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FLUID CONVECTION IN SEDIMENT-HOSTED RESERVOIRS DUE TO THERMAL ACTION OF DIKES AND SILLS

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Flow regimes of pore fluid were studied, with the steam-water phase transition near cooling intrusions of dikes and sills in a sedimentary basin taken into account. The study was based on a computer program simulating a one- or two-phase flow of the steam-water mixture depending on the phase state of the fluid. An actual section of the Yenisei-Khatanga basin northwest of the Siberian Platform is analyzed. Two types of models were used: (1) intrusion of a sill into the basement of the basin, beneath a reservoir bed, and between two reservoirs; (2) the same situations but with a vertical dike and an off-branching sill. Simulations were carried out to study the effect exerted on convection by pore fluid of a single sill or a dike combined with sill that intrude into the sediment. The results of the calculations permit prediction of the beginning of convection depending on the type of intrusion, its location in the section of sedimentary basin, initial temperature, and physical parameters of rocks. Patterns of evolution of temperature field and velocities of fluid flows in the sediment around cooling intrusions have been obtained. The problem is important for predicting the behavior of both water and hydrocarbon fluids in basins where trap magmatism is expressed.

Fluid flow, convection, sedimentary basin, numerical modeling

INTRODUCTION

The effect of basalt (trap) intrusions on the convective flow and phase state of water and hydrocarbon fluids in the host rocks remains to be studied. This problem is of great practical interest, especially with regard to the sedimentary basins which experienced basic magmatism. Examples include sedimentary basins of Central and Arctic Siberia, where recurrent effusions of basaltic magma and its intrusions into the sedimentary cover occurred in the Permian to Early Triassic. Most likely, the magma intruded permeable sedimentary rocks under shallow conditions, the pore pressure being close to hydrostatic at these depths [1]. The magmatic heat disturbs the heat balance and leads to free-convection flows. The geology of these basins implies that the convection flows of the fluid could have played an important role in the mass and heat transfer.

Since the 1970s, considerable progress has been made in developing mathematical models for free and forced convection of a water fluid in a porous medium. In general, the studies were concerned with hydrothermal systems or processes of cooling around magmatic intrusions. First models were restricted to single-phase conditions: Norton and Knight [2] ignored the two-phase region, and Cathles [3] supposed that the fluid at each point was either vapor or liquid. Schubert and Straus [4] estimated the effect that thermophysical properties of fluid such as viscosity, compressibility, heat expansion etc. exert on heat convection. Critical Rayleigh numbers for the convection onset were obtained as functions of the thickness of the porous bed, geothermal gradient, and temperature difference across the bed. Also, attempts were made to describe the convection of a two-phase fluid in a porous medium on the basis of simplified models [5]. The steam-water mixture is simulated in this work by a homogeneous model, i.e., when the only Darcy velocity is responsible for the mass flow of the mixture, the thermodynamic and transport properties of the mixture depend only on relative amounts of the phases.

A better understanding of the physics of fluid convection in the Earth's crust significantly advanced the frontiers of the use of relevant models in hydrogeology [6, 7], geology of sedimentary formations [8], theory of ore formation [9], metamorphic petrology [10], oil geology [11, 12], and in other fields of geosciences. The state-of-the-art theory of fluid flow is described in many monographs and reviews [13]. The theory of reaction flows in porous media is now a basic tool for simulating mineral formation, phase transitions, and fluid-rock interaction.

The studies of the last decade emphasize the interaction of a fluid with the porous medium [14–16]. The subject was the different rheology of the solid phase: elastic, viscous, elasto-viscous, and compacted with varying viscosity. The filtering medium was, as a rule, a one-velocity homogeneous liquid with constant or variable properties. In the Balashov and Yardley model [17], the interaction of fluid and the rock matrix under strain is complicated by solid-phase decarbonation so that the permeability changes. The issue under consideration was the one-dimensional problem of filtration under isothermal conditions and at a constant pressure gradient with the water-hydrocarbon fluid of variable composition. Thus, in the majority of the earlier works the flowing substance was a homogeneous fluid without phase transitions. Our approach principally differs from the earlier approaches: We suppose that the movable medium is able to pass from one phase state to another, its properties being radically changed.

To decribe the effect of trap magmatism on hydrocarbons, several earlier works suggested an analytical model for thermal alteration of rocks which contain dissipated organic matter (OM) near parallel sills. Dynamics of the temperature field in the interintrusive space has been examined. A remarkable inference is that the conditions of OM transformation near the basalt sills are favorable for gas HC rather than for liquid HC [18, 19]. The extent of thermal "destruction" of oil pools were studied in [20] in the framework of conductive heat transfer, and the heat conductivity of rocks was estimated as "effective", with a fluid convection in a porous medium taken into account. It was shown that the magmatic heat exerted a great influence on the processes of generation of hydrocarbons dissipated over the volume of the sedimentary basin. Dikes and sills of basic magma could led to degradation of kerogen and formation of gas and liquid oil fractions. The magma-supplied heat could participate in the generation of hydrocarbons at the cost of transformation of the OM dissipated in sedimentary rocks as well as in thermal destruction of oil and its degassing. How completely and quickly the oil will be destructed depends on the rate at which the products of reaction are removed from the zone of contact with basalts. Given a considerable temperature gradient in the near-intrusion rocks and a sufficiently low viscosity of heated oil, convection provides the removal of gases and light HC, whereas heavy solid and semi-solid products such as cox, tar, and so like remain in place. Driven off from the contact with basalts, gases and light HC are able to react with oil components at a distance from the intrusion. Ultimately, this process leads to evolution of composition of liquid HC toward deasphaltization and increased proportion of light products in the deposit. Its intensity depends on particular ratios of thicknesses of the intrusion and oil reservoir [21].

MATHEMATICAL MODEL AND FORMULATION OF THE PROBLEM

Numerous works deal with solidification and crystallization of dikes and sills [19, 22, 23]. Rather than consider the immediate problems of magmatism, we investigated the processes occurring chiefly outside the intrusive bodies, considered as a thermal source. Formulation of the problem of filtration and convection of a fluid in a porous medium of sedimentary basin involves some specific issues: (1) shallow conditions, which are responsible for the hydrostatic initial distribution of pore pressure, (2) nonuniform structure of the region, composed of alternating beds of fluid seals and reservoirs, and (3) higher values of porosity and permeability as compared with a deeper geologic setting.

The method of solving the problem with a pure water fluid. The problem of modeling the convective flows of a water fluid is solved in a 2D formulation on the basis of nonstationary equations of nonisothermal hydrodynamics. The main assumptions are as follows: Darcy relationship holds true for a two-phase current; effects of capillary pressure are neglected; all the phases (matrix, i.e., solid rock, liquid, and gas) are considered being in state of local thermal equilibrium; reaction of solid rock matrix to the fluid's motion is negligible.

The system of equations is as follows. The equation of energy conservation is used in the form [24]

$$\rho \frac{DH}{Dt} = \mathrm{div}\,(K_\mathrm{m}\,(T)\nabla T) + \frac{DP}{Dt}, \tag{1}$$

where H is enthalpy, T — temperature, P — pressure, $D/Dt = \partial/\partial t + u\nabla$ is the operator of material derivative, ρ — density, $K_\mathrm{m}(T)$ — heat conductivity of the medium. If the liquid moves at a velocity much lower than the velocity of sound, the movement-induced changes in pressure are so low that can be neglected and the last term in the right-hand part of (1) can be discarded [25]. Unlike the universally admitted P-T, the variables P-H unambiguously define the thermodynamic state of the system in both one- and two-phase states. Writing expressions for the density of mixture, complete enthalpy, and volume saturations of phases in the fluid

$$\rho = S_w \rho_w + S_s \rho_s, \tag{2}$$

$$H = (S_w \rho_w H_w + S_s \rho_s H + (1 - n)\rho_r H_r)/\rho, \tag{3}$$

$$S_w + S_s = 1 \tag{4}$$

and, introducing corresponding quantities into the differentiation term, we obtain the relationship of heat balance on convection of two-phase fluid through the porous medium

$$\frac{\partial[nS_w \rho_w H_w + nS_s \rho_s H_s + (1 - n)\rho_r H_r]}{\partial t} - \nabla[\rho_s H_s \cdot V_s] - \nabla[\rho_w H_w \cdot V_w] - \nabla(K_m \nabla T) = 0, \tag{5}$$

where n is porosity, $S_{w(s)}$ is volume water (steam)-saturation, k is absolute permeability, $k_{rw(s)}$ is relative (phase) permeability of water (steam), $\mu_{w(s)}(P, T)$ and $\rho_{w(s)}(P, T)$ are dynamic viscosity and density of water (steam), H is enthalpy, K_m is effective heat conductivity of porous medium, indices w, s, r stand for water, steam, and rock, respectively.

The equation of fluid motion is the Darcy approximation for a two-phase flow in a porous medium, valid for slow inertion-free currents:

$$V_s = -\frac{kk_{rs}}{\mu_s}(\nabla P - \rho_s g \nabla Z), \tag{6}$$

$$V_w = -\frac{kk_{rw}}{\mu_w}(\nabla P - \rho_w g \nabla Z),$$

where $\nabla z = \{0; 1\}$ is the unit vector along the Z-axis, g is the acceleration due to gravity. This approximation is valid when the Reynolds number $\mathrm{Re} = Vd/\nu$, calculated from mean pore diameter, falls between 1 and 10 [8, 26]. At maximum velocities of flow in all versions of the model (see below) $V = 10^{-6}$ m/s, size of pores $d = 10^{-3}$ m, and minimum viscosity of a fluid in the supercritical domain $\nu = 10^{-7}$ m^2/s the Reynolds number is Re = 0.01, which permits us to use (6) as equations of motion.

The equation of mass conservation is formulated in terms of density of a two-phase medium:

$$\frac{\partial\,[n\,(S_w \rho_w + S_s \rho_s)]}{\partial t} - \nabla[\rho_s V_s] - \nabla[\rho_w V_w] = 0. \tag{7}$$

The equation of fluid state describing the relationship between thermodynamic variables of pressure, temperature, and density in the form $P(T, \rho)$ closes the system of equations. We used the Haar-Gallagher-Kell equation of state to describe properties of pure water fluid in the interval of temperatures of 0 to 1200 °C and pressures of 1 to 10,000 bars [27]:

$$P = P_{base} + P_{resid},$$

$$P_{base} = \rho RT \left\{\frac{3 + 33y + 133y^2}{3(1 - y)^3} + B_{\nu,2}\,\rho - 14y\right\},$$

$$P_{resid} = \frac{\rho^2 e^{-\rho}}{\hat{C}C} \sum_{i=1}^{36} g_i \left(\frac{T_c}{T}\right)^{l_i} (1 - e^{-\rho})^{k_i - 1} +$$

$$\rho^2 \sum_{i=37}^{40} g_i\,\rho_i^{-1} \exp(-\alpha_i\,\delta_i^{k_i} - \beta_i\,\tau_i^2)\,\{l_i\,\delta_i^{l_i - 1} - \alpha_i\,k_i\,\delta_i^{k_i + l_i - 1}\},$$

$$y = b_{\nu,2}\,\rho/4, \tag{8}$$

$$B_{\nu,2} = \sum_{j = 0, 1, 2, 4} B_j \left(\frac{T_c}{T}\right)^j,$$

$$b_{\text{v},2} = b_1 \ln (T_c/T) + \sum_{j=0,3,5} b_j \left(\frac{T_c}{T}\right)^j,$$

$$\delta_i = (\rho - \rho_i)/\rho_i, \;\; \tau_i = (T - T_i)/T_i,$$

where g_i, k_i, ρ_i, T_i, α_i, β_i, δ_i, B_i, b_i are empirical approximating coefficents; C and $\hat{C}$ are converted dimension constants whose values are given in [27]. Equation of state (8) takes into account changes of thermodynamic properties near the critical point of water (critical parameters: P_c = 22.055 MPa, H_c = 2086.0 kJ/kg, T_c = 373.98 °C, $\rho_c = 3.94 \cdot 10^{-2}$ MPa·s) as well as on curves of phase transitions (condensation and evaporation). Thermodynamic parameters of fluid in the subcritical domain are calculated in advance from (8) with the help of interpolation bicubic splines and are kept in the memory as direct-access files.

Thermophysical coefficients depend on the type of rocks and are supposed to be as follows:

— heat capacity and heat conductivity of water-saturated rocks and intrusion were considered equal, with c_p = 1 kJ/(kg·K) and K = 2 W/(m·K) (they can be specified as temperature-dependent in the form of linear, step, or delta-function);

— relative phase permeabilities were given as linear functions of water- and steam-saturation [28]:

$$k_{\text{rw}} = \frac{S_{\text{w}} - S_{\text{wr}}}{1 - S_{\text{wr}} - S_{\text{sr}}}, \tag{9}$$

$$k_{rs} = 1 - k_{\text{rw}},$$

where S_{wr} and S_{sr} are residual or critical water- and steam-saturation in the fluid below which water (steam) becomes an immobile phase; our calculations use S_{wr} = 0.3 and S_{sr} = 0; porosity and permeability are characterized by the type of rocks (i.e., are functions of spatial coordinates) with n = 0.01; 0.05; 0.25 and $k = 10^{-18}$, 10^{-16}, 10^{-14} m^2 for intrusions, country sedimentary rocks, and reservoirs, respectively (Fig. 1).

The nonstationary boundary-value problem with initial data is to find a P-T field and a field of phase velocities that satisfy (2)–(9), the boundary and initial conditions within a given region. The boundary conditions specify

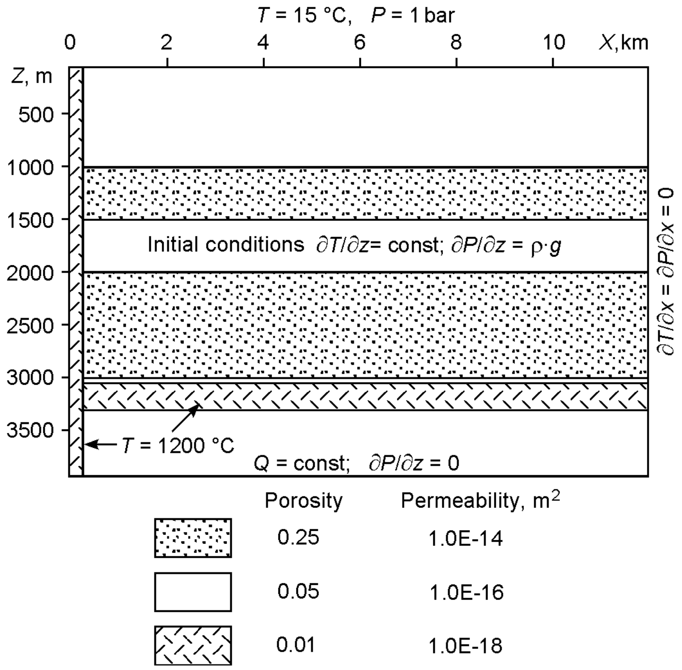

Fig. 1. Geometry of calculated region, model parameters, boundary and initial conditions of the problem of pore fluid filtration in a nonuniformly layered medium composed of seals (light areas), reservoirs (speckled), and magmatic intrusions (hatched). Symbols and values of porosity and permeability for each kind of rocks are given at the bottom.

constant pressure (P = 1 bar) and temperature (T = 15 °C) on the top surface (conditions of free surface), zero horizontal gradients of pressure and temperature on the sides of the region (conditions of symmetry), constant heat flow, and no vertical gradient of pressure (condition of impermeability) on the bottom (see Fig. 1). The initial conditions require the definition for distribution of pressures and temperatures within the region. It is supposed that at t = 0 the fluid experiences hydrostatic pressure and temperature corresponds to the geothermal gradient 25 °C/km everywhere but the regions in which basic intrusions at an initial temperature of 1200 °C were emplaced.

Equations (2)–(9) were numerically solved on the basis of the finite-difference approach from [29]. The computation algorithm is based on an implicit finite-difference scheme with iterations according to the Newton-Raphson method [30]. To solve a linearized matrix equation, an algorithm of successive upper relaxation built into the Newton-Raphson iteration [30] was used. The program block calculating thermodynamic and transport parameters of the fluid at each step retrieves the precalculated and stored data for the fluid temperatures ranging from 0 to 1200 °C and pressures from 1 to 10 kbar. Equation of state (8) defines whether a given point of space contains a compressed liquid (S_w = 1), a mixture of steam and water ($0 < S_w < 1$), or overheated steam (S_w = 0). Above the critical point the differences in properties between liquid- and steam-disappear and, therefore, the value of water- and vapor-saturation in (5), (7), and (9) becomes uncertain. For the calculation, the properties in the supercritical region are considered such as if the system would contain two phases with identical properties, and the saturation gradually decreased from S_w = 1 (liquid) on the curve of evaporation to S_w = 0 (vapor) on the curve of condensation, radially around the critical point. In the absence of vapor or liquid, the saturation $S_{w(s)}$ of the missing phase is zero, and that of the available phase, unity. It is also supposed that the relative permeability $k_{rw(s)}$ of the missing phase is zero, and that of the available phase, unity. Thus, equations (2)–(9) are reduced to the corresponding expressions either for the region of compressed liquid, or for the region of overheated vapor, or for the two-phase region.

Equations (2)–(9) describe the two-phase convective flow of heat and mass in the steam-water-rock system. However, having replaced the state-equation module, we can use them to describe the gas–oil–oil-producing rock system.

RESULTS OF MODELING OF THE EFFECT OF INTRUSIONS OF SILLS AND DIKES ON CONVECTION IN A POROUS MEDIUM

We consider two types of models — basic magmatism in the form of a single horizontal sill and the case of intrusion of a vertical dike, off which a horizontal sill branches. The main difference of this model from the previous ones is that the heating laterally and from below is not constant but proceeds in a spike: At first, "instant" intrusion occurs resulting in a sill or a dike, and then they gradually cool. It is supposed that a dry basalt melt initially heated to 1200 °C is injected by "active" intrusion [9]. Crystallization accompanying magma solidification and release of volatiles are beyond consideration.

The target of modeling is a 12×4 km area near the side of a sedimentary basin, heterogeneous in structure of sediment fill (Fig. 1). The left side boundary is the axis of symmetry of the intruded dike and the right boundary is drawn far enough to minimize its effect. (The thickness-to-length ratio for the lower sediment-hosted reservoir bed is 1:12, and for the upper reservoir, 1:24) The upper and lower reservoirs are set 500 m and 100 m thick, respectively. The thickness of the upper cap rock is 1000 m, and the thickness of weakly permeable bed between the reservoirs is 500 m. This structure of the sill-sedimentary complex is a simplified model for the structure of the upper part of the Ust'-Yenisei depression, the western structure of the Yenisei-Khatanga rift basin [32] and northern continuation of the Tunguska sedimentary basin [33]. Values of porosity, permeability, and thickness correspond to measurements in boreholes [1].

The structure of the sedimentary section is a sequence of Mesozoic deposits of the Ust'-Yenisei depression (from top to bottom):

— Paleocene and Upper Cretaceous deposits (to a depth of 1 km) with predominant clay and aleurite sediment;

— the Dolgan Formation (Cenomanian Stage) 500 m thick, composed chiefly of sands and sandstones with high storage and filtration properties. The most frequent values of open porosity 28–32% and permeability 500–1000 mD characterize this formation as a regional reservoir;

— the Yakovlev Formation (Albian-Aptian) 500 m thick, composed of clay-aleurite rocks, with higher contents of clays than in the under- and overlying reservoirs. According to laboratory determinations of porosity and permeability the formation is considered a regional cap;

— the Malaya Kheta and Sukhaya Duda Formations (Barremian-Valanginian Stage of the Upper Cretaceous) 1000 m thick, composed of weakly cemented sandstones and sands with predominant reservoir properties.

Table 1

Critical Values of Rayleigh Numbers and Wave Numbers at the Beginning of Convection in a Horizontal Pore Layer

Boundary condition at the lower bound	Boundary conditions at the upper bound	Ra_{cr}	a_{cr}
1.3	1.3	39.48 ($4\pi^2$)	3.14
1.3	2.3	27.10	2.33
2.3	2.3	12	0
1.3	1.4	27.10	2.33
2.3	1.4	17.65	1.75
1.3	2.4	9.87 (π^2)	1.57 ($\pi/2$)
2.3	2.4	3	0
1 or 2.4	1 or 2.4	0	0

Note. Numbers in first two columns mean the type of boundary conditions at the lower and upper bounds, respectively. For the upper and lower surfaces the boundary conditions for temperature and fluid flow are as follows: 1 — constant temperature, 2 — constant heat flow, 3 — rigid impermeable boundary (no flow), 4 — constant pressure (free surface). The case considered in this paper is framed.

Position of the dike and sills in the section of the basin is variable in models. However, their thickness and physical parameters are taken typical of the intrusive bodies of the Tunguska and Yenisei-Khatanga basins [33–35].

Model for a single horizontal sill. In the layered porous medium heated from below, the convection begins when the dimensionless Rayleigh criterion

$$Ra = \frac{\alpha g \Delta T H \rho^2 c_p k}{\mu \chi} \tag{10}$$

reaches some critical value. Here, α, μ, ρ, and c_p are the heat expansion, viscosity, density, and heat capacity of the fluid, respectively; k and χ are the permeability and heat conductivity of the fluid-saturated rock; H, ΔT are the thickness and temperature difference across the bed. In the Boussinesq approximation for a liquid with constant properties, the critical number Ra and corresponding wavenumber of the convective cell $a_{cr} = 2\pi H/\lambda$ (H, thickness of layer, λ, wavelength, or width of cell) [36] are given in Table 1.

Calculations were carried out using a model for "instant" emplacement of the sill, to study the convection in a layered medium composed of a pair of reservoirs situated above one another. To create an initial temperature perturbation, it was supposed that the temperature left of the sill is higher than in the rest of the layer. It looks as if magma was supplied from the source from left to right. The initial perturbations were given at several nodes of calculation grid left of the sill, while everywhere else the temperature was even.

The sill at the bottom of the basin. It was supposed that the sill 250 m thick was intruded into the rocks at a depth of 3750 to 4000 m, which is 750 m deeper than the bottom of the lower reservoir. Calculations show that this intrusion causes no convection either in the lower or in the overlying reservoirs (Fig. 2, *A*). The lower reservoir is heated up to 150–275 °C, and the maximum temperature difference across the layer of 1000 m is about 170 °C. Substituting given physical parameters of the layer yields the Rayleigh criterion Ra = 4.5, which does not exceed the critical value for the onset of convection in a porous layer with boundary conditions on horizontal surfaces framed in Table 1. In the absence of convection, the sill cools due to conduction (Fig. 2, *A*), unlike the case with the developed convection shown in Fig. 2, *B*, which will be considered below.

Sill contacting the lower reservoir. In this version, the effect of sill intrusion at 3.0–3.25 km was considered, i.e., at immediate contact with the lower reservoir. Figure 3 shows the convection patterns 1500, 3500, 50,000, and 100,000 years after the sill emplacement. Temperature disturbance on the left side of the sill serves a perturbation causing development of convection in the overlying reservoir. A local maximum of 1000 °C was set left of the sill, whereas everywhere else the initial temperature was 900 °C. The initial perturbation moves as a temperature wave along the strike of reservoir (from left to right) and does not travel into the neighbouring strata: the 50 °C isotherm, running within the upper reservoir, remains nearly parallel to the surface. The movement of

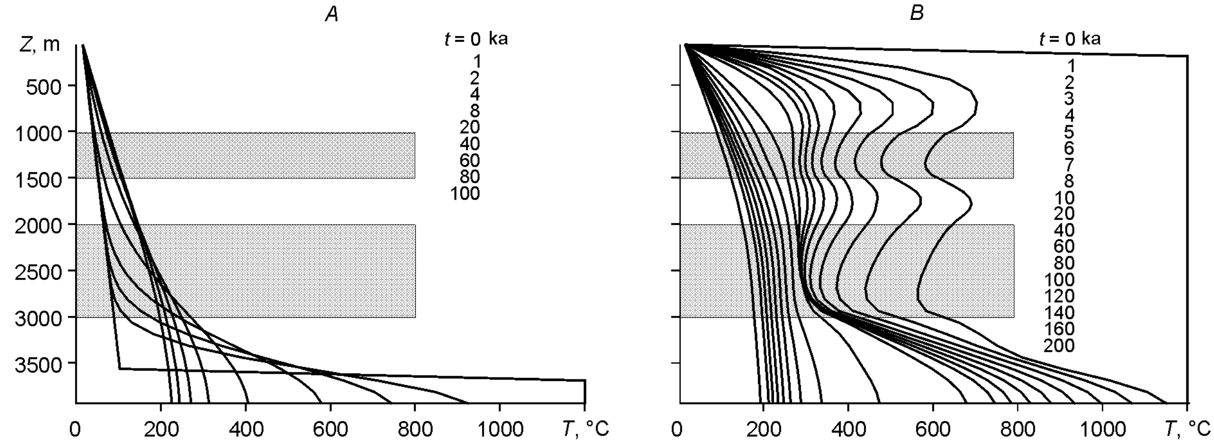

Fig. 2. Comparison of temperature profiles for model of sill cooling given no convection and developed convection in reservoirs. Geotherms are drawn for times (ka) shown in the right. *A*, evolution of temperature above a single cooling sill 250 m thick at the bottom of the basin. *B*, evolution of temperature at the contact of a dike and host rock in a model whose results are shown in Fig. 6. Vertical profiles are drawn along the contact of dike with host rock. Speckled gray is the location of reservoirs.

the convection front can be determined by the velocity of propagation of the temperature disturbance throughout the stratum from the point of initial perturbation. This velocity can be estimated from Fig. 3 at 4000 m/1500 years = 2.7 m/year. It appears to approximate the rate of fluid filtration.

Single sill between two reservoirs. Quite frequently, a sill intrudes between two reservoirs. For this case we modeled the emplacement of a 125 m thick sill with an initial temperature of 1200 °C into the middle part of the intermediate layer 500 m thick. Thus, the sill was in contact with none of the reservoirs. Figure 4, *A* shows a cooling history of a sheet intrusive body and warming up of the surrounding rocks. After about 100,000 years, the temperature profile comes back to its initial state. The temperature gradient in the upper reservoir does not exceed 150–160 °C/500 m for the whole period of sill cooling and decreases with time to the initial gradient 25 °C/km. The maximum Rayleigh number calculated from the parameters of the upper layer is 13.5. Reminiscent to the sill at the bottom of the basin (see Fig. 2, *A*) this value does not exceed the critical Rayleigh number for the onset of convection (Table 1). Hence, the temperature profiles correspond to conductive cooling.

The main result of this series of calculations is that the loss of fluid stability occurs only in the near reservoir, whereas in far layers it is not observed. If the intrusive body is emplaced at some distance from the contact with the reservoir, it is possible to calculate a ratio between the thickness of intrusion H and distance to reservoir L at which convection takes place. In particular, at $L/H = 0.5$ and 0.66 there is no convection, and the stability is lost at L/H between 0 (at the immediate contact) and 0.5. More exact estimation requires additional calculations.

Model for simultaneous emplacement of a dike and a sill into a water-saturated layered medium.

A dike and a sill between two reservoirs. In the preliminary models a single sill is "detached" from the source. It seems more likely, however, that a magma melt was intruded through a vertical fracture (feeder dike), with a horizontal sill going off it. In all cases with a dike, free convection is caused by heating simultaneously from below and laterally in the region to be modeled. The initial temperature of the dike, as well as of the sill, is set at 1200 °C. The temperature of magma in feeder dike is assumed to be the same in depth. Thus, both a horizontal and a vertical gradients of temperature are created at the moment of emplacement of intrusions. Unlike in previous models, free-convection flows form in the two reservoirs, irrespective of the location and thickness of the intruded sill. However, the dynamics and duration of the existence of cellular convection are different for different models.

Figure 5 shows the initial stage of convection 2000 and 3000 years after magma intrusion. A 250 m thick sill is intruded into an intermediate layer between reservoirs. In the upper reservoir, which is heated laterally and from below, the isotherms acquire an oscillating form and a wave of thermal perturbations propagates as in the model in Fig. 3. This does not occur in the lower reservoir. Lateral warming up leads to the formation of a single cell near the cooling wall of the dike. Owing to the active circulation of fluid, more intense cooling occurs there, and the isotherms, which are initially nearly parallel to the contact of the dike, are displaced to the left, toward

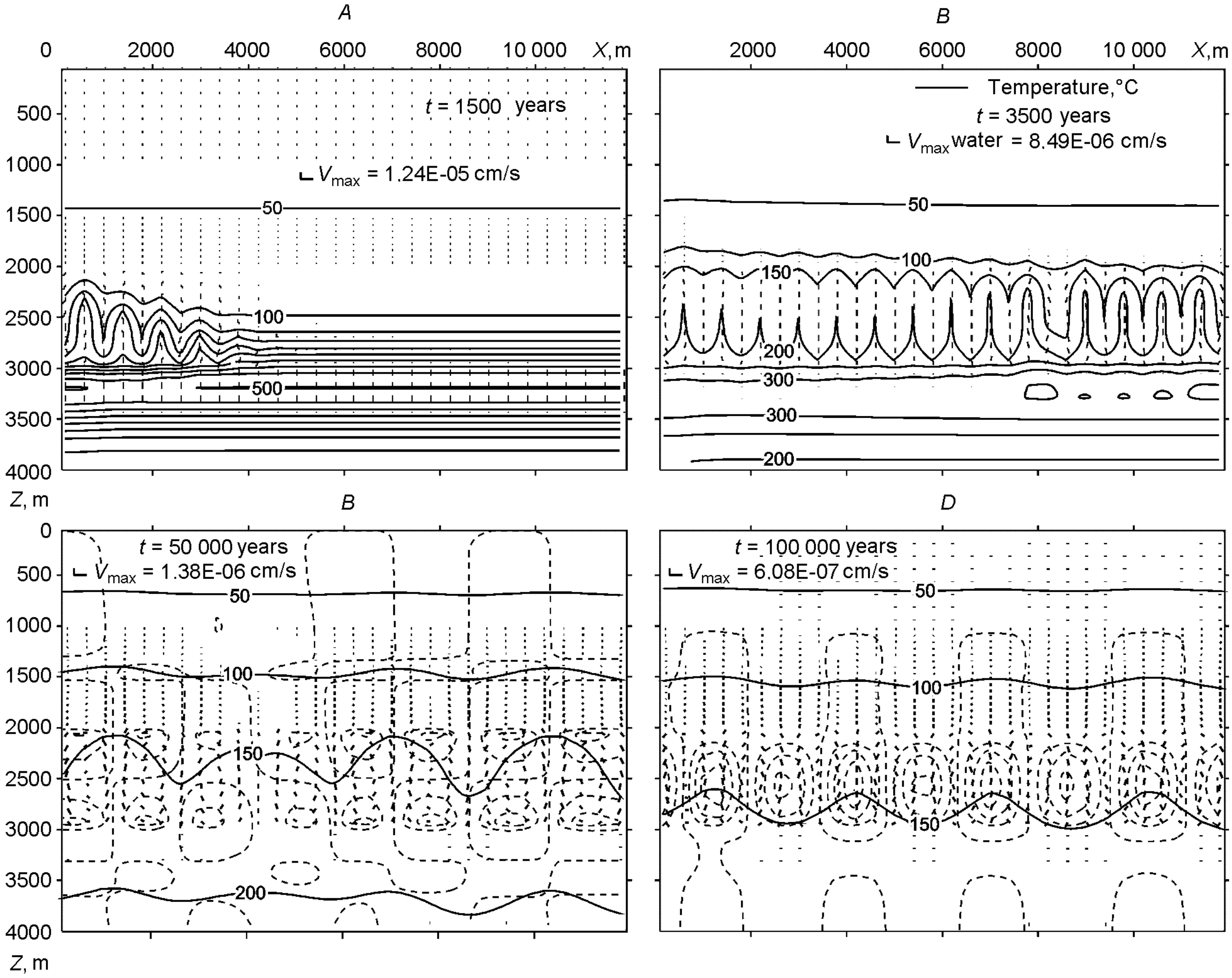

Fig. 3. Evolution of temperature and fluid flow for the model of intrusion of a single sill 250 m thick at the immediate contact with the lower reservoir for times of 1.000, 3500, 50,000, and 100,000 years after intrusion. The upper contact of the sill is at a depth of 3000 m. The initial temperature of intrusion is 900 °C; the initial perturbation δ = 100 °C was set at the left side of the sill. Step of the isotherms is 50 °C. Dashed line shows isolines of velocities of horizontal V_x cm/s (fig. *C*) and vertical V_z cm/s (fig. *D*) filtration. Here and in the following figures, vectors of fluid velocity are given in logarithmic scale, with scales of a vertical and a horizontal components indicated for each plot. The vertical axis is twice the scale of the horizontal axis.

the axis of symmetry. Since the lower reservoir is heated laterally and from above, the convective instability does not develop in it at a distance from the dike. Above the upper contact of the horizontal sill and along the contact with the dike, the pore fluid boils up. In Fig. 5, these domains are marked by bold primes. Boiling up of a pure water fluid is possible at depths of no more than 2–2.5 km, when the pressure does not exceed the critical value.

A dike and a sill at the bottom of the basin. Figure 2, *B* shows the cooling history of the host rock at the dike exocontact 250 m apart from the axis of symmetry of the model. At the initial moment temperature is assumed to be equal in depth (1200 ℃). The dike cools nonuniformly: more quickly at the contact with reservoirs and more slowly, at the contact with seals. The maximum difference in temperatures between the reservoir and seal does not exceed 110–130 ℃. The temperature in the carrier bed appears to be lower than in the cap rocks during the fluid circulation. Then this effect disappears at the stage of conductive cooling. After a lapse of 40,000 years the effect of convection becomes markedly weaker, and the form of geotherms becomes typical of conductive cooling. Thus, the duration of the convection stage, when the temperatures in the reservoir and in the seal markedly differ, is 20–40 ka under indicated parameters. With time, the dike and the sill cool until equlibrium initial temperatures.

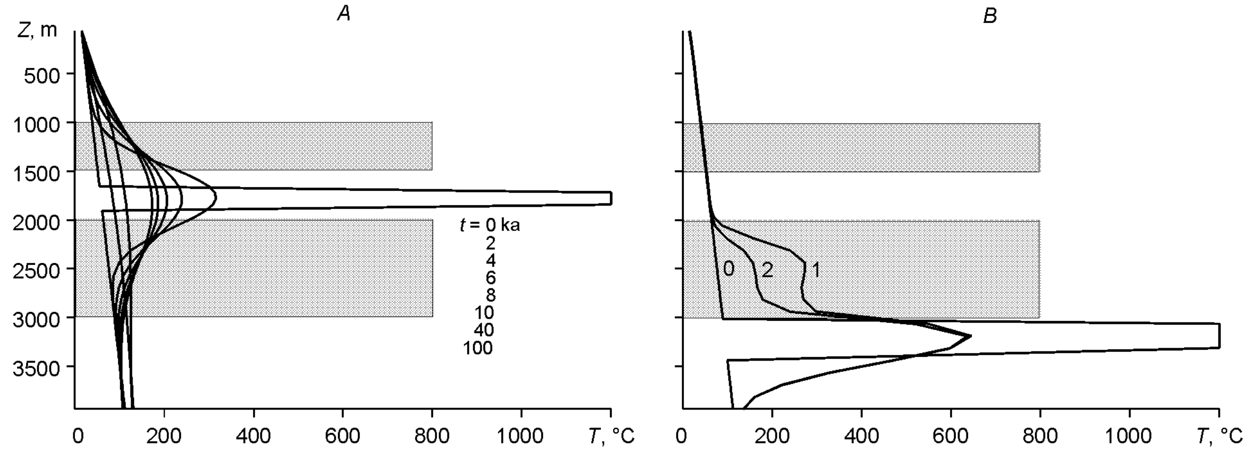

Fig. 4. *A*. Temperature profiles for the model of a 125 m thick sill intruded between reservoirs. Profiles are drawn through the center of model area for times (ka) shown right. *B*. Temperature profiles for the model of a 250 m thick sill intruded beneath the lower reservoir. Digits at the curves in Fig. 3, *B* mean: 0, temperature distribution on sill intrusion; 1, temperature profile along the ascending flow; 2, along the discending flow 3500 years after the intrusion under developed convection in the lower reservoir. The location of the reservoirs is shown in gray.

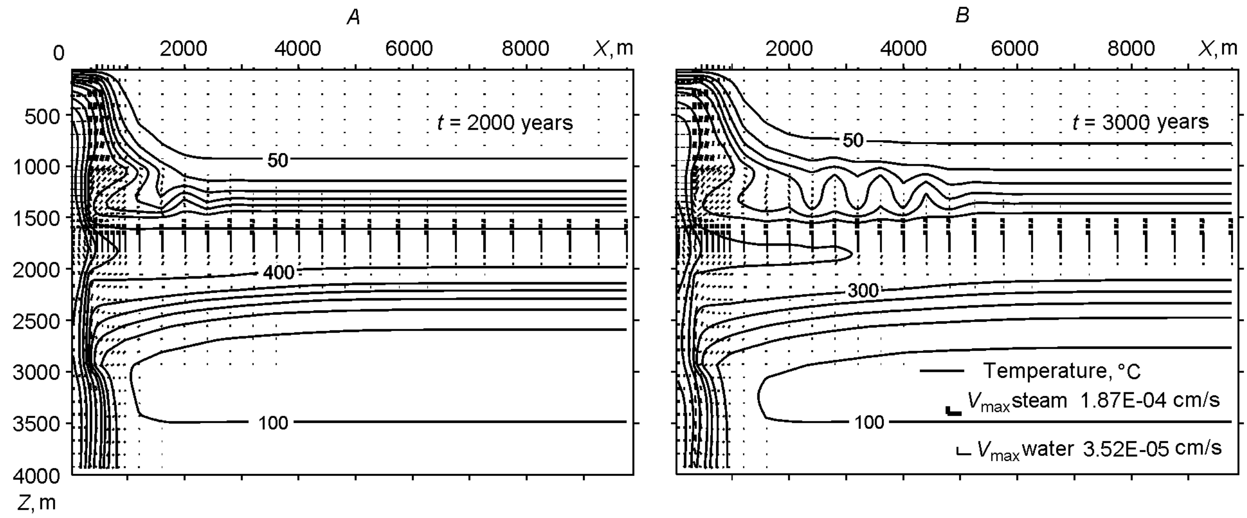

Fig. 5. Onset of development of convection (after 2000 and 3000 years) caused by the intrusion of a vertical dike (left) and a sill between reservoirs. Convection in the upper reservoir is associated with propagation of a "temperature wave". The interval between isotherms is from 50 to 300 °C at 50 °C and then 100 °C. Afterwords fluid passes into gas phase above the upper contact of the sill and in contact with the dike at shallow depth. The boiling zone is shown by bold primes.

Figure 6 shows the history of convection and temperature from early times (1000 years) to nearly complete extinction of convection (200,000 years), when the temperature decreased nearly to the initial one (less than 150 °C). At a depth of less than 2 km boiling occurs at the dike-host rock contact, but ceases 5000 years later. The width of the zone of overheated vapor and two-phase state of fluid does not exceed the thickness of the dike. As magma cools (Fig. 6, *C*, *D*), the fluid everywhere in the region to be modeled is in single-phase state and continues to circulate in both lower and upper reservoirs, but in the lower reservoir the convection attenuates more slowly. The

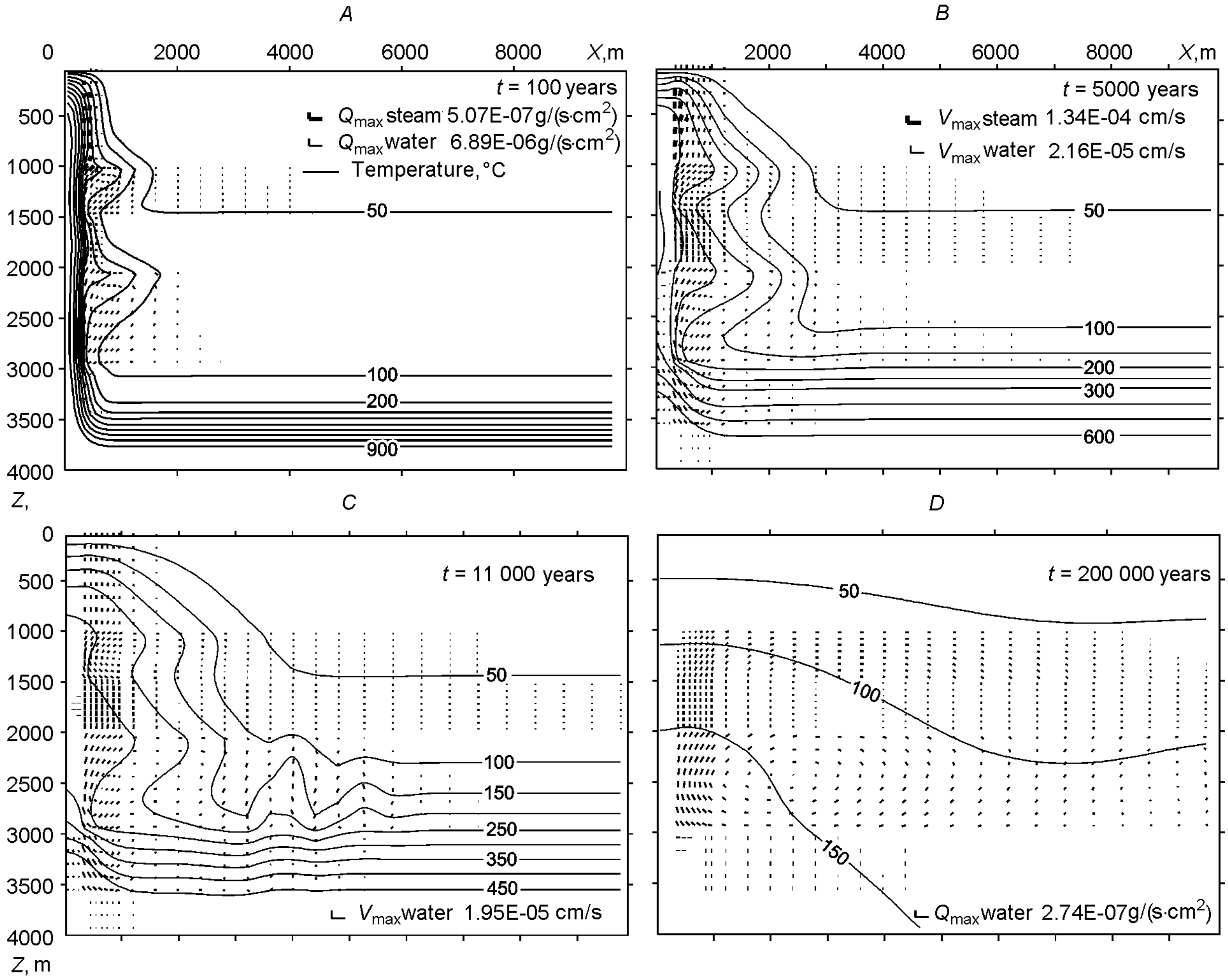

Fig. 6. Results of modeling of a combined model for cooling of a dike (at the left boundary) and a 250 m thick sill at the bottom of the basin with an initial temperature of 1200 °C. Values of isotherms are given in °C. The plots are given for the times 1000 (*A*), 5000 (*B*), 11,000 (*C*), and 200,000 (*D*) years after intrusion. The boiling zone is shown by bold primes. The main cause of convection is the presence of a dike (horizontal temperature gradient), whereas the effect of a sill is less pronounced.

isotherms at this stage do not become S-shaped as in Fig. 6, *A–C*, and the temperature at each point of the medium decreases. After a lapse of 200,000 years, two convective cells 8 and 2 km wide are preserved in the lower reservoir. Later they merge into one cell, as in the upper reservoir where the flow of fluid is unidirectional, and at later stages of the process the convective cell entrains the intermediate layer between the reservoirs (Fig. 6, *D*).

As in the previous model, the isotherms are oscillating in the reservoir which is closer to the sill. In the model under consideration this is the lower reservoir where several convective cells form for 10,000 years (Fig. 6, *C*). In a remote reservoir, convection is governed only by lateral heating, and at any time there is only one cell with a narrow ascending and a wide descending plumes. At late times, there is only one "global" cell, and the movement of fluid particles is traceable from the lower reservoir to the upper one through the intermediate layer and back to the lower reservoir, i.e., "penetrating" convection takes place.

A dike and a sill in contact with lower reservoir. This situation implies that the magma is intruded as a vertical dike, with an off-branching sill which is in contact with the lower reservoir. In this case convection is governed by two factors: lateral heating and heating from below. The results for times of 2000, 6000, and 8000 years after magma intrusion are shown in Fig. 7. After a lapse of 2000 years, cells appear in the lower reservoir: a 3–3.5 km wide cell near the contact and a number of narrow cells about 1 km in width equal the thickness of the reservoir. The periodical structure of convection forms under the effect of uniform heating by the intruded sill.

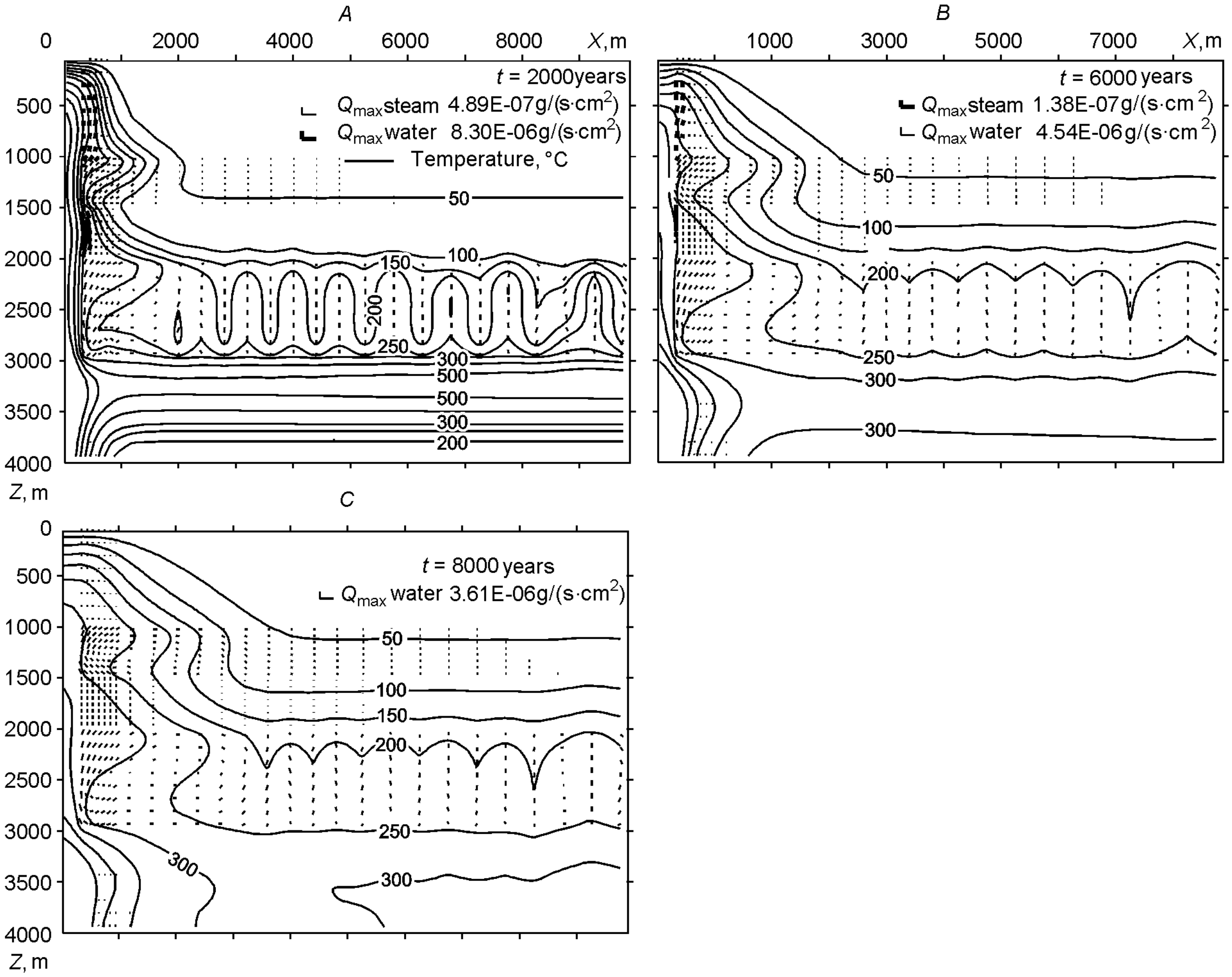

Fig. 7. Development of convection in the model for emplacement of a dike and a sill 250 m thick at the immediate contact with the lower reservoir for times of 2000 (*A*), 6000 (*B*), and 8000 (*C*) years after intrusion. The periodic structure of the thermal field in the lower reservoir is caused by the effect of the sill (cf. Fig. 6). At its contact with the dike boiling occurs in two zones separated by the reservoir (see the text for comments).

In the upper reservoir a single cell forms with a localized ascending flow near the vertical dike and a wide descending flow. It has no periodical structure because the sill is rather distant. A domain of overheated vapor forms at the contact with the dike above and below the upper reservoir (Fig. 7, *A*, *B*). At the contact with the reservoir vapor does not form, because the heat is advected from this domain by a vertical current of fluid and cooling is more effective.

As can be seen from Fig. 7, a quasithermostatic domain forms in the lower reservoir, between the isotherms 200 and 250 °C, which is stable for about 10,000 years. From the very beginning of cooling of intrusions, a temperature gradient appears in this reservoir: from 200 °C at the roof to 250 °C at the bottom of the layer. Because of intense convection of fluid, the temperature gradient in this layer decreases, given steeper gradients above and below. This model differs from the previous one in that the sill makes contact with the reservoir, and cells form along the whole length of the layer. Only one cell exists in the upper reservoir; the lower reservoir abounds in them, but with time the left cell enlarges at the cost of the others.

These calculations show that a dike creating a horizontal temperature gradient leads to a loss of fluid stability in the layer irrespective of the place where the sill is intruded. The sill creates a vertical temperature gradient which disturbs the fluid stability only when the gradient is above some critical value.

IMPLICATIONS OF MODELING OF AQUEOUS FLUID CONVECTION FOR PREDICTION OF BEHAVIOR OF HYDROCARBON FLUID

Studies of the behavior of water-steam system with consideration of phase transition are important for many hydrothermal and hydrogeological processes. In addition, we believe that this simple system can be a certain analog of the behavior of liquid hydrocarbon–gas system.

It is usually supposed that petroleum forms in fine-grained sedimentary rocks by thermal diagenesis of organic matter with subsequent primary migration of hydrocarbons (HC) into carrier beds (permeable) and secondary migration into oil and gas traps [37]. Forces of floating and capillarity play an additional role in case of phase immiscibility, but these factors are ignored in our approach. Geochemical and temperature aspects of generation and alteration of oil have been studied quite well, whereas mechanisms of migration and accumulation at a scale of sedimentary basin remain unclear. Regional flows in the process of secondary migration provide the main transport mechanism necessary for accumulation of hydrocarbons from dispersed centers of generation and their concentration in oil traps [36]. Depending on thermodynamic conditions, the process of secondary migration can include the transport of HC either in the form of immiscible drops or as a one-phase mixture of hydrocarbon fluid [38]. The fluid convection caused by magma heat should be regarded as a process of secondary migration. Therefore, it is important to know the phase state of the mobile medium. Modeling of phase behavior on the basis of state equation shows that mostly during its generation, expulsion, and migration, the natural oil fluid remains one-phase. Only at depths of less than 3 km, significant phase fractionation can occur [39]. At greater depths, the phase fractionation is possible only for gas-saturated fluids, in which the fraction of liquid phase is small.

To replace the water-steam system by the natural gas-oil system, we must know thermodynamic properties of hydrocarbons especially with high molecular weights. At present, well-developed approaches to modeling of phase state and thermodynamic properties of HC fluids are available (so-called PVT modeling) which are based on a state equation derived from the law of corresponding states. When the law of corresponding states is applied to a mixture, a common practice is to determine critical parameters of a hypothetical pure substance having the same coefficients of state equation as the mixture [40]. To compare critical properties of pure water and some natural hydrocarbon mixtures, we report values of their critical parameters [41] (Table 2).

Table 2 shows that the critical point of water falls into the region of critical parameters for hydrocarbons of varying composition. Thus, we can consider the water fluid as an analog of some hypothetical mixture of hydrocarbons with pseudocritical properties close to those of a light gassy oil. The onset of convection of the HC fluid is governed by the Rayleigh number for the porous medium. Given the same parameters of the layer (thickness, 1000 m; permeability, 10^{-14} m^2; temperature gradient, 0.025 °C/m), the Rayleigh numbers for oil and water are close, Ra = 20 for the water and Ra = 12.8 for oil fluid, if we use viscosity, density, and other parameters of the fluid at formational pressure and temperature (here, at 45 MPa and 170 ° C, from [13]). Thus, a 30% increase in temperature gradient will cause convection of the HC fluid (the Rayleigh number will exceed the critical value). The magmatic intrusions in the form of sills at an initial temperature of 1000–1200 °C create necessary temperature perturbations (simultaneously decreasing the fluid viscosity) capable to initiate convection of both water and hydrocarbon fluids.

For comparison, we may use results of numerical modeling from [42]. The cellular pattern of convection was governed in this case by the kerogen-oil transition. A small local maximum of methane content in kerogen in the left portion of the region (Fig. 8) initiates convection caused by gravitational instability at the kerogen-methane phase transition. The initial disturbance leads to the rightward propagation of a wave, generating

Table 2

Critical Parameters of Water in Some Types of Oil after [28, 41]

Type of fluid	Critical pressure *P*, bars	Critical temperature *T*, °C
Water	220.55	373.98
Condensate	493	234.6
Light oil	191.8	400.8
Medium oil	179.2	334.4
Heavy oil	46.8	478.3

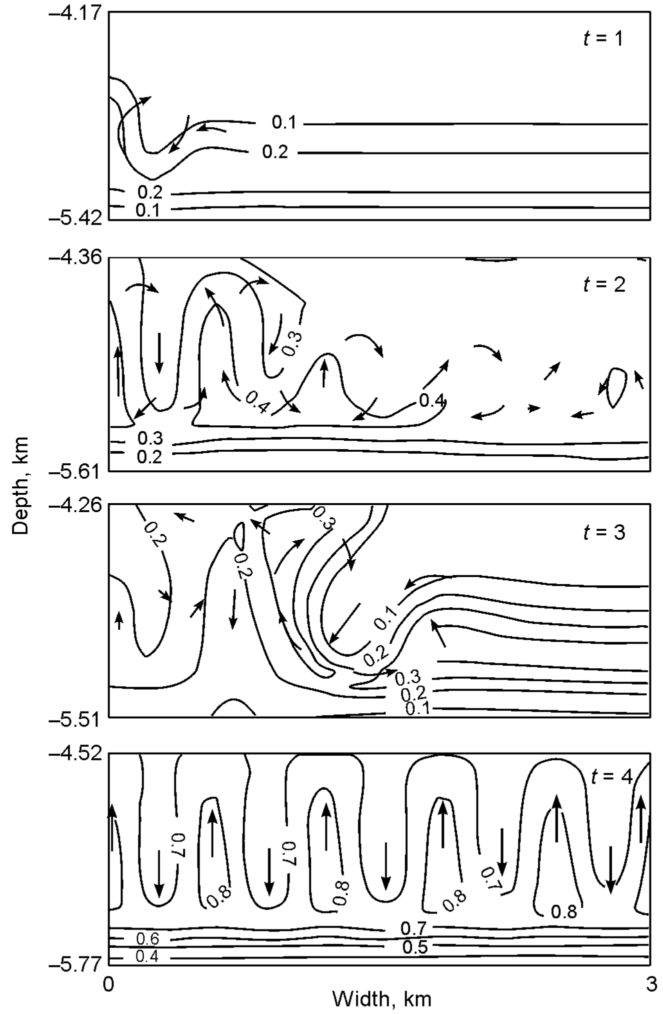

Fig. 8. Development of convection caused by the kerogen-methane phase transition, after the results of modeling in [42]. Isolines of concentration of methane and direction of fluid flow velocities are given for four successive times.

convective cells. We have obtained a similar pattern of propagation of heat convective wave in models with the water fluid for a single sill (Fig. 3) as well as in the presence of a dike and a sill (Fig. 5).

CONCLUSIONS

The two-dimensional model has shown that the earlier one-dimensional approaches in estimates of the degree of heating of the host rocks by magmatic intrusions are not satisfactory. Convection in highly permeable fluid-conducting layers can create wide zones of heating above horizontal sheets and near vertical magmatic bodies. In the same zones, a domain of existence of gas phase forms, whose lifetime is about 5000–6000 years at given parameters of rocks.

The results of the numerical modeling given in Figs. 2–7 may be considered a qualitative model of behavior (in respect to the character of convection and phase state) of hydrocarbons in sedimentary layers under the effect of magmatic heat sources of various kinds.

One of the important results of modeling of *P-T* conditions during magma intrusion in a sedimentary basin is the prediction of the existence of a zone with low thermal gradient in the reservoir where convection evolves. With respect to the hydrocarbon fluid this can create favorable conditions for maturation and improvement of oil quality when the intense mixing of liquid occurs and then nearly thermostatic phase lasts more than 10,000 years.

Temperature of 200–250 °C and pressures of 200–300 bars correspond to the two-phase region for a light HC mixture [40] with critical parameters indicated in Table 2. Thus, in reservoirs of sedimentary basins of the Yenisei-Khatanga or Tunguska types at depths of less than 3 km, where the phases are separated in an equilibrium (undisturbed) state, the sill intrusion can transfer the hydrocarbon fluid from a subcritical state to single-phase supercritical state. This transition will favor the faster migration of hydrocarbons due to lowering their viscosity.

Comparison of model results for pure water fluid and HC fluid with the kerogen-methane transition [42] clearly shows that the water-steam transition can approximate the kerogen-liquid oil transition. To quantitatively estimate the possibilities of convection with consideration of phase state of natural petroleum mixtures of different compositions, we plan to carry out further experiments by changing the program module for calculation of thermodynamic and transport properties of the fluid.

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