaction, t is the time, and x is the diffusion distance for the given time. More specifically, x corresponds to the distance from the mineral boundary, where the influence of diffusion on the initial concentrations of atoms is negligible and the concentration gradient approaches zero. This expression allows us to estimate either the scale of diffusion at the given time and temperature or time necessary for the realization of diffusion for the given distance and temperature. In a more general case, when element concentration in the mineral at the distance x from the boundary differs from the initial concentration, diffusion time will be higher than that derived from equation (1). In this case, the right-hand side of the equation will be higher than one, not necessarily

constant, and dependent on concentration ratios (Putnis and MacConell, 1983):

$$Dt/x^2 \approx (C_p - C_x)/(C_x - C_0), \quad (2)$$

where C<sub>p</sub> is the initial concentration in the phase and C<sub>x</sub> and C<sub>0</sub> are the concentrations at the distance x and at the boundary, respectively, in the time moment t from the onset of diffusion.

The kinetics of the olivine-spinel exchange reaction was studied by Ozawa (1984), who demonstrated that interdiffusion in spinel was the main controlling factor of the rate of this reaction. He also reported the temperature dependence of the Mg-Fe interdiffusion coefficient of this reaction:

$$D_{os} = 30.0 \exp(-40000/T) \text{ (cm}^2/\text{s)}. \quad (3)$$

To minimize the influence of olivine and spinel compositions on cooling rate estimates, Ozawa (1984) used calibration diagrams including the diameter of spinel grains and the closure temperature of the

exchange reaction, which was calculated from the composition of the central part of a spinel grain and olivine composition measured far from the spinel. In this work, the kinetics of the olivine–spinel reaction was calculated from an approximate expression similar to equation (2) with the parameter in the right-hand side calculated by the mathematically more correct model of Ozawa (1984). For an interval of spinel grain size of 0.1–1.0 mm and cooling rates from  $10^{-4}$  to  $1^\circ\text{C}/\text{year}$ , the relationships between grain size, temperature, and time calculated by Ozawa are adequately approximated by the simplified formula:

$$4D_{\text{os}}t/d^2 \approx 17.0, \quad (4)$$

where  $D_{\text{os}}$  is defined by equation (3) and  $d$  is the diameter of the spinel grain in centimeters. The average error of temperature estimates by approximate equation (4) as compared with the values obtained by the Ozawa (1984) expression is  $\pm 15^\circ\text{C}$ .

A petrographic and microprobe investigation of mantle spinel peridotites allowed us to accept a reliable assumption that the average diameter of spinel grains in the rocks studied was approximately uniform, about 0.3 mm. Spinel grains, up to 3 mm across, occur in some samples. In addition to the central portions of such large grains, their rims are usually analyzed as well as medium-sized and small grains. The averaging of sizes and compositions results in approximately the same grain size of 0.3 mm. In general, the plausibility of such an assumption increases with increasing number of samples characterizing an object and the number of analyzed points in spinel in each sample. Similar values of average diameter were obtained for spinel grains in the oceanic spinel peridotites of Hess Deep (Dick and Natland, 1996) and ophiolite peridotites of the Miyamori (Ozawa, 1984) and Tamvatnei massifs (Palandzhyan, personal communication, 1998).

The temperature dependence of the Ca–Mg interdiffusion coefficient in clinopyroxene was given by Sanford and Huebner (1979):

$$D_{\text{pp}} = 0.00389 \exp(-43400/T) \text{ (cm}^2/\text{s)}. \quad (5)$$

Similar to the olivine–spinel reaction, the relationships between diffusion scale, temperature, and time can be approximated by equation (2), whose right-hand side can be estimated from orthopyroxene compositions in oceanic spinel peridotites. According to our data, the initial CaO content of orthopyroxene averages at 2.7 wt %, the average CaO content near the contact with clinopyroxene in the rim parts of grains is 1.2–1.5 wt %, and the lowest content, 0.8–1.0 wt % is observed directly at the contact with clinopyroxene. Thus, the value of the “constant” of this reaction is estimated as 3.5–4.0. The compositions of minerals from ophiolite peridotites (Bazylev *et al.*, 1993) allowed us to conclude that this parameter was no lower than 4.0. Thus, the diffusion parameters of the two-pyroxene

exchange reaction can be estimated semiquantitatively from the expression:

$$D_{\text{pp}}t/x^2 \approx 4.0, \quad (6)$$

where  $D_{\text{pp}}$  is defined by equation (5) and  $x$  (cm) is the distance from the analyzed point to the  $\text{Cpx–Opx}$  interface.

A reasonable assumption with respect to the whole data set on mantle spinel peridotites is that all the measurements of orthopyroxene rims, orthopyroxene in contact with clinopyroxene, and matrix near clinopyroxene lamellae (as well as clinopyroxene compositions in similar positions) were carried out at a certain average distance from the phase contact. This distance is normally from 10 to 20  $\mu\text{m}$  in the case of grain boundaries and from 5 to 10  $\mu\text{m}$  in the case of lamellae. Thus, the average value of  $x$  can be taken as 10  $\mu\text{m}$ .

For a certain diffusion scale, equations (4) and (6) relate the closure temperature of the reaction with time, and their estimation is possible if another relation between them will be defined. The character of rock cooling, i.e., change in temperature with time, can be used for this purpose. The validity of the method and assumptions can be assessed on the basis of an analysis of the simplest model of constant-rate cooling within the whole temperature interval.

#### CONSTANT RATE OF PERIDOTITE COOLING

At constant cooling rate, the relationship between rock temperature and time is the following:

$$T = 1250 - V_i t, \quad (7)$$

where  $T$  is the temperature ( $^\circ\text{C}$ ) in the time moment  $t$  (y) after melt separation at the cooling rate  $V_i$  ( $^\circ\text{C}/\text{year}$ ).

The calculation of the closure temperature of exchange reactions was performed by the iterative solution of equations (3), (4), and (7) for the reaction  $\text{Ol–Spl}$ ; and (5), (6), and (7) for the reaction  $\text{Cpx–Opx}$  at  $d = 0.3$  mm and  $x = 10$   $\mu\text{m}$ . Within this model and at fixed  $d$  and  $x$ , any value of cooling rate corresponds to certain values of the closure temperatures of the olivine–spinel Fe–Mg exchange reaction ( $T_{\text{os}}$ ) and the two-pyroxene Ca–Mg exchange reaction ( $T_{\text{pp}}$ ) (Fig. 1). The diagram  $T_{\text{os}}-T_{\text{pp}}$  is convenient for the interpretation of the thermal history of peridotites. A point on this diagram corresponds to a certain value of cooling rate, and combinations of various cooling rates are shown by lines, whose position depends on the diffusion scales of the exchange reactions under consideration.

#### Geodynamic Consequences

In contrast to spinel peridotites from ophiolite and xenoliths, the peridotites of mid-ocean ridges show rather narrow ranges of temperatures estimated by the two exchange reactions:  $995 \pm 34^\circ\text{C}$  for the two-pyroxene thermometer and  $793 \pm 51^\circ\text{C}$  for the olivine–spinel

thermometer. The difference between these values is  $202 \pm 47^\circ\text{C}$ . Parkinson and Pearce (1998) estimated, independently (most of their data were not used in this study), the average closure temperature of the olivine-spinel exchange reaction for oceanic peridotites as  $806 \pm 75^\circ\text{C}$ . This value is virtually identical to our estimate.

The narrow range of temperatures obtained for the MOR spinel peridotites suggests that the cooling of these rocks occurred through a single mechanism characteristic of the MOR environment. Calculations (Fig. 2) demonstrate that within the temperature interval of the closure of exchange reactions, the cooling rate of oceanic peridotites was rather high and  $0.01^\circ\text{C}/\text{year}$  on average with an order of magnitude variations. Assuming that the cooling rate of oceanic peridotites was constant through the whole subsolidus history and the temperature interval of cooling was  $1250^\circ\text{C}$ , the time between melt segregation and exposure of peridotites at the ocean floor was approximately 125000 years. Taking the probable depth interval of melt segregation, 5–23 kbar (Sobolev, 1997), the ascent velocity of peridotite can be estimated as 12–50 cm/year. This estimate is an order of magnitude higher than the probable ascent velocity of mantle material beneath mid-ocean ridges (1–10 cm/year), which, according to mantle convection models, must be close to half of total spreading rate (Reid and Jackson, 1981). It is evident that the assumption of the constant-rate cooling of oceanic peridotites from 1250 to  $0^\circ\text{C}$  and estimates of the rate of this process from the data on the closure temperatures of exchange reactions ( $750$ – $1050^\circ\text{C}$ ) resulted in significant errors in geological interpretation.

Nevertheless, the preliminary analysis of the results obtained from the simplest model suggests the validity of our approach and the plausibility of assumptions and constraints in the diffusion model of exchange reactions. Physical models of mantle peridotite cooling accounting for a particular geodynamic environment should be used to refine some parameters of geologic significance.

### ASCENT OF PERIDOTITE IN THE AXIAL ZONE OF MID-OCEAN RIDGES

In general terms, three possible cooling mechanisms must be considered in the case of oceanic mantle peridotites: adiabatic (convection) cooling; conductive cooling; and cooling due to fluids circulation via convection beneath mid-ocean ridges (Lister, 1972; Bottinga and Allegre, 1973; Cannat, 1993).

#### Adiabatic Cooling

Adiabatic cooling results from a pressure decrease during rock ascent. Its rate ( $V_p$ ,  $^\circ\text{C}/\text{year}$ ) is controlled by adiabatic gradient (about  $0.4^\circ\text{C}/\text{km}$ ) and ascent velocity ( $V$ ,  $\text{cm}/\text{year}$ ):

$$V_p = 0.4 \times 10^{-5} V. \quad (8)$$

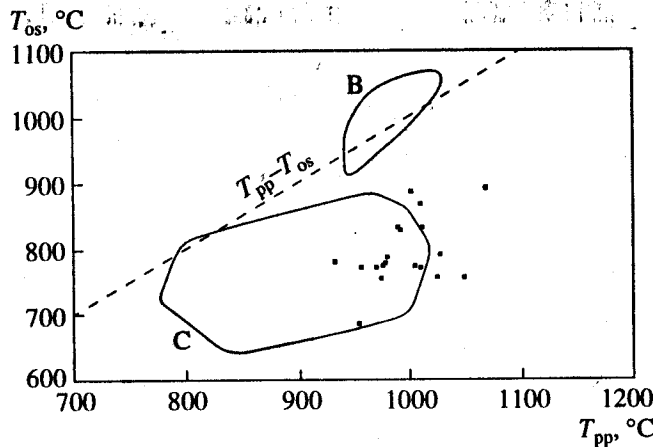


Fig. 1. Closure temperature of the two-pyroxene (Ca-Mg),  $T_{pp}$  and olivine-spinel (Fe-Mg),  $T_{os}$  exchange reactions in oceanic spinel peridotites.

B is the field of typical peridotite xenoliths (Bazylev and Silant'ev, 2000) and C is the field of peridotites from ophiolites (Bazylev and Silant'ev, 2000).

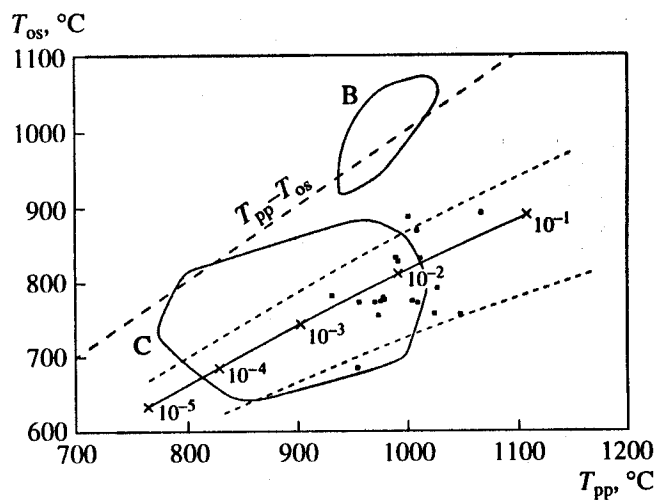
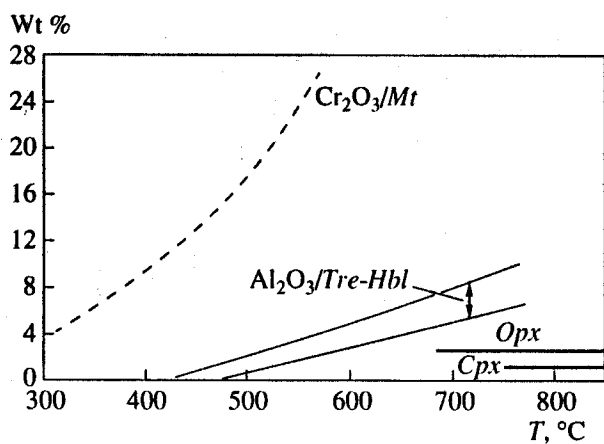


Fig. 2. Comparison of the real closure temperatures of exchange reactions in natural spinel peridotites with temperatures calculated from the model of constant cooling rate (from  $0.1$  to  $10^{-5}$   $\text{H}^\circ\text{C}/\text{year}$ ).

Calculation parameters: initial temperature,  $1250^\circ\text{C}$ ; average grain diameter of spinel,  $d = 0.3$  mm; and pyroxene compositions analyzed at a distance of  $x = 10$   $\mu\text{m}$  from grain contacts. Dotted lines show the field of closure temperatures at variations in the parameters:  $d$ , from  $0.1$  to  $0.5$  mm and  $x$ , from  $5$  to  $20$   $\mu\text{m}$ . See Fig. 1 for other symbols.

According to the model of passive spreading, the ascent velocity of mantle material beneath the axial parts of MOR can be taken as half of the total spreading velocity. At an ascent velocity of mantle material in the axial spreading zone of  $5$   $\text{cm}/\text{year}$ , the adiabatic cooling rate is  $2 \times 10^{-5}$   $^\circ\text{C}/\text{year}$ , which is two–three orders of magnitude lower than the cooling rate of MOR peridotites estimated above. Thus, the contribution of the adi-



**Fig. 3.** Changes in the characteristics of metamorphic minerals (chrome-rich magnetite and calcic amphibole) from spinel harzburgites and lherzolites versus metamorphic temperature.

Aluminum content in calcic amphibole depends on temperature and rock composition and is shown, therefore, as a field. Also shown are temperature intervals of the stability of orthopyroxene–clinopyroxene–amphibole association.

abatic term into the total cooling rate of MOR peridotites is negligible and can be ignored.

#### *Convection Cooling: The Upper Lithospheric Layer*

The convection heat transfer defines the character of cooling of the upper lithospheric layer within MOR owing to the circulation of pore fluids (Lister, 1972). In general, the intensity of circulation drops with increasing depth and decreasing rock permeability. Nevertheless, the highest temperatures of metamorphism related to fluid circulation beneath MOR is about 650°C in metabasic rocks (Silantyev, 1995) and 800°C in metamorphosed peridotites (Bazylev, 1992).

The temperatures of metamorphism caused by the circulation of hydrothermal fluids in peridotites can be determined by the olivine–spinel geothermometer; solvus configuration in medium-temperature spinels (Sack and Ghiorso, 1991); spinel–chlorite–olivine and amphibole–chlorite geothermometers (Bazylev *et al.*, 1990); and, if metamorphic clinopyroxene occurs, two-pyroxene geothermometer (Wells, 1977). Temperature can be constrained from the compositional characteristics of metamorphic minerals in peridotite, in particular, the alumina content in calcic amphibole (tremolite and hornblende) and the chromium content of magnetite coexisting with ferrite–chromite. Figure 3 shows variations of these parameters with temperature estimated by the above-mentioned geothermometers. Metamorphic orthopyroxene–amphibole assemblages are characteristic of temperatures above ca. 680°C and clinopyroxene–amphibole, above ca. 750°C.

On average, a temperature of 550–600°C is the most common maximum value of hydrothermal metamor-

phic conditions for both metabasic rocks and metamorphosed peridotites from various MOR segments (Bazylev *et al.*, 1990; Silantyev, 1995; Bazylev, 1998), which can be considered as typical of the lower limbs of convection fluid flows. This temperature is evidently much lower than the closure temperature of the exchange reactions under consideration (Table 1). Thus, the cooling of oceanic peridotites at temperatures close to the closure of the exchange reactions occurred mainly beyond the direct influence of the convective fluid circulation.

Since convection is at least an order of magnitude more efficient in heat transfer than conduction, the latter mechanism can be ignored at the first approximation for the upper permeable layer of the mid-ocean ridge lithosphere.

The thickness of the permeable layer of the lithosphere in a particular MOR segment can be estimated from the maximum temperature of metamorphism resulting from fluid circulation in the rocks of this segment. Such estimates are based on the assumption that the geothermal gradient was constant through the whole depth interval. Proceeding from the direct petrologic estimates of parameters corresponding to the case of the deepest fluid penetration into the MOR lithosphere in the northern slope of the Atlantis Fracture Zone valley, Atlantic Ocean (820°C and 7.0–7.5 kbar; Bazylev, 1992), the gradient is estimated as 35°C/km. This parameter can also be assessed from the depth of the 750°C geotherm, which was calculated from the geophysical model accounting for the segmentation geometry of axial MOR zones as 15 km in the zones of large transform faults and 7 km in the portions of the rift valley remote from transform faults (Neuman and Forsyth, 1993). The corresponding geothermal gradient values are 50°C/km and more than 100°C/km. The latter value is evidently overestimated, because the maximum temperature of hydrothermal metamorphism in areas remote from transform faults is usually lower than 750°C and the respective geotherm lies below the permeable lithospheric layer. An estimate of 100°C/km does not characterize the thermal gradient of the circulation zone. Nevertheless, it is reasonable to expect elevated values of geothermal gradients in the segments of rift valleys remote from transform faults. As a first approximation, these variations can be ignored and a value of 40°C/km can be accepted as average geothermal gradient in the permeable layer of the MOR lithosphere. The thickness of this layer is estimated by the relation,

$$a_h = T_h/40, \quad (9)$$

where  $a_h$  is the thickness of the fluid circulation zone; i.e., the upper permeable lithospheric layer in kilometers and  $T_h$  is the temperature at the base of this layer (°C) corresponding to the maximum metamorphic temperature in the rocks of this segment (Fig. 4).

*Conductive Cooling: Lower Lithospheric Layer*

Cooling via conductive heat transfer occurs in the mantle peridotites of MOR in the lower, fluid-impermeable layer of the lithosphere. The temperature of the lower boundary of this layer ( $T_0$ ) corresponds to the conditions of melt separation and can be taken as 1250°C. The temperature of the upper boundary ( $T_h$ ) corresponds to the temperature of the maximum depth of fluid penetration. The cooling of mantle peridotite in this layer beneath the axial zone of mid-ocean ridges can be modeled by the cooling of a hot half-space with constant temperature on the surface,  $T_h$ . Changes in temperature ( $T$ ) with time ( $t$ ) in any point at the distance  $a$  from the upper boundary can be calculated from the equation (Jaeger, 1968):

$$(T - T_h)/(T_0 - T_h) = 1/2 - 1/2 \operatorname{erf}(a/(2\sqrt{kt})), \quad (10)$$

where  $k$  is the thermal diffusivity of mantle peridotite, 0.0085 cm<sup>2</sup>/s (Parker and Oldenburg, 1973);  $a$  is the distance, cm; and the error function is defined as,

$$\operatorname{erf}(u) = \frac{2}{\sqrt{\pi}} \int_0^u \exp(-z^2) dz. \quad (11)$$

Since mantle peridotites rise beneath mid-ocean ridges, the distance of heat transfer for a given point decreases with time depending on the ascent velocity of mantle material:

$$a = a_0 - Vt, \quad (12)$$

where  $a_0$  is the distance from the point to the upper boundary of the layer of conductive heat transfer in the initial moment,  $t = 0$  (moment of melt separation) and  $V$  is the ascent velocity of mantle material.

Thus, the closure temperature of the exchange reaction in MOR axial zone peridotites can be calculated by the simultaneous solution of equations (5), (6), (10), and (12) for the two-pyroxene reaction or (3), (4), (10), and (12) for the olivine-spinel reaction. Particular temperatures depend on the magnitude of the parameters  $a_0$ ,  $T_h$ , and  $V$ , and the effects of these parameters are not equal. The maximum possible variation in the upper boundary temperature of the conductive heat transfer layer (from 750 to 250°C) changes the closure temperatures of the exchange reactions by 20–40°C, which is equivalent to the effect of changing  $a_0$  by a factor of less than two (Fig. 5a). The maximum variation in the ascent velocity of mantle material beneath MOR (from 12.8 to 0.8 cm/year) changes the closure temperature of the exchange reactions by 20–60°C, which is also approximately equivalent to a two-fold increase in  $a_0$  (Fig. 5b). Since the closure temperatures of the exchange reactions in various MOR zones vary within 110–210°C (Table 1), the parameter  $a_0$  is the most important factor affecting these parameters: a two-fold decrease in the thickness of the conductive heat transfer zone raises the closure temperature by 40–65°C (Figs. 5a, 5b). It is possible to assess the thickness of the conductive heat transfer zone (i.e., the thickness of the

$a$ , km

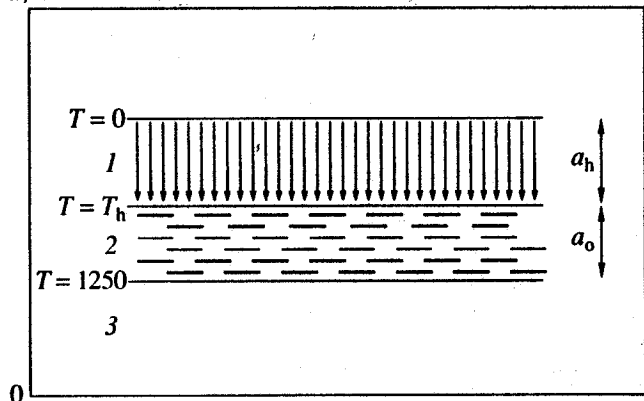


Fig. 4. A schematic cross-section of the lithosphere of the axial part of MOR. 1, Upper layer permeable for seawater; 2, lower layer impermeable for seawater and melt; and 3, sublithospheric mantle (peridotite with interstitial melt).

lower, impermeable layer of the MOR lithosphere) from the closure temperatures of exchange reactions. At an average ascent velocity of mantle material of 3.2 cm/year, the dependencies of the thickness of the impermeable layer of the lithosphere ( $a_0$ , km) on the closure temperatures of exchange reactions ( $T_{pp}$  and  $T_{os}$ , °C) and the temperature of the upper boundary of the conductive heat transfer zone ( $T_h$ , °C) can be approximated by the following equations:

$$\log(a_0) = 6.57 - 4 \times 10^{-4} T_h - 0.0059 T_{pp} \quad (13)$$

and

$$\log(a_0) = 7.86 - 7 \times 10^{-4} T_h - 0.0087 T_{os}. \quad (14)$$

The estimates of the thickness of the lower lithospheric layer by equations (13) and (14) are independent. Thus, the mean of the two values is a more accurate estimate for any particular MOR segment.

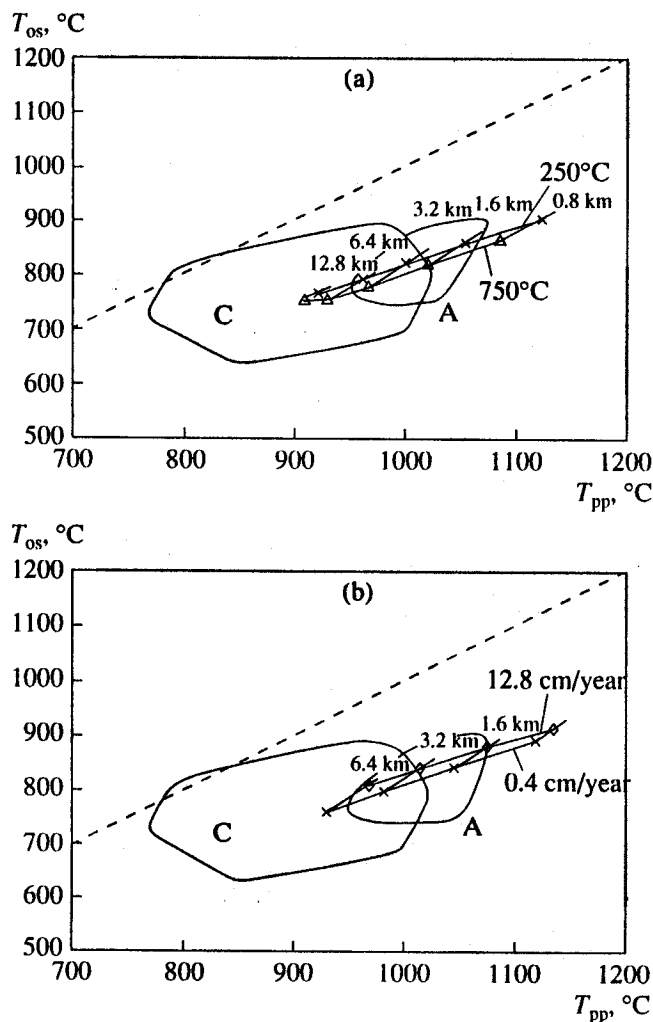
The simultaneous solution of equations (5) and (6) allows calculation of time necessary for the attainment of the closure temperature of the two-pyroxene exchange reaction ( $t_q$ ) during rock cooling. For  $x = 10 \mu\text{m}$ , such a dependency is the following:

$$t_q = 3.265 \times 10^{-11} / \exp(-43400/(T_{pp} + 273)), \quad (15)$$

where  $t_q$  is in years and  $T_{pp}$  is in degrees centigrade.

*Intersection Zones of a Rift Valley and a Transform Fault*

A number of models of the structure of the oceanic lithosphere near the intersection of a rift valley and a transform fault assume that the newly-formed lithosphere of the rift valley cools owing to lateral conductive heat transfer into an older thick and cold lithosphere through the zone of the transform fault (Fox and Gallo, 1984). However, taking into account that large fracture zones are characterized by the deepest penetration of



**Fig. 5.** Comparative influence of the thickness of the lithospheric layer of conductive cooling, the temperature at the base of the permeable layer of the lithosphere,  $T_h$  (a), and the ascent velocity of mantle material,  $V$  (b) on the closure temperatures of exchange reactions in MOR peridotites.

(a)  $V = 3.2$  cm/year,  $T_h = 250$ – $750°C$ , and  $a = 0.8$ – $12.8$  km; (b)  $T_h = 250°C$ ,  $V = 0.4$ – $12.8$  cm/year, and  $a = 0.8$ – $6.4$  km. A is the field of typical MOR peridotites and C is the field of ophiolitic peridotites.

fluids into the oceanic lithosphere (Kimball *et al.*, 1985; Bazylev, 1992; Cannat and Seyler, 1995), it is more reasonable to suggest that transform fault valleys are zones, where lithosphere cooling through convective hydrothermal heat transfer occurs to considerable depths. The lateral conductive cooling of contacting lithosphere blocks takes place independently in two separate zones divided by a central zone, where changes in temperature with depth within the zone of fluid circulation is controlled by the geothermal gradient.

At such a structure of the areas of axial MOR zone intersection by transform faults,  $a_0$  obtained from equations (13) and (14) for peridotites ascending near the

faults will correspond to the minimum distance from the point of melt segregation to the surface of fluid penetration zone rather than to the vertical thickness of the conductive cooling layer.  $a_h$  and  $T_h$  determined for these peridotites will be much lower than the depth and maximum temperature of fluid penetration in the fault zone, which will control the rate of conductive cooling of peridotites under such circumstances. Calculations show that at moderate slope of this surface (up to ca  $60°$ ), equations (13) and (14) yield reasonable values. At steeper slopes,  $a_0$  (and, consequently,  $a$ ) will be lower than the true value. However, ridge areas with such a high gradient of the depth of hydrothermal fluid penetration are probably rather small (less than 8 km from the boundary of the zone of the deepest fluid penetration, which can be accepted as corresponding to the flank of the fault valley). Thus, within the proposed model, it is not necessary to consider the areas of MOR intersection by the valleys of transform faults separately from the axial parts of the MOR remote from transform faults.

### Geodynamic Consequences

A principal geological consequence of the model is that it allows estimation of the thickness of the lithosphere in the central part of mid-ocean ridges (a) and, hence, the depth of the separation of last melt portions from mantle peridotite on the basis of the closure temperatures of solid-phase exchange reactions, orthopyroxene–clinopyroxene and olivine–chrome spinel, and temperatures of metamorphism caused by the descending limbs of hydrothermal systems. Moreover, it is possible to estimate the time between melt separation from peridotite and closure of the two-pyroxene exchange reaction in the peridotite ( $t_q$ ) and the exposure of the rock on the ocean floor ( $t_{ex}$ ).

The calculations demonstrated (Table 2, Fig. 6) that the average thickness of the lower impermeable layer of the oceanic lithosphere beneath the central zone of mid-ocean ridges was 4.4 km and varied from 0.8 to 6.6 km (except for the western flank of the Romanche Fracture Zone in the Atlantic Ocean). The closure of the two-pyroxene exchange reaction in peridotite occurs on average 35 ka after melt separation from the rock. During this time, mantle peridotite cools by  $250°C$  and moves away from the zone of melt separation by 1.1 km. These average values suggest that the lower portion of the impermeable layer of the oceanic lithosphere is characterized by very high geothermal gradient, ca.  $250°C/km$ .

Another inference from these estimates is that the pressure of the closure of the two-pyroxene (Ca–Mg) exchange reaction in MOR peridotites is lower than the pressure of separation of last melt portions from these rocks by a negligible value, ca. 0.3 kbar. Thus, pressure at the closure temperature of the two-pyroxene exchange reaction in oceanic peridotite is essentially equal to the pressure of melt segregation.



Table 2. Additional parameters and average calculated data for the spinel peridotites of MOR

Location (dredge, hole, fracture zone)	Region	$T_h$ , °C	$a_0$ , km	$a$ , km	$P$ , kbar	$t_q$ , Ma	$t_{ex}$ , Ma	Cr#, Spl	$V_s$ , cm/year	Al <sub>2</sub> O <sub>3</sub> liq, wt %
43° N	MAR*	<u>650</u>	3.9	20.2	6.6	0.04	1.6	0.50	2.5	14.5
Atlantis	MAR	820	2.8	23.3	7.6	0.02	1.7	0.28	2.8	16.7
Hole 920, Kane	MAR	<u>650</u>	4.2	20.4	6.7	0.04	1.4	0.29	3.0	16.8
Hole 670A, Kane	MAR	<u>650</u>	2.5	18.8	6.2	0.01	1.3	0.23	3.0	17.5
Hole 395, Kane	MAR	550	6.0	19.8	6.5	0.04	1.3	0.38	3.0	16.0
Dredge 8, 56, 15–20	MAR*	650	2.2	18.5	6.1	0.03	1.3	0.46	2.8	15.0
Dredge 77, 15–20	MAR*	560	2.5	16.5	5.4	0.03	1.2	0.56	2.8	14.0
Dredge 71, 15–20	MAR*	500	1.7	14.2	4.8	0.02	1.0	0.62	2.8	13.2
Dredge 68, 15–20	MAR*	580	4.8	19.3	6.3	0.05	1.4	0.53	2.8	14.2
Romanche (east)	MAR	550	1.7	15.4	5.1	0.02	0.9	0.17	3.5	18.4
Romanche (west)	MAR	550	19.3	33.1	10.6	0.08	1.9	0.10	3.5	17.3
Owen	AIR	560	4.1	18.1	6.0	0.02	2.3	0.20	1.6	17.9
Vernadsky	AIR	600	4.2	19.2	6.3	0.04	1.3	0.30	3.0	16.8
Vityaz'	AIR	670	3.9	20.7	6.8	0.01	1.1	0.23	3.8	17.4
Vema	AIR	350	6.6	15.3	5.1	0.01	0.8	0.29	3.9	17.2
Station 5324	SWIR	690	4.5	21.8	7.1	0.07	2.6	0.16	1.7	18.0
Atlantis II	SWIR	550	6.0	19.7	6.5	0.14	2.3	0.16	1.7	18.2
Hole 895	GSC	550	0.8	14.6	4.9	0.004	0.6	0.53	5.0	14.4
Dredges 4312, 4318	MCSC	350	2.7	11.4	3.9	0.02	1.3	0.17	1.8	18.7

Note: Estimates of the maximum temperature of hydrothermal metamorphism ( $T_h$ ) are mainly after Bazylev *et al.* (1990), Bazylev (1992), and unpublished data of the authors. Underlined values are our approximate estimates from individual published compositions of metamorphic minerals. An average temperature of 550°C (shown in italics) was accepted if no data on the metamorphism of peridotite were available.  $a_0$  is the thickness of the lower impermeable layer of the lithosphere;  $a$ , total thickness of the lithosphere ( $a = a_0 + T_h/40$ );  $P$ , pressure at the base of the lithosphere ( $P = a/3.2 + 0.3$ , where  $P$  in kilobars,  $a$  in kilometers, and 0.3 is the average correction for water pressure);  $t_q$ , time between melt separation and the closure of the two-pyroxene exchange reaction; and  $t_{ex}$ , time between melt separation and exposure of rocks on the ocean floor. The estimates of total spreading velocities ( $V_s$ ) for sampling sites are given after Galushkin and Ushakov (1978). The concentration of Al<sub>2</sub>O<sub>3</sub> in the last melt portions separated from peridotite was estimated on the basis of spinel composition and assessed pressure (Bazylev, 1995).

According to the calculation method, the thickness of the upper permeable layer of the oceanic lithosphere beneath the central zone of mid-ocean ridges depends strongly on the maximum temperature of hydrothermal fluid penetration, which is known only for some of the MOR regions considered. The thickness of this layer averages at 14.6 km and varies from 8.8 to 20.5 km. Thus, the total thickness of the lithosphere in the places of spinel peridotite sampling within MOR averages at 19 km and varies from 11.4 to 23.3 km. An exception is the above-mentioned western part of the Romanche Fracture Zone, where the total thickness of the lithosphere is estimated as 33 km.

If the ascent velocity of peridotites is half the total spreading velocity, spinel peridotites come to the ocean floor on average in  $1.4 \pm 0.5$  Ma after melt separation from them. This allows us to address on a quantitative basis the problem of the genetic correlation of basalts, gabbroids, and peridotites found in the same sites within MOR (Silantyev *et al.*, 1995).

## DISCUSSION

### *Structure of the Lithosphere and Characteristics of Igneous Processes of Mid-Ocean Ridges*

The specific character of the geodynamic setting of mid-ocean ridges results from the fact that the processes of magmatism, subsolidus recrystallization of residues, their crustal metamorphism, and exposure on the ocean floor are interrelated within a single dynamic event. This allowed us to estimate the thickness of the lower lithospheric layer from the rate of peridotite cooling during subsolidus metamorphism and the thickness of the upper lithospheric layer from the temperature of crustal metamorphism. Thus, data on metamorphic conditions provides evidence on the structure of the lithosphere and depth of melt separation from solid residues.

Our values of lithosphere thickness in the axial parts of mid-ocean ridges (from 11.4 to 23.3 km and up to 33 km in the western part of the Romanche Fracture Zone) are in general consistent with estimates derived from geochemical (Niu and Batiza, 1991) and geophys-

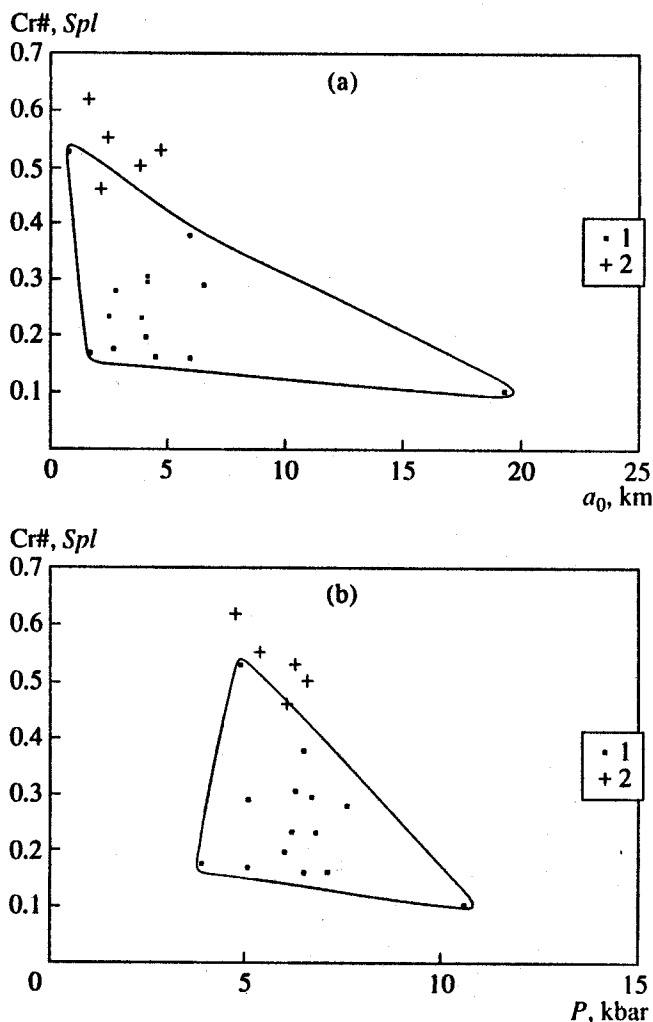


Fig. 6. Correlation of Cr/(Cr + Al) ratio of spinels from MOR spinel peridotites with the calculated thickness of the impermeable layer of the lithosphere,  $a_0$  (a) and calculated pressure at the separation of last melt portions (b).

1, Peridotites from normal MOR segments; 2, peridotites from geochemically anomalous MOR segments; and 3, peridotites from the western Romanche Fracture Zone, Atlantic Ocean.

ical data (Neuman and Forsyth, 1993). Proceeding from the estimates of lithosphere thickness, the pressure of last melt separation from the spinel peridotites of MOR varied from 3.9 to 7.6 kbar (10.6 kbar at the western Romanche Fracture Zone), which is consistent with values obtained from studies of melt inclusions in minerals from oceanic basalts (4–15 kbar; Sobolev and Dmitriev, 1989).

The concentration of alumina in melt can be estimated from the chromium content of spinel in the residual peridotite and the pressure of separation of the last melt portion (Bazylev, 1995):

$$\text{Al}_2\text{O}_3\text{liq} = [1000(1 - \text{Cr}\#, \text{Spl})]^{0.444 - 0.021P}, \quad (16)$$

where alumina content of the melt is in weight percents and  $P$  is the pressure in kilobars. According to the calculations (Table 2), alumina concentration in last melt portions separating from the oceanic peridotites was 14.4–18.7 wt % and 16.0–18.7 wt % in normal segments of slow and moderately spreading ridges. These values are similar to the estimates of  $A_8$  (concentration of  $\text{Al}_2\text{O}_3$ , wt % recalculated to the reference degree of fractionation corresponding to an MgO content of 8 wt %) in basalts from the normal segments of mid-ocean ridges, 14.5–17.0 (Niu and Hekinian, 1997).

It should be kept in mind that the above calculations of the structure of the oceanic lithosphere in the axial parts of mid-ocean ridges were based on the data on those segments of ridges, where mantle peridotites were found. The fact of mantle peridotite occurrence in the axial zones of ridges is at odds with the classic layered model of the oceanic lithosphere. However, the data currently accumulated allow us to believe that areas of the oceanic crust with distorted structures are rather widespread in MOR (Cannat, 1993).

The considerable difference between the closure temperatures of the two-pyroxene and olivine–spinel exchange reactions (116–293°C) in the spinel peridotites of MOR suggests that there was no high-temperature isothermal stage in the geologic history of oceanic spinel peridotites; i.e., they cooled continuously from the moment of melt separation to the exposure on the oceanic floor.

The fact that spinel peridotites appear to be brought to the ocean floor in the axial part of mid-ocean ridges is in apparent conflict with the occurrence of magma chambers beneath these areas (Cannat, 1993). During their ascent near a magma chamber, spinel peridotites must have experienced repeated heating or an isothermal stage, which are not recorded in the rocks studied. This contradiction can be solved suggesting that the MOR areas with mantle peridotite outcrops are spatially separated from those containing magma chambers (Cannat, 1993). It is conceivable also that magma chambers are enclosed by plagioclase-bearing peridotites, wehrlites, or dunites rather than spinel lherzolites or harzburgites, i.e., by the rocks that were not considered in this work. In any case, the characteristics of the thermal history of oceanic peridotites suggest that crustal magma chambers beneath ridges either occur locally or operate discontinuously during short time periods.

#### Characteristics of Geochemically Anomalous MOR Segments

The estimated thickness of the lithosphere in the axial zones of geochemically anomalous MOR segments (17.7 km on average) does not differ significantly from that of normal MAR segments (21.8 km on average) and Indian Ocean ridges (19.1 km on average).

The ascent velocity of mantle material is one of the factors controlling the degree of mantle partial melting beneath MOR (Nicolas, 1986; Niu and Hekinian, 1997). According to the model of mantle convection, this velocity is proportional to spreading velocity. In fact, there is a correlation between chrome content of spinel and spreading velocity in oceanic spinel peridotites from normal segments (Fig. 7). It is remarkable that spinel peridotites from geochemically anomalous MOR segments (Azores and Fifteen Twenty) show strong deviation from this correlation recording an unusually high depletion independent of spreading velocity. In addition to the peridotites of geochemically anomalous zones, peridotites from the western flank of the Romanche Fracture Zone deviate from this correlation though in a different direction (Fig. 7). These rocks differ from typical oceanic peridotites in lower closure temperatures of exchange reactions and, consequently, higher calculated depths of melt separation. Furthermore, these rocks show the lowest degree of depletion for oceanic peridotites (Bonatti *et al.*, 1993). Because of this, the relation of these rocks to the modern magmatism of the Mid-Atlantic Ridge is questionable.

The spinel peridotites of geochemically anomalous MOR segments show a number of mineralogical and petrologic characteristics that are not typical of the peridotites of normal MOR segments, but are common in the peridotites of some ophiolite massifs and fast-spreading MOR segments (Hole 895). Among such characteristics are high Cr/(Cr + Al) ratios of chrome spinel (0.46–0.62), low pressure of separation of the last melt portion (4.8–6.6 kbar), and low alumina contents of segregated melts (13.2–15.0 wt %  $\text{Al}_2\text{O}_3$ ) (Table 2). The high degree of depletion of mantle residues and shallow melt segregation in the case of peridotites from Hole 895 correlate with the rapid ascent of mantle material (Fig. 7). The combination of these features in the peridotites of geochemically anomalous MOR segments results from the additional influence of a local nontectonic factor, most likely, the input of water and presumably potassium into the magmatic system (Bonatti *et al.*, 1992; Cannat *et al.*, 1992).

## CONCLUSIONS

(1) The closure temperatures of the two-pyroxene (Ca–Mg) and olivine–spinel (Fe–Mg) exchange reactions in the mantle spinel peridotites of mid-ocean ridges record the thermal regime of the rocks at subsolidus conditions after melt separation. The average closure temperatures of the two-pyroxene ( $997 \pm 29^\circ\text{C}$ ) and olivine–spinel ( $801 \pm 47^\circ\text{C}$ ) exchange reactions suggest rapid (ca.  $10^{-2}^\circ\text{C}/\text{year}$ ) and continuous cooling of the rocks from melt separation conditions to temperatures below  $650^\circ\text{C}$ .

(2) Two layers are distinguished in the lithosphere of axial MOR zones differing in the dominating mechanism of heat transfer. The upper lithospheric layer is permeable for hydrothermal fluids circulating beneath

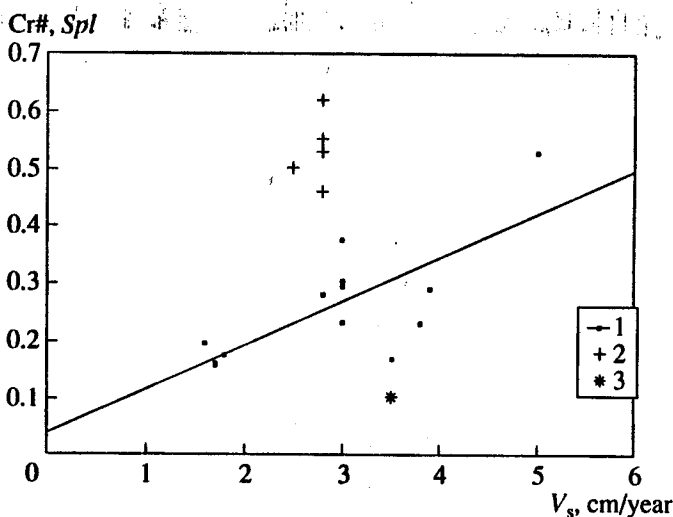


Fig. 7. Correlation of the Cr/(Cr + Al) ratio (Cr#) of spinels from oceanic spinel peridotites with spreading velocity,  $V_s$ . 3, Peridotites from the western part of the Romanche Fracture Zone, Atlantic Ocean.

See Fig. 6 for the other symbol explanation.

MOR. Heat transfer in this layer occurs via convection. The lower lithospheric layer is fluid-impermeable, and conductive heat transfer prevails in this zone.

(3) The thickness of the upper layer was estimated from the highest temperature of hydrothermal metamorphism recorded in spinel peridotites assuming constant geothermal gradient within this layer. The average value is 15 km at variations from 9 to 21 km. The thickness of the lower layer is estimated from the closure temperatures of exchange reactions in the spinel peridotites of MOR, the average value is 4.4 km at variations from 0.8–19 km.

(4) The total thickness of the lithosphere in the axial zone of MOR within the segments studied varies from 11–33 km and averages at 18.4 km. This corresponds to a pressure interval of 3.9–10.6 kbar for the separation of last melt portions from the peridotites.

(5) The peridotites of normal MOR segments show a positive correlation between the Cr/(Cr + Al) ratio of spinel and spreading velocity and a negative correlation between the former value and the calculated depth of melt separation. This suggests a decrease in the thickness of the lithosphere with increasing spreading velocity in normal MOR segments.

(6) The peridotites of geochemically anomalous MOR segments deviate from the correlation between the Cr/(Cr + Al) ratio of spinel and spreading velocity in that their spinel is unusually rich in chromium, which results from a high degree of melting of the rocks owing to nontectonic factors, which do not occur in normal segments.

(7) The spinel peridotites of the MOR axial zones studied, including those from flanks of a rift valley, do not bear evidence of repeated heating or an isothermal

high-temperature stage. This suggests that no large magma chambers exist beneath the sampled segments of the axial parts of MOR.

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