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Thermal structure of the axial zone in mid-oceanic ridges. Part 1. Formation and evolution of axial magma chamber

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Abstract. We considered the origin and evolutionary conditions for the axial magma chamber (AMC) in the rift zone of mid-oceanic ridges (MOR), in the framework of the continuous-discrete spreading model. Calculations show that the depth to the chamber top depends on spreading rate, periodicity of the tectono-magmatic cycle, fracturing in the crust, hydrothermal activity and thermophysical properties of MOR rocks. Numerical modelling suggests that the quasi-stationary AMC structure in the axial MOR zone likely does not exist if the spreading rate is less than 3 cm yr^{-1}. At greater spreading rates this structure is developed in the crust of axial MOR zone over a period of 150,000–200,000 years. Analysis of evolution of the AMC chamber shape on its cooling stage demonstrates that an interruption between the periodic intrusions of more than 100,000 years results in the disappearance of the magma chamber.

Introduction

In the last decade the rift zones of mid-oceanic ridges (MOR) and back-arc basins have attracted considerable attention from marine geologists and geophysicists in connection with the discovery here of deep-sea polymetallic sulfide deposits (DPS). Studies in this direction have stimulated both direct methods of immediate search of structures associated with metalliferous occurrences (submersible observations, sea-floor photosurveying, acoustic sounding, portrayals of structures in geophysical fields) and indirect methods based on analysis of deep deposition conditions, facilitating establishment of some regularities of its location. Unfortunately, the available geological-geophysical information is insufficient for revealing the regularities in location of the ore-bearing hydrothermal systems and their association with one or another sea-floor structure in the axial MOR zone. What is more, it is still difficult to substantiate the occurrence time and duration of action of hydrothermal sources. A knowledge of these regular features would be useful in working out the strategy of expensive surveys on DPS in the axial MOR zones. For these reasons it has become important to simulate (theoretically and experimentally) the basic deep processes in the rift MOR zones starting from complex geological and geophysical information on the structure and evolution of the region.

In the present paper we consider a numerical model of formation and further development of axial magma chambers and relation of these processes with distribution of the hydrothermal activity zones.

Prior to presentation of our modeling, we shall touch on the main conclusions on structures and evolution of the axial magma sources following from the available geological and geophysical data.

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Geological-Geophysical Information on the Existence of a Magma Chamber

Even the first studies of rift MOR zones have revealed a great difference in spreading rates, i.e., in the growth rates of new crust along accreting plate margin. These rates vary from 1–4 cm yr^{-1} on slowly divergent MOR such as the Mid-Atlantic, Hakkel and others, to 8–16 cm yr^{-1}, i.e., within the East Pacific Rise (EPR). Topographic features of the axial ridge zones correlate with the variations of spreading rate [*Dubinin et al*, 1992; *Macdonald*, 1982]. Deep graben with numerous faults and terraces along the flanks is characteristic of the slowly spreading rift zones. The outlined terraces are located 1.5 to 2 km over the graben floor. The relief of the rift zone on the fast diverging ridges represents an arched rise of 2 to 6 km width complicated by often discontinuous top graben of 1–1.5 km width, with wall elevation of only tenths of meters with respect to the bottom [*Lonsdale,* 1989; *Macdonald et al.,* 1984]. Within the rift MOR zones are identifiable neovolcanic areas associated with the recent volcanic extrusions of great bulk. They occupy the central part of the rift zones, and their width decreases from 5–8 km for slowly divergent ridges [*Ballard and Andel*, 1977; *Zonenshein et al.*, 1989] to 1–3 km for steeply divergent ridges [*Macdonald*, 1982; *Macdonald and Fox*,1983] as the spreading rate increases. The majority of active hydrothermal outcrops and attendant hydrothermal structures was found within the neovolcanic zones [*Gente et al.*, 1988; *Haymon et al.*, 1991; *Hekinian et al.*, 1985; *Karson and Brown*, 1988; *Krasnov et al.*, 1988; *Lisitsyn et al.*, 1989]. Outward from the rift axis, the neovolcanic zones give way to the crustal tectonic fracturing zones of 1–2 km width, and then to faulting zones [*Ballard and Andel*, 1977; *Ballard et al.*, 1981; *Caress et al.*, 1992; *Edwards et al.*, 1991; *Lonsdale*, 1978; *Macdonald and Luyendyk*, 1977; *Searle and Laughton*, 1977; *Ushakov et al.*, 1979]. The joint systems of these zones serve the feeders through which cold sea water gains access to deep horizons in the vicinity of the intracrustal magma chamber. Then heated water reaching the neovolcanic area through the deep micro- and macrojoints extrudes, forming numerous hydrothermal outcrops.

Analysis of available geophysical information leads to the conclusion that the distribution of the resulting hydrothermal fields is closely related to shape evolution, occurrence depth and melt-saturation of the axial magma chamber in the crust. Compelling evidence for the existence of low-velocity layers associated with the crustal magma reservoirs was obtained only on the rapidly and moderately divergent ridges. These layers were recognized from the specialized seismic experiments ROSE [*Ewing and Meyer*, 1982; *Lewis and Garmany*, 1982] and MAGMA [*Burnett et al.*, 1988, 1989; *Caress et al.*, 1992; *Harding et al.*, 1989; *McClain et al.*, 1985;] in rift zone of the ERP at 12–13°, as well as from numerous seismic studies at 9–10° N in the ERP [*Herron et al.*, 1980; *Kent et al.*, 1990; *Toomey et al.*, 1990; *Vera et al.*, 1990], in the Juan de Fuca Ridge [*Koski et al.*, 1987; *Morton et al.*, 1987; *Rohr et al.*, 1988] and other regions. Based on these data the chamber top is fixed at 1.5–2.5 km depth for the fast spreading ridges and at 2–3 km depth for the ridges at intermediate rate; the chamber width is estimated at 2–4 km.

On the whole, the geological-geophysical studies in various sites of fast or moderately divergent ridges also argue for the existence of relatively stable structures of the axial magma chamber extended over tenths of kilometers and characterized by distinct discrete-periodic magma eruptions on the surface of the neovolcanic zones [*Detrick et al.*, 1987; *Sinton and Detrick*, 1992]. Modern geophysical instruments do not allow tracing crustal magma chamber on slowly divergent ridges with spreading rates smaller than 3.5 cm yr^{-1} (i.e., the rift MOR zone at latitudes 23° N [*Purdy and Detrick,* 1986], 37°N [*Fowler*, 1976], 45° N [*Fowler*, 1978] and other sites).

Estimates of extrusion frequency based on determining the degree of freshness of lava flows and thickness of thin depositional covers show that substantial lava extrusions in the axial zone of rapidly divergent ridges occur, on the average, every 100–1000 years [*Haymon et al.*, 1991; *Lonsdale*, 1978; *Macdonald*, 1982]. The volcanic phase is replaced by the phase of hydrothermal activity and further by the tectonic phase. For the slowly divergent ridges, analysis of thickness of fresh volcanics and considerations of sizes and space distributions of volcanos in the neovolcanic area demonstrate that magmatic eruptions take place here, on the average, every 5000–10,000 years [*Atwater*, 1979]. The stated periodocities of tectono-magmatic cycles in the axial MOR zones must be of crucial importance for distribution of hydrothermal activity and fields of the deep-sea polymetallic sulfides.

Results of seismic experiments conducted in the axial zones of the MOR with high and moderate rates indicate the existence of two major reflecting horizons pertaining to the boundaries of the low-velocity zone. The first of them is observed at the depth of 1–3 km. It is seen 1–4 km away fromhe axis and, as mentioned above, is associated with the occurrence of the axial magma chamber in the crust (this magma source, as a rule, is not observed on the slowly divergent ridges). The second reflection horizon, in contrast to the first, is of more stable character and is recorded at distances as far as 15 km from the ridge axis and at depths up to 4–7 km [*Barth et al.*, 1991; *Detrick et al.*, 1987; *Garmany*, 1989; *Harding et al.*, 1989; *Herron et al.*, 1980; *Rohr et al.*, 1988; *Vera et al.*, 1990]. This horizon is associated with the Moho discontinuity and, presumably, with the rise of the asthenosphere top in the axial MOR zones; the characteristic width of the rise is 20 to 30 km (the complete width) and occurrence depth of the asthenosphere top is 5 to 10 km. This conclusion also is supported by the depth to earthquake foci in the rift zones (for the most part the depths lie between 3 and 7 km) [*Huang and Solomon*, 1987; *Huang et al.*, 1986; *Sykes*, 1970], anomalously high background values of heat flow, density models of the rift zone [*Cochran*, 1979; *Hall et al.*, 1986; *Lewis*, 1981, 1982; *Madsen et al.*, 1984] and thermal models of spreading [*Galushkin and Dubinin*, 1993; *Parker and Oldenburg*, 1973; *Rosendahl et al.*, 1976; *Sleep, 1969*, 1978, 1991; *Sorokhtin,* 1973; *Suetnova*, 1991; *Wilson et al.*, 1988].

Thermal Model of Magma Chamber

Thus, geophysical data support the evolution of the axial crustal magma sources (either steady-state as at rapidly and moderate spreading, or episodic as at low-rate spreading) against the background of comparatively stable rise of the asthenosphere top (Figure 1).

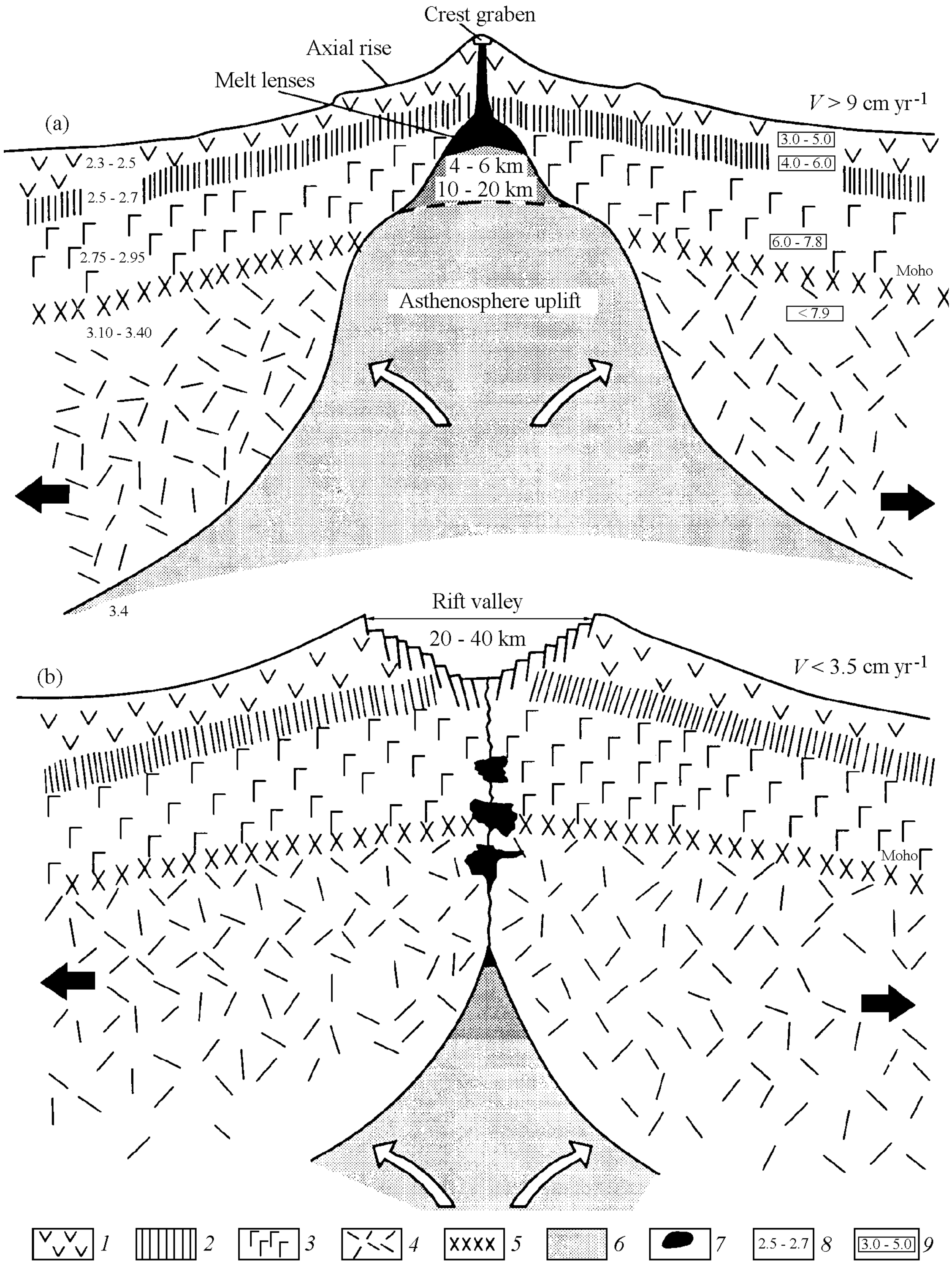

Figure 1. Schema of the deep lithosphere structure in axial MOR zones at (a) high spreading rate, and (b) low spreading rate; 1) volcanics, 2) dike complex, 3) gabbro and megagabbro, 4) upper-mantle ultramafic rocks, 5) interlayering of mafic and ultramafic rocks, serpentinites, 6) asthenosphere at various density inversion, 7) melt springs, 8) density of rocks, g cm^{-3}, and 9) seismic velocities, km s^{-1}.

The prolonged lifespan of such a rise with episodic intrusions justifies development of a simple thermal model designed for numerical analysis of processes of formation of the axial magma chamber. Our model takes into account the influence of the processes like crust accretion, circulation of hydrothermal liquids and movements of basaltic melts on the thermal state of ridge axial zone by help of displacement of the crust thermal field for the intrusion periods, effective thermal conductivity for hydrothermal heat exchange and episodic redetermination of temperature within the presumptive region of lens melt accumulation at the top of the chamber. Despite the assumptions, this approach made it possible to effectively analyze the temporal and spatial aspects of formation and evolution of crustal magma sources without solving an enormous system of heat-mass-transfer equations. According to the model, the chamber is formed in the uppermost lithosphere layer including the crust. This layer is bounded on its underside by the horizontal top of the lithosphere at the temperature $T = 1200°$ C. The thickness of the layer ranges between 4 and 6 km for high spreading rates ($V > 6$ cm yr^{-1}), and may increase to 8–10 km for low spreading rates ($V < 3.5$ cm yr^{-1}). The temperature definition domain is a rectangle with the base at the asthenosphere top and the upper bound at the ocean floor. The boundary conditions for temperature have the form

$$T = 0°\mathrm{C} \text{ at } z = 0,\ T = \mathrm{TM} = 1200°\mathrm{C} \text{ at } z = \mathrm{ZM}$$

The width of the rectangular domain (XM) exceeds its thickness (ZM) by a factor of 3–10, which justifies the condition of zero temperature gradient on the right-hand boundary

$$\frac{dT}{dx} = 0 \text{ at } x = \mathrm{XM}$$

The condition at the axis ($x = 0$) directly reflects our view of the axial magma chamber forming as a result of repetitive intrusions at the spreading axis. As this takes place, the thermal regime is periodically renewed. The intrusion repetition is supported by continuous tension state characteristic of the lithosphere in the axial rift zone. Each of these intrusions is associated with the crust growth by the value of $2\Delta x = 2V\Delta t$, where V is the average half-rate of spreading and Δt is the time interval between sequential intrusions. The duration of the intrusion itself is much smaller than the Δt interval between the intrusions, and therefore we consider the intrusion in our model as a momentary process. Occurrence of intrusion implies that the axial fracture of the width $2\Delta x = 2V\Delta t$ is filled with magma. The intrusion process is repeated every Δt interval.

For numerical estimation of the thermal consequences of such a process, we worked out a FORTRAN program. The intrusion process in computer simulation was described by rewriting or replacing the temperature within the axial intrusion width Δx with the temperature TM, which is close to the basalt melting temperature. At the same time the temperature distribution within the entire area outside of the intrusion ($x > \Delta x$) at the moment before the intrusion was displaced along the horizon by the distance Δx. In doing so, the temperature distribution in the crust immediately after intrusion takes the form :

$$T(x, z, t_{\mathrm{int}} + 0) = T(x - \Delta x, z, t_{\mathrm{int}} - 0) \text{ at } x > \Delta x$$

$$T(x, z, t_{\mathrm{int}} + 0) = \mathrm{TM} \text{ at } 0 \le x \le \Delta x \qquad (1)$$

This temperature distribution was established each time after the next regular intrusion repeated at intervals Δt. The temperature relaxation in the intervening period between the intrusions is described by the non-stationary heat conduction equation :

$$\frac{d}{dt}(\rho C_p T) = \frac{d}{dx}\left(K\frac{dT}{dx}\right) + \frac{d}{dz}\left(K\frac{dT}{dz}\right) \qquad (2)$$

In the difference approximation for the heat conduction equation the steps Δx and Δz were increased in the geometrical progression from the minimal values at $x = 0$ and $z = 0$ to maximal ones at $x =$ XM and $z =$ ZM. The minimal step in the axes x and z was defined by the half-width of the intrusion. Therefore, for instance, if the intrusion of the 50 m width originates every 1000 years (spreading half-rate averages 5 cm yr^{-1}), then the minimal time step determined from the stability condition of the solution and estimated as $\Delta t \approx \Delta x^2/\kappa$ amounts to 75 years for the normal thermal conductivity of the crust (0.006 cal cm^{-1} s^{-1} °C^{-1}) but 50 and 25 years for the "hydrothermal" values of crustal heat conductivity increased relative to the normal values by 1.5 and 3 times, respectively. For the intrusions of the 5-m width for every 100 years ($x_{\min} = z_{\min} = 5$ m) at the same spreading rate, the minimal step in time Δt should be diminished by about 100 times. Over the interval between the intrusions when the temperature regime relaxes, the time step may be increased as compared to the minimal one, but only when 10 to 20 steps have elapsed after the intrusion. On the strength of the above, the major problem of the solution discussed here is the great time required for the computer run of the evolution variant of axial magma chamber from the time of its origination in the crust to the quasi-stationary state. Important elements of the model are an approximate treatment of the hydrothermal heat transfer in the crustal rocks of the axial rift zone of spreading. Heat exchange in the region of hydrothermal activity (neovolcanic and adjacent fracture zones) was described by introducing the efficiency coefficient of thermal conductivity into the model of crustal rocks of the given block. This coefficient is in excess of the normal thermal conductivity, so that the temperature gradient in the hydrothermal block of the model has been reproduced close to the mean value observed in the real hydrothermal block. Clearly the temperature distribution within the block of the effective heat conduction differs from the temperature inside the real hydrothermally active block, especially if the latter contains local channels for water exchange. However, approximation through the effective thermal conductivity proves to be a suitable way to describe evolution of the magma chamber, taking into account the important integral heat-exchange properties of surrounding rocks and neglecting the minor details of temperature distribution. It enables us to circumvent consideration of the cumbersome system of heat-mass transfer equations but in doing so, the origin of magma chamber remains a complicated thermal problem. Nevertheless, the effective thermal conductivity approximation is justified by recent research showing that the hydrothermal systems in the axial spreading zones are very complex, multilevel and polyphase structures, whereas we just deal with the integral information on their activity reflected in the heat flow distribution on the sea floor. A particular form of the effective thermal conductivity variation with depth and distance from spreading axis has a substantial effect on the formation of the chamber. Therefore, below we begin our discussion of the modeling results with just this question.

Effective Thermal Conductivity Distribution and Shape of Magma Chamber

One would reasonably expect that an increase in the effective thermal conductivity leads to an increase in heat losses in the crust and, as a consequence, to a deeper top of the chamber. Specific computations conducted for a simplified model with the homogeneous thermal conductivity of rocks provide support for this expectation. An example is given by the case with spreading half-rate of 5 cm yr^{-1} (intrusion of $\Delta x = 5$ m for every 100 years) with normal thermal conductivity of crust ($K = 0.006$ cal cm^{-1} s^{-1} °C^{-1}) resulting in the chamber top at about 1 km depth after 13,000 years from the intrusion occurrence; but at $K = 0.06$ cal cm^{-1} s^{-1} °C^{-1} the chamber top arises at the 4.8 km depth after the same time. This is also supported by computations with variable values of the thermal conductivity (Figure 2).

Variation of the thermal conductivity with depth and with distance from spreading axis substantially affects the shape and location of the magma chamber top. This situation is illustrated in Figure 2. All curves represented here correspond to the model of episodic spreading with intrusions of the 50-m width every 1000 years; the results are given for the time $t = 160{,}000$ yr. The appropriate thermal conductivity distributions $K(x)$ are shown in the right upper corner of Figure 2. It is characteristic of all distributions in Figure 2 that the chamber top deepens very slowly with distance from the axis, so that the chamber half width becomes greater than 4 km. This is, however, inconsistent with the seismic data, which limit the half width to the values 1.5 to 2.5 km, even for the rapidly divergent ridges. Such a requirement is satisfied best by the shape of chamber top corresponding to the curve (e) in Figure 2. This shape is compatible with the complex thermal conductivity distribution shown in the upper part of Figure 3 and described as follows

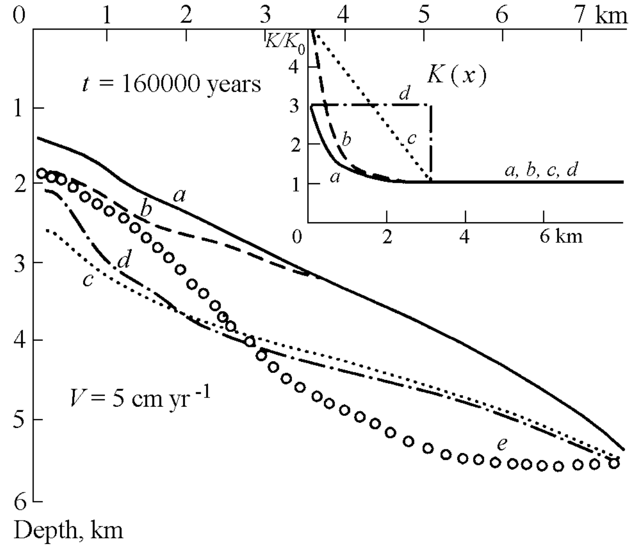

Figure 2. Location of chamber top at various thermal conductivity $K(x)$ distributions (the upper right corner). See text.

$$K(x,z,t)/K_0 = \begin{cases} 1 + 4e^{-2x} & 0 < x < 1 \\ 1.5 & 1 < x < 2 \\ 3 & 2 < x < 4.5 \\ 7.5 - x & 4.5 < x < 6.5 \\ 1 & x > 6.5 \\ 1.5 & T > T_s(\text{chamb}) \end{cases} \qquad (3)$$

In distribution (3) we indicate the central interval $x < 0.5$ km with high hydrothermal heat transfer ($K/K_0 > 3$), domain of moderate hydrothermal activity within the neovolcanic zone ($K/K_0 \geq 1.5$), fracturing zone

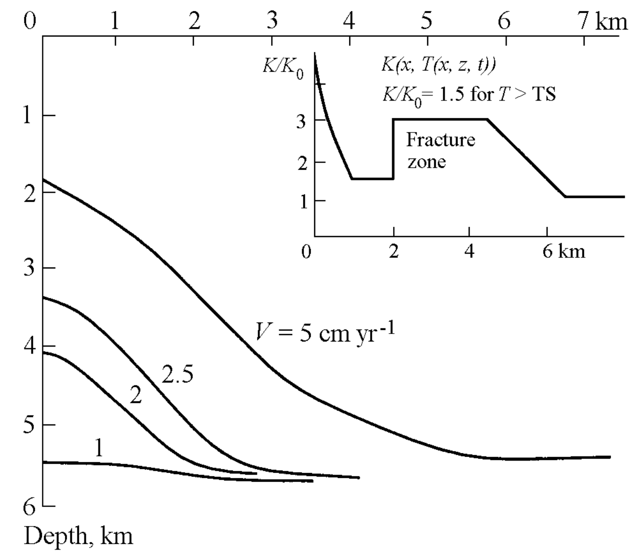

Figure 3. Change in chamber shape with spreading rate. Intrusions are 5 m every 100 years ($1/2V = 5$ cm yr^{-1}), 50 m/2000 yr ($1/2V = 2.5$ cm yr^{-1}), 50 m/2500 yr ($1/2V = 2$ cm yr^{-1}) and 50 m/5000 yr ($1/2V = 1$ cm yr^{-1}). The thermal conductivity distribution is shown in the upper right corner.

$2 < x < 4.5$ with high contribution of hydrothermal transport ($K/K_0 = 3$), and the range $4.5 < x < 6.5$ with a gradual drop in hydrothermal activity and reconstruction of the normal thermal conductivity. Furthermore, the distribution $K(x, z, t)$ from (3) accounts for the fractures not being able to propagate deep into the magma chamber and, therefore, the hydrothermal region must be bounded in depth, at least, by the basalt solidus isotherm. The effective thermal conductivity of rocks inside the chamber is difficult to determine. We assumed that it is 1.5 times greater than the normal thermal conductivity. In fact, while the temperature in the chamber is kept high $T_s < T < 1205°$C, being sufficient for basalt partial melting, the degree of this melting cannot ensure the predominance of convective heat transfer. Thus, the effective thermal conductivity increases insignificantly here.

Effect of Spreading Rate on the Formation and Evolution of Magma Chamber

The effective thermal conductivity (3) can be considered as one of the probable distributions $K(x, z)$ reflecting, in a very general form, the intensity of hydrothermal heat exchange in the crust (variation of intensity with depth and distance from spreading axis). We used this distribution in the next series of computations designed for analyzing the important problem of the shape and sizes of the axial magma chamber under the influence of the spreading rate. The modeling results are presented in Figure 3. They show that a decrease in the spreading half-rate by two times (from 5 to 2.5 cm yr^{-1}) results in the chamber dipping by 1.5 km and a decrease in the half width of the chamber by 1 to 1.2 km. At the half-rate of $V = 1$ cm yr^{-1}, the stationary axial chamber exists as a 5.5-km-deep rise no more than 300 m high, and cannot be distinguished by seismic methods. The half-rate $V = 1.5$ cm yr^{-1} can be accepted for the lower bound of the crustal chambers distinguishable with geophysical methods. In this case the chamber has an elevation of not more than 0.5 km relative to its walls, and the half width of the chamber amounts to 0.5–1 km.

Evolution of Magma Chamber in the Regime of Its Formation and Cooling

Note that, strictly speaking, our formulation of the problem does not lead to the stationary or asymptotic shape of the magma chamber. The chamber shape continuously varies but as time passes these variations remain prominent only at the very far flanks. Therefore, we can introduce the notion of the asymptotic or stationary chamber shape, meaning a shape established during a time on the order of 0.2–0.3 mil yr, which is comparable to the geological lifetime of similar structures (0.5–1 mil yr). The modeling results shown in Figure 4 illustrate dynamics of the process in which the chamber shape approaches the "stationary" one. Here

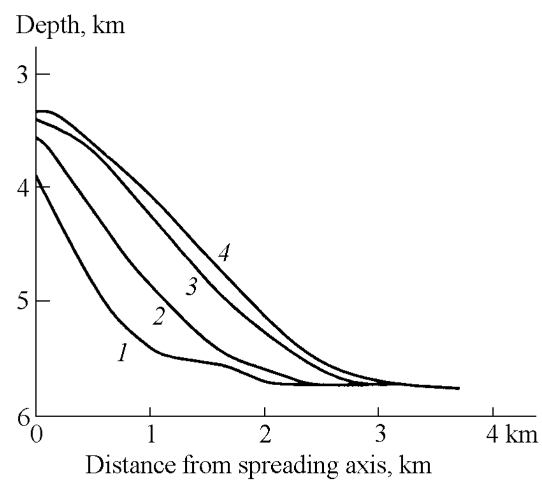

Figure 4. Variation of chamber shape from beginning of the process to the stationary state. $V = 2.5$ cm yr^{-1}, intrusion at 50 m/2000 yr. (1) 35,000 yr, (2) 60,000 yr, (3) 140,000 yr, and (4) 350,000 yr (thermal conductivity from (3), see text).

we give an example at the half-rate of opening $V = 2.5$ cm yr^{-1}. However, this is characteristic of the all rates considered. In all cases the chamber shape, except for the far-away flanks, differs from the asymptotic form by no more than 5% even within 100–150 ths yr after the intrusion time, i.e., after 40–60 magma intrusions at the spreading axis.

Figure 5 demonstrates the dynamics of the process which is inverse to one considered above, that is, the relaxation of thermal state of the cooling magma chamber. As regards the chamber shape prior to the cooling, we obtain it by considering intrusions of half width 50 m every 1000 years ($V = 5$ cm yr^{-1}) over 320,000 years. As Figure 5 suggests, a marked change in the chamber shape (a two-fold increase in the half width) happens, on the average, each 20–30 thousand years after the inception of cooling. Intrusion interruption of 100 mil yr or more results in the disappearance of the magma chamber. Understanding the relaxing thermal regime in the chamber is critically important for analysis of hydrothermal evolution in terminating the related arm of axial spreading zone, i.e., in jumping the spreading axis. It is conceivable that other causes are also responsible for significant deviations from periodicity of dike intrusions when the thermal state of the source has an opportunity to relax noticeably. These final stages of hydrothermal activity are of particular interest for the origin of sulfide deposits, because in such cases the deposits do not overlap with the further extrusions of lava flow and very likely are preserved on the oceanic floor. It should be emphasized particularly that in our modeling we assumed that the intensity of hydrothermal activity did not vary during the cooling of the magma source. This assumption is not quite correct for the time represented in Figure 5. Therefore, one can expect that the cooling beginning from the time $t \approx 30{,}000$ yr slows down in comparison with the results in Figure 5. The next development of the model takes into consideration a reduction in the hydrothermal activity together with dipping of the lower boundary of hydrothermal water in the crust, as the chamber cools.

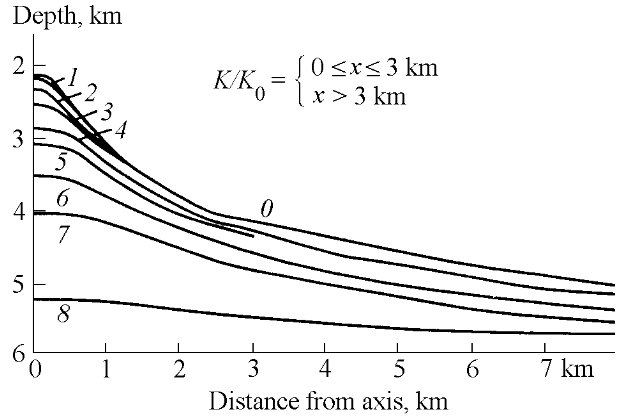

Figure 5. Chamber cooling. The initial shape (curve 0) was found at episodic 50-m intrusion for every 100 years over 320,000 years. The cooling times (thous. years) are (1) 1.2, (2) 6.6, (3) 10.4, (4) 20.4, (5) 30.4, (6) 50.4, (7) 80.4 and (8) 180.4.

Effect of Intrusion Parameters on Shape and Evolution of the Chamber

To summarize the review of our modeling, we shall touch on the method itself. The examples in Figure 2 refer to the spreading half-rate of 5 cm yr^{-1}. This kind of process can be simulated by intrusions of the 50-m width every 1000 years, as well as of the 25-m width every 500 years; some other possibilities exist as well. Geophysical and geological data give no way of choosing the correct variant. Therefore, analysis of the above solutions brings up the natural question: How much do the results obtained depend on the accepted values of intrusion frequency? To examine this problem we made a series of comparative computations: a) for the spreading half-rate $V = 5$ cm yr^{-1} with intrusions of 5-m width every 100 years (5 m/100 yr), 25 m/500 yr, and 50 m/1000 yr; b) for $V = 2.5$ cm yr^{-1} with intrusions of 25 m/1000 yr and 50 m/2000 yr; and c) for $V = 1.0$ cm yr^{-1} with intrusions of 50 m/5000 yr and 100 m/10,000 yr. The calculations showed that after a lapse of time from the beginning of the process giving more than 20 cycles of intrusions at maximal width (from the above variants), the chamber top ceases to depend on the width and related frequency of intrusion with accuracy up to 100–200 m.

Conclusion

In spite of the definite simplifications in our formulation of the problem and methods of its solution, the represented model provides, to a first approximation, a rough idea of the space-time scaling of the shape and thermal regime of the axial magma chamber. This, in turn, allowed us to estimate the intensity of hydrothermal activity involved in the formation and relaxation processes in the magma chamber. Analysis of the modeling results leads to the following conclusions.

1) Formation of the axial magma chamber is substantially related to discrete character of intrusions and lava extrusions under the conditions of continuous tension of the lithosphere in the axial spreading zones.

2) The magma chamber shape and occurrence depth of its top are closely connected with the intensity of hydrothermal activity in the crust and with regular trends in hydrothermal variations outward from the axis.

3) The frequency of intrusions as well as spreading half-rate and hydrothermal heat exchange in the crust are of the utmost significance for the occurrence and existence of a stable crustal magma chamber and the evolution of its shape. In particular, the existence of a stable chamber is unlikely if the frequency of intrusions is small and the related spreading half-rate is lower than 1.5 cm yr^{-1}. The shape of the magma chamber varies with a change in the spreading rate (frequency of intrusions).

4) Interruption between intrusions covering 100 thousand years and more leads to closing of the magma chamber.

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