ISSN 0742-0463, Journal of Volcanology and Seismology, 2012, Vol. 6, No. 3, pp. 184–196. © Pleiades Publishing, Ltd., 2012.
Original Russian Text © N.S. Zhatnuev, S.N. Rychagov, V.I. Vasil'ev, E.V. Vasil'eva, 2012, published in Vulkanologiya i Seismologiya, 2012, No. 3, pp. 59–72.

The Influence of Rheologic Crustal Properties of the Crust on the Location of Ore-Forming Hydrothermal Magmatic Systems

N. S. Zhatnuev[a], S. N. Rychagov[b], V. I. Vasil'ev[a], and E. V. Vasil'eva[a]

[a] *Geological Institute, Siberian Branch, Russian Academy of Sciences, 6a ul. Sakh'yanovoi, Ulan-Ude, Russia 670047*

e-mail: zhat@gin.bscnet.ru

[b] *Institute of Volcanology and Seismology, Far East Branch, Russian Academy of Sciences, Petropavlovsk-Kamchatskii, Russia 683006*

e-mail: rychsn@kscnet.ru

Received March 30, 2010

Abstract—Based on a comprehensive study of hydrothermal magmatic systems at island arcs and a review of available mechanisms that cause elasto-plastic deformation in rocks, we considered the conditions for interaction between a convective magmatic cell and a convective hydrothermal cell in different rheologic zones of the crust. Three models have been developed to describe the generation of hydrothermal circulation systems: (1) the magma chamber is localized in a plastic zone, (2) partial and (3) complete penetration of the chamber into a brittle crust. It is shown that the last of these models is highly consistent with the structure of present-day high-temperature hydrothermal magmatic systems at depths greater than 1.0–1.5 km and with the structure of Miocene to Pliocene ore-bearing volcano-plutonic complexes that are eroded to different depths in different geologic blocks within these complexes.

DOI: 10.1134/S0742046312030062

INTRODUCTION

A hydrothermal system is a part of the Earth's crust of regular structure above a source of heat supply and in the region of its influence within which a favorable combination of geologic bodies, permeable zones, and hydrogeological structures creates conditions for heat energy to be transported from certain depths to the ground surface by fluid convection in a liquid or gaseous phase [*Struktura* ..., 1993]. The structure of a hydrothermal system is composed of rocks, hot (cooling) magma bodies, hydrothermally altered zones, permeable tectonic faults and faults that have been welded with secondary minerals, regions of boiling hydrothermal occurrences, water-bearing horizons and aquicludes, zones of thermodynamic and geochemical barriers and other elements of geological space that involve phase transformation of material and the generation of ores and minerals influenced by chemical reactions that either release or absorb heat.

Any model for hydrothermal systems in areas of present-day volcanism assumes the presence of a magma chamber or a hot intrusive body at depth as the main source of heat [Aver'ev, 1966; *Gidrotermal'nue* ..., 1976; Kononov, 1983]. However, it is only in recent years that it was possible to prove the presence of a transition zone between a near-surface hydrothermal convective cell and a deep-seated magmatic convective cell in the structure of a high-temperature hydrothermal systems; this research was based on deep drilling data, a detailed study in the structure of long-lived volcanogenic metallization centers, and deep drilling on andesitic volcanoes at island arcs. Such geological structures were defined as long-lived, ore-forming, hydrothermal, magmatic, convective systems in the ocean–continent transition zones [Rychagov, 2003], to be referred to below as hydrothermal magmatic systems.

The present authors use theoretical concepts and thermal hydrodynamic models, as well as materials from multidisciplinary studies of hydrothermal magmatic systems at island arcs and the Au–Ag–Cu–Mo–... ratios in porphyry, as well as epithermal ore and geothermal fields, to suggest a new model to describe the interactions between magma chambers (hot intrusive bodies) and gas–liquid circulation systems in different rheologic crustal zones.

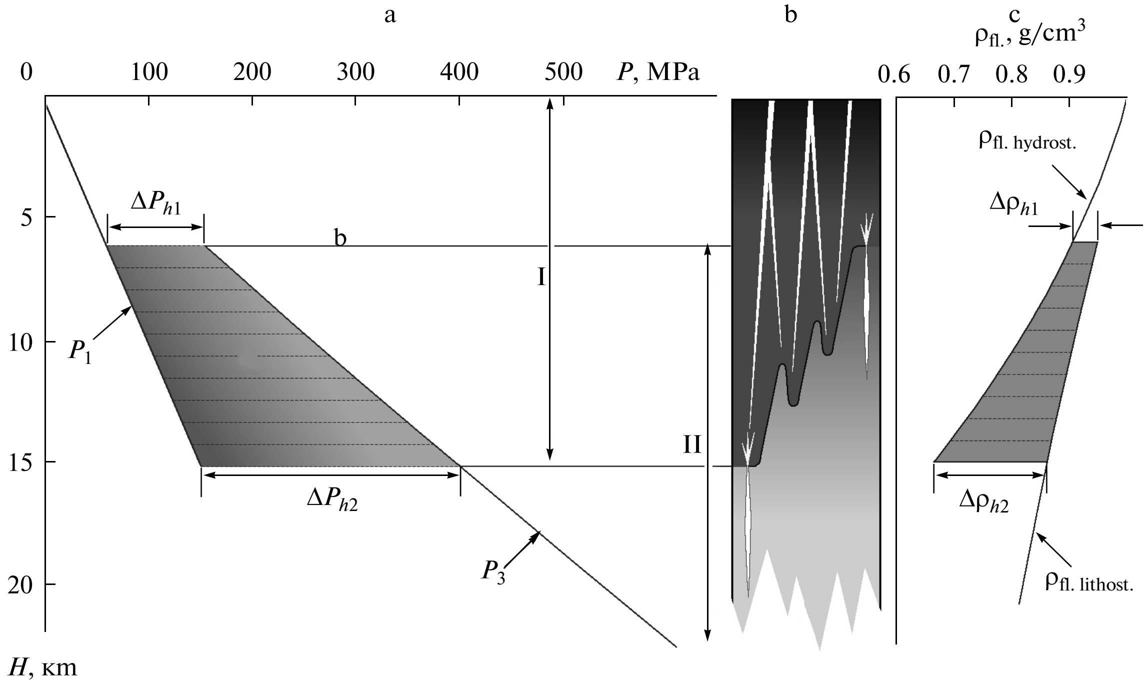

Fig. 1. A diagram that shows the elastic/plastic transition in the crust down to a depth of 20 km: (a) variation in the lithostatic/hydrostatic pressure relationship as a function of depth in relation to crustal rheology, (b) cross section of the model system, (c) fluid density contrast in the system along the vertical (ρ_{h1} and ρ_{h2} for the case of the plastic/brittle transition occurring along the 30°/km geotherm.

P_1 is fluid hydrostatic pressure in fissures in the brittle region, P_2 is the fluid pressure in pores, which is equal to the lithostatic pressure in the plastic region, P_3 is the transition zone from P_1 to P_2 in relation to the physical state of host rock. At different transition depths we also have different pressure contrasts (ΔP_{h1} and ΔP_{h2}). I is the zone of open fissures in the brittle region, II is the zone of closed fissures or no fissures in the zone of plastic deformation.

THE ELASTO-PLASTIC TRANSITION IN THE EARTH'S CRUST AND THE EVOLUTION OF HYDROTHERMAL MAGMATIC SYSTEMS: THE CONCEPT

It is known that brittle deformation is replaced by plastic deformation as one goes downward from the ground [Vashchilov, 1984; Ivanov, 1970, 1990; Nikolaevskii, 2001; Pavlenkova, 2001]. However, the depth of this elastic/plastic transition is estimated differently. For example, S.N. Ivanov's assertion is that continental crust is subject to brittle deformation down to depths of 6–10 km with plastic deformation dominating greater depths [Ivanov, 1990]. His other thesis is that the fluid exists in open cracks under hydrostatic pressure within the brittle crust, while below it resides in isolated cavities and pores in the conditions of the maximum lithostatic pressure [Ivanov, 1970, 1990]. According to Nikolaevskii [2001], the true plasticity of the geological medium is reached at the Moho; this makes the rocks below that interface impermeable. It is supposed that the zone of elasto-plastic transition in oceanic crust is shallower, but still possibly below the Moho. For example, according to Bazylev [1992], oceanic water in Atlantis Fracture Zone penetrates to depths of about 10–20 km, which must be in the upper mantle for oceans.

The different estimates of the elastic/plastic transition depth may be due to the fact that different geodynamic settings have different rates of deformation. If a rock behaves as a plastic material when it is under a relatively slow deformation, then it will exhibit elastic properties as the rate increases. For example, according to Fyfe et al. [1978], the yield limit as a limestone specimen is compressed occurs in the stress range between 1 and 10 MPa with the respective rates of deformation between 1.2×10^{-7} s and 1.9×10^{-1} s. It seems that one has to assume a certain range for the brittle–elastic transition corresponding to extremely low and extremely high rates of geological movements.

Figure 1 shows a plausible range for the elastic/plastic transition in the depth–pressure coordinates [Kissin, 2001]. To a first approximation, the solutions that reside in the upper (brittle and cracked) part of the earth's crust are under their own hydrostatic pressure; if the fluid density is assumed to be 1 g/cm^3, then the pressure plot will correspond to the P_1

line. Actual pressures may differ by a value that is proportional to the decrease in fluid density over depth due to thermal expansion. The P_2 line shows a possible fluid pressure when the rocks in isolated pores and cracks are plastic. The pressure under these conditions is controlled by the lithostatic load and depends on rock density in the depth range between the ground and the depth where the cracked pore fluid system is found. When an isolated crack that contains a fluid penetrates from the plastic zone into a brittle region, which is quite possible for a migrating crack (the mechanism that is responsible for crack migration in a plastic medium was discussed previously in [Zhatnuev, 2005]), there must be a sharp drop in fluid pressure from lithostatic to hydrostatic with accompanying adiabatic cooling and the precipitation of dissolved material. The value of this pressure drop will be different for different depths of the transition from plastic to elastic rocks. For example, the drop will be ΔP_{h1} for a depth of 6 km, which is much lower than that for 15 km (ΔP_{h2}, see Fig. 1a). Accordingly, we shall have different changes in fluid density, which will vary between ρ_{h1} and ρ_{h2} in the range of these depths (see Fig. 1c). As the fluid density increases, so does the fluid dissolving capability under higher temperatures.

Whatever the depth of the elastic/plastic transition, this transition is important in controlling the location of magma chambers, ore magmatic complexes, and geothermal fields. The transition from the brittle to the plastic state does not produce cracking and porosity, as well as large-scale convective motions in the fluid, and must limit the generation of hydrothermal systems. In practice, however, deep boreholes and microseismic probing revealed zones (restricted volumes, block, or horizons) that have contrasts in their petrophysical rock properties. This is related to zones of lower rock density at different depths, as far down as 10–12 km [*Kol'skaya* ..., 1984], that are saturated with fluids in the form of gas and/or liquid. Such zones are oil and gas traps; they widely occur at the boundaries of older platforms, within volcanic belts, and in areas of tectono-magmatic activation [Rychagov, 2003; Khristoforova et al., 1999].

Depending on rock mechanical properties (brittleness and plasticity), the hydrothermal systems that are generated by a magma chamber may develop along the ways specified by the scenarios that follow.

Scenario 1. The chamber is in a brittle crust (fragment 1A in Fig. 2). As hot magma is being emplaced, a plastic shell of host rock forms around the new magma chamber [Efimov and Ershova, 1999]; the shell prevents meteoric water from contact with the edge parts of the cooling intrusion and hampers mass exchange between the melt and the hydrothermal occurrences that occur in the host-rock crack–pore space. Nevertheless, convection arises at some distance from the magma chamber in rocks that have not been subjected to plastification owing to poor heating of the geological material during this stage. The convection gives rise to a flow of hydrothermal solutions that is upward near the chamber boundaries and downward at the periphery of the heated part of the system.

The convection of solutions that are dominated by meteoric water drives mass transport in the host rocks and the upward heat transfer. As the magma cools and crystallizes, the edge part of the intrusion becomes brittle, the front of plastic rocks recedes into the chamber, and the solutions penetrate into the intrusive body along contraction cracks (see fragment 1B in Fig. 2). The meteoric material of the solutions and the juvenile magmatic material are mixed. At the same time, the magmatic fluid separates to form an independent phase as a result of retrograde boiling during the melt crystallization; the pressure in that phase exceeds the strength of the impermeable plastic shell and the phase breaks from the chamber, afterwards mixing with meteoric hydrothermal water.

Scenario 2. The magma chamber is partially emplaced in a brittle crust (see fragment 2A in Fig. 2). Most of the chamber is within the plastic zone. The convection of meteoric solutions in the zone of brittle deformation drives mass transport in the host rocks and transfers heat upward to the ground surface. As crystallization and cooling proceed, the apical part of the massif loses plasticity, it experiences brittle cracking and is "washed out" by hydrothermal solutions. In contrast to the first scenario, the part of the intrusion that is in the plastic zone cools at a lower rate, because there is no convective transport of heat outward. As a result, the lower part of the magma chamber remains in a molten state, is isolated in the plastic zone, and hardly exchanges material with the hydrothermal discharges at all. All of this increases the time of melt crystallization in the chamber and favors the preservation of juvenile composition for the igneous rock that is thus being generated.

Scenario 3. The chamber completely remains in the plastic zone (see fragment 3A in Fig. 2). Magma convection transports heat to the upper part of the chamber. Conductive heat transport into the host rocks mostly occurs via the apical zone of the intrusive body, which is related to convection in the magma. No convective mass transfer with the host rocks occurs. The only event that is still possible is batch separation of fluids that contain mineral material, with the separation occurring during melt crystallization due to retrograde boiling or decompressional degassing in case the magma is oversaturated with fluids. The convection of

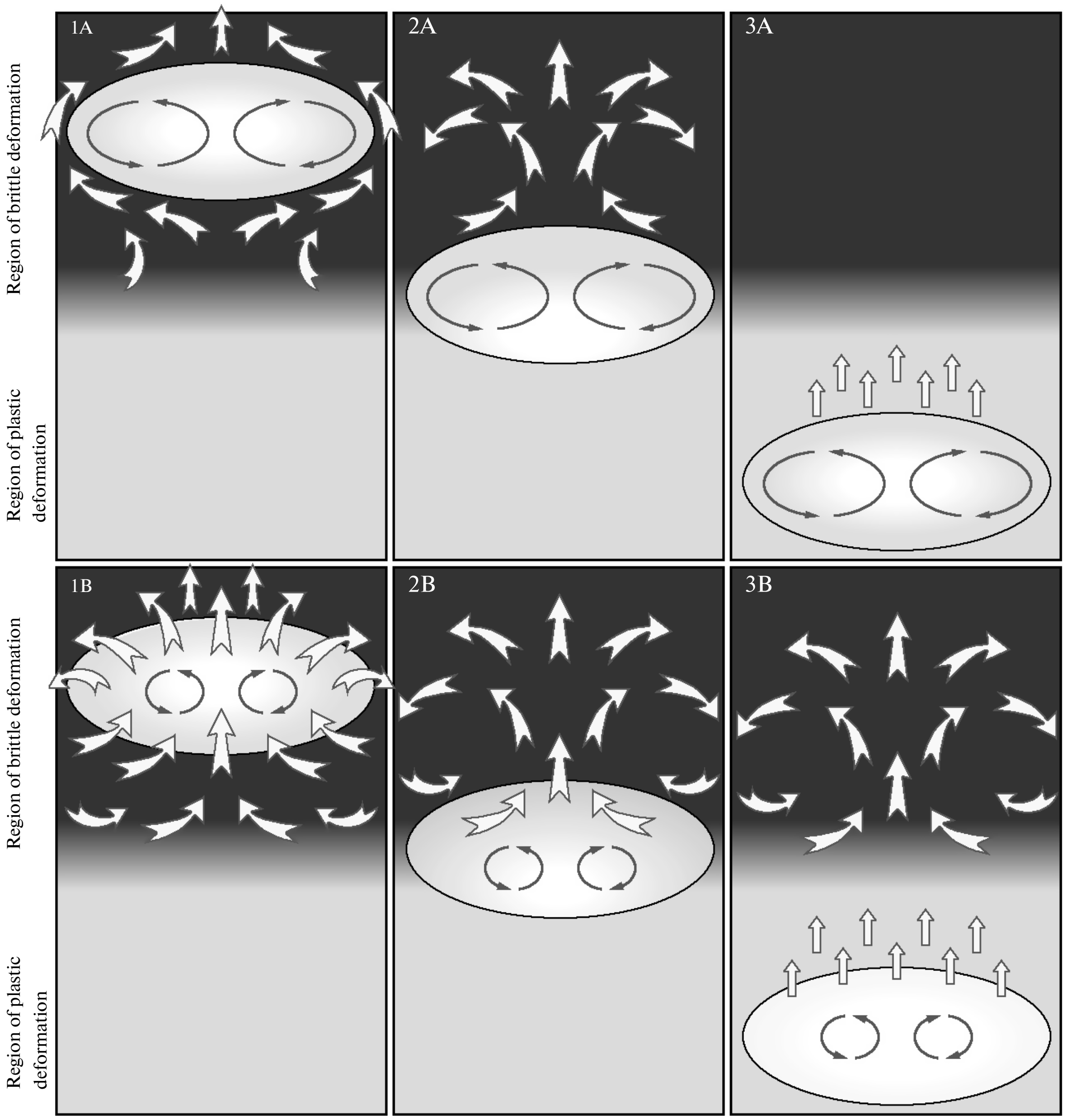

Fig. 2. Emplacement of magma chambers in different rheologic zones of the crust and patterns of convective hydrothermal flows in relation to the location of the heat source.

In the region of brittle deformation the fluid pressure in the host rocks is hydrostatic and the heat-and-mass transport is convective. In the region of plastic deformation the fluid pressure in host rocks is equal to lithostatic pressure, the heat transport is conductive, and the transport of material occurs by diffusion. (1) location of the magma chamber in the zone of brittle deformation, (2) location of magma at the boundary between the zones of elastic and plastic deformation, (3) location of the magma chamber in the zone of plastic rock; A the initial stage of the system evolution after magma emplacement, B evolution of the convection in the system after partial magma crystallization. The larger arrows indicate the convection of solutions in the zone of brittle deformation, smaller arrows show the conductive heat flow from the magma reservoir located in the zone of plastic rock; light circular arrows indicate the magma convection in the chamber.

meteoric solutions mostly occurs by conductive heating of the rocks in the brittle part of the section above the magma chamber (see fragment 3B in Fig. 2). The mineral composition of hydrothermal solutions is not affected by juvenile material, it completely forms by interaction with the host rocks. In contrast to the first two scenarios, the intrusive body enclosed in the plastic zone will crystallize for a long period, since con-

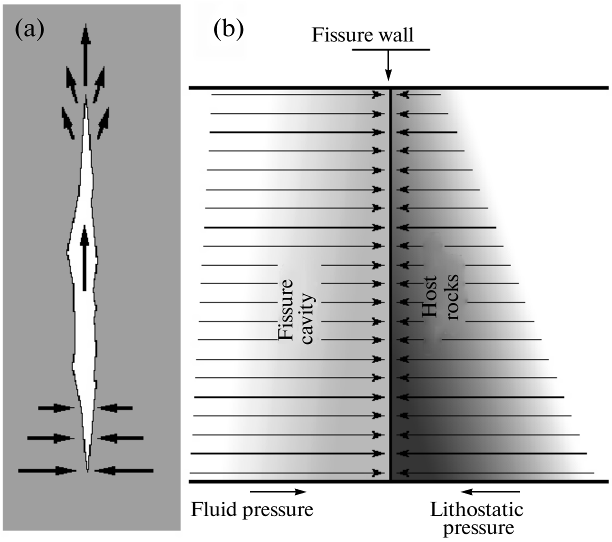

Fig. 3. Mechanism for the upward motion of a fissure cavity filled with a fluid in a plastic rock: (a) model of the fissure cavity (arrows at the top show the hydraulic fracture at the "head"; the arrows at the bottom indicate the compression of the cavity by lithostatic pressure, the arrow inside the cavity shows the direction of fluid movement), (b) vectors of fluid and lithostatic pressure that act on the cavity wall.

ductive heat exchange is much less effective compared with convective exchange.

It follows from the above that hydrothermal magmatic systems will evolve in different ways, depending on the depth where the paternal intrusive body is located. Three evolution scenarios are possible. The first consists in the localization of magma chambers in the zone of brittle deformation; the second is localization at the boundary between the zones of brittle and plastic deformation; and the third is in the zone of plastic deformation. Each of these models has its own crystallization of the magma melt and has its own specific features in the generation of hydrothermal systems.

In the first case the magmatic system (the subvolcanic chamber) cools rather rapidly and the crystallized massif is "washed out" by hydrothermal solutions. The process involves practically complete mixing of juvenile and meteoric material. In the second case fast cooling of the apical part of the massif occurs and partial mixing of juvenile and meteoric material. During the process the deeper part of the massif may be evolving for a long time and this involves the deep magmatic differentiation of the material. In the third case the conductive cooling of the intrusive body results in long-continued differentiation of the melt in the magma chamber, with the last phase generating pegmatites and vein bodies combined with the separation and migration of magma fluids by the mechanism discussed in [Zhatnuev, 2005]. "Contamination" of the massif with meteoric material is virtually ruled out. At the same time, low-temperature hydrothermal systems may form by conductive heating of the geological medium in the brittle region; these systems have eminently meteoric compositions of their solutions during the initial phase of the evolution. In the first and second cases the hydrothermal systems involve, during the initial phases, the generation of thick (300–500 m or greater) steam zones whose boundaries are effective geochemical barriers, including ore barriers [Zhatnuev et al., 1996].

In the third case (the retrograde boiling of magma) the residual water–silicate liquid may separate and migrate upward along moving cracks as far as the zone of brittle deformation. The mechanism of crack migration of fluids in the zone of plastic deformation was developed previously [Zhatnuev, 2005]. The mechanism is essentially as follows. An excess pressure arises in the "head" part of a crack that is filled with the fluid and is in the gradient field of lithostatic pressure, with the excess pressure exceeding the rock strength. It fractures the plastic rocks above the crack and, at the same time, the lithostatic pressure compresses the crack around its tail (Fig. 3a). The distributions of lithostatic and fluid pressure on the crack wall in relation to crack height are shown in Fig. 3b. This process makes the crack move against the lithostatic pressure vector toward the zone of brittle deformation where the fluid is under its own hydrostatic pressure. The transition from lithostatic to hydrostatic pressure, as the break-through from the plastic to the brittle zone occurs, produces an adiabatic expansion of the fluid accompanied by considerable cooling and precipitation of the dissolved material. For this reason the zone of elastic-to-plastic transition shows intensive generation of secondary minerals, in particular, silicification and filling of cavities, pores, and cracks with tridymite, cristobalite, opal, chalcedony, and quartz. When the fluid is thus ejected from cracks, its density drops dramatically. The drop varies in relation to the depth of the elastic/plastic transition (Fig. 4). Figure 4a shows the isolines of the water density in P–T coordinates with due account of the elastic/plastic transition at a depth of 8 km. The rocks are porous and cracked in the zone of brittle deformation (within 8 km depth), which allows the water solutions to be under their own hydrostatic pressure. The depth–pressure relationship is that for liquid water. Below an 8 km depth the solutions have lithostatic pressure, hence the slope of the water isopleths is significantly altered. The same slope of isopleths is also characteristic for the field of basaltic melt whose solidus can also be seen in the diagram. The projection of the water isopleths onto a hypothetical cross section of the

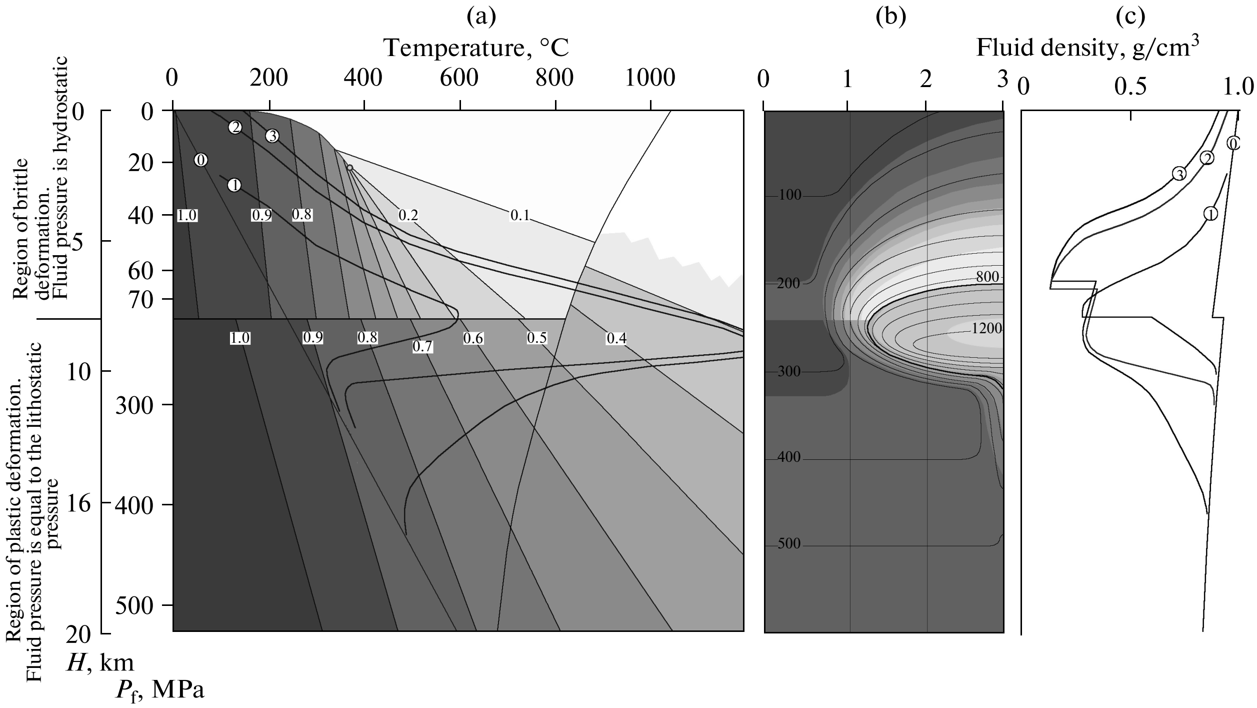

Fig. 4. Fluid density contrast between the zone of plastic deformation and the zone of brittle deformation. (a) isolines of water density in the P–T coordinates and with incorporation of the elastic/plastic transition at a depth of 8 km, (b) a hypothetical section with isotherms across the magma chamber and projections of water isopleths from diagram (a) onto the section, (c) variation of water density in sections 1–3 (numerals in circles).

Diagram (a) also displays the solidus of water-containing basaltic magma: the position of water isopleths in the field of basaltic melt is the same as that in the zone of plastic deformation; we also plot here geotherms of a hypothetical magma chamber, the numerals in circles at the geotherms indicating the temperature distributions in sections 0–3 in fragment (@c). Heavy line in diagram (b) shows the basalt solidus as projected onto the section.

magma chamber (see Fig. 4a) shows that water density experiences a considerable contrast at the elastic/plastic boundary. As well, if the melt contains free water in bubbles, its density will exceed that of water in the brittle zone at the same depth. Figure 4c clearly shows contrasts in fluid density at the elastic/plastic transition with a geothermal gradient that has not been disturbed by the thermal anomaly of the magma chamber (plot 0) and in the temperature field of the magma chamber (plot 1). In the former case the density contrast is small, while in the latter case it is much higher (about 0.3 g/cm^3). As well, a large contrast in the density of water fluid (about 0.2 g/cm^3) occurs at the boundary between the host rocks and the melt (plots 2 and 3).

The mechanism of fissure eruptions that discharge basaltic melts and a model for the evolution of a system of fissures and dikes above major peripheral and deep-seated magma chambers were developed by Fedotov [1984]. He discussed the existing theoretical concepts, examined a great wealth of continuously recorded seismological data as well as in-situ volcanological and other observations conducted during the Great Tolbachik Fissure Eruption to prove that the ascent of basaltic magma occurred from upper mantle depths by narrow fissure channelways, with individual fissures being only 1.0–1.3 m across. The high speed of magma ascent (its value is estimated as between 2 and 20 cm/s on average) is also due, among other causes, to an excess pressure of gas-saturated magma in the head parts of the fissures [Fedotov, 1984].

The present authors used different data (knowledge in the geological structure of hydrothermal systems and intrusive bodies, geochemical parameters, the gas–water–rock interaction in the zone of transition from hydrothermal to magmatic convective cells) but arrived at similar inferences concerning the mechanisms that operate in the opening of fissures in an elasto-plastic medium. Based on [Fedotov, 1984] and on the mechanism proposed by Zhatnuev [2010], it can be hypothesized that a connected, long, magma-conducting set of fissures is generated by emplacement of a water–silicate liquid (melt or fluid) into individual fissures that are close in time and space.

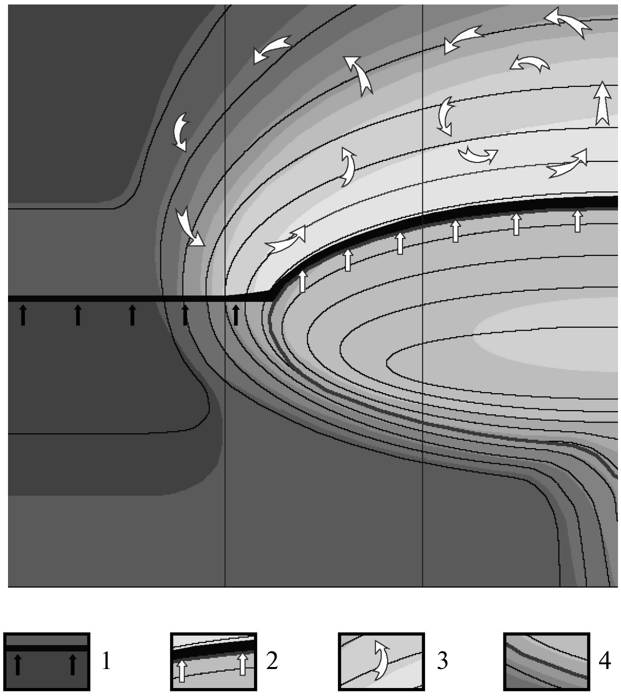

Fig. 5. The positions of mineralization zones in the section at the boundary of the plastic/brittle transition. (1) mineralization zone that forms as the fissure solution comes out from plastic rocks into brittle rocks, (2) same, as the solution comes from the melt into the host rocks, (3) convection of hydrothermal fluids in brittle fissured porous rocks, (4) isotherms and the melt solidus.

It was noted above that when in the thermal field of a magma chamber, hydrothermal fluids transport heat and material along open pores and fissures by convection. The transport and deposition of material will occur as the P–T conditions are changing over the section of the hydrothermal magmatic system. At the same time, if one accepts the hypothesis of fissure movement for fluids [Zhatnuev, 2005] under the conditions of plastic deformation, as well as the possibility of material being transported outward in bubbles and fissures from a boiling and crystallizing magma, one can then assume a dramatic expansion of the fluid as it

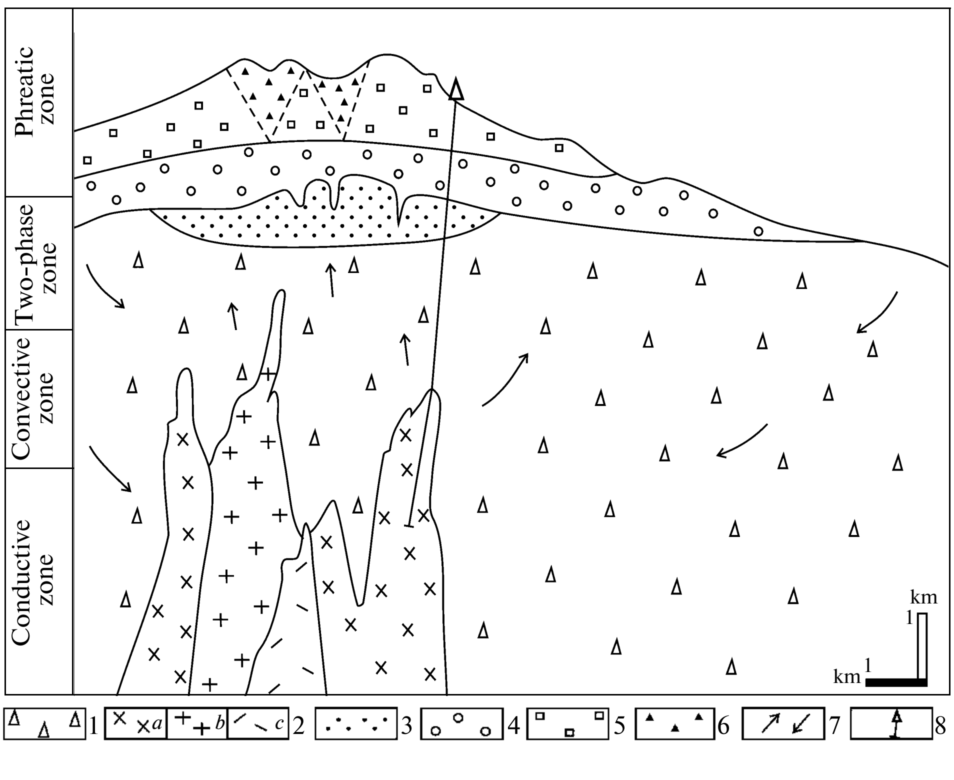

Fig. 6. A conceptual model for an island-arc hydrothermal magmatic system based on the study of epi- and meso-thermal and porphyry deposits at the Philippine island arc [Corbett and Leach, 1998].

(1) volcanogenic–sedimentary host rock sequence, (2) multi-phase intrusion: a basic rocks, b intermediate rocks, c acidic rocks; (3) zone where deep fluids mix with sulfate–bicarbonate solutions and the region where ore minerals are deposited, (4) zone of sulfate–bicarbonate hydrothermal fluids, (5) zone of meteoric water infiltration, (6) explosion craters of volcanic or hydrothermal origin, (7) direction of fluid flows, sea or mixed waters, (8) borehole.

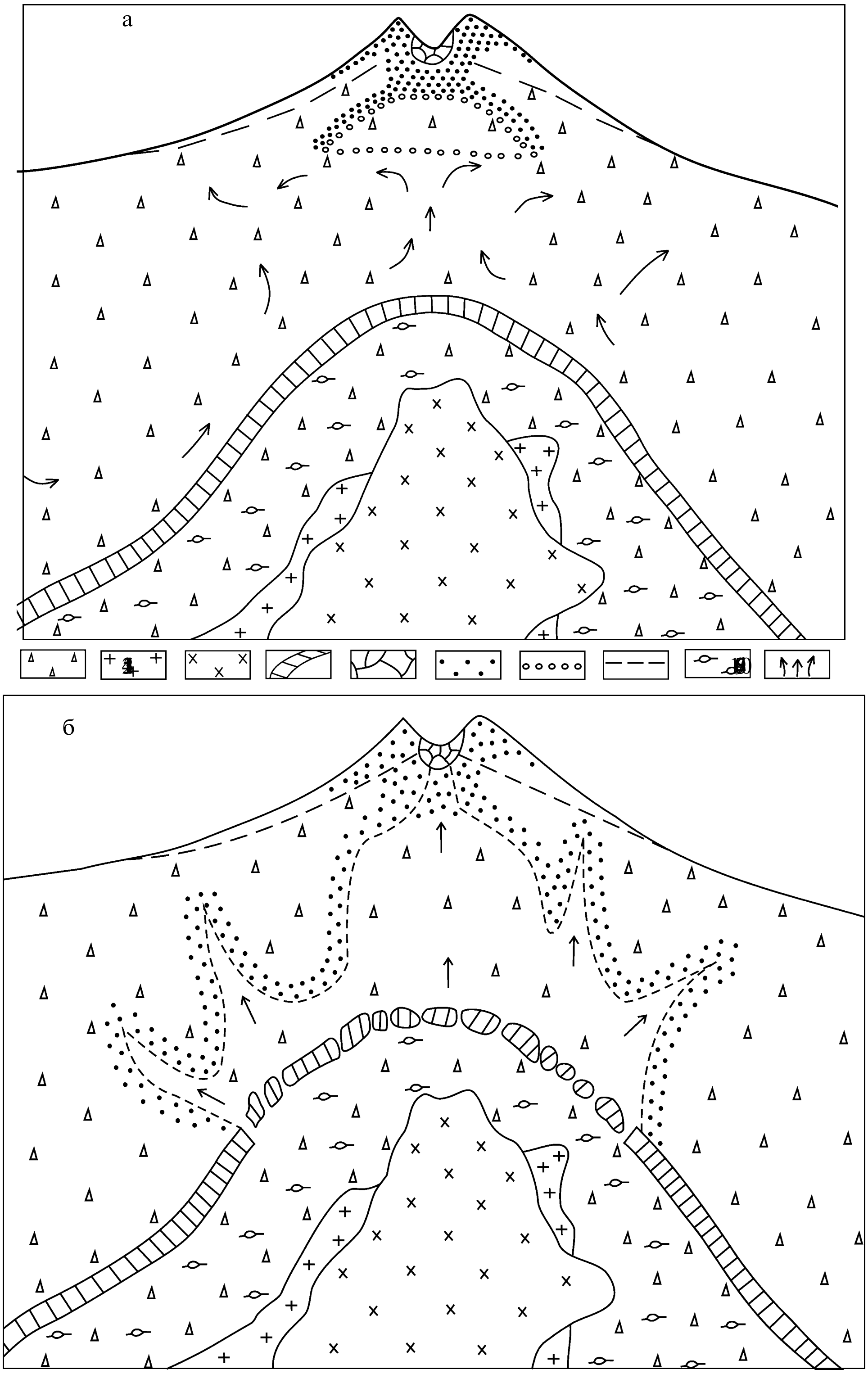

Fig. 7. A diagrammatic model of transition from magmatic to epithermal (hydrothermal) conditions in subvolcanic structures near the top of emplaced plutons [Fournier, 1999]. @a transition from brittle to plastic medium in the temperature range between 370°C C and 400°C, b episodic and temporary penetrations of magmatic fluids that come to the hydrothermal system above.

(1) volcanogenic-sedimentary host rock sequence, (2) first phase of the intrusion, (3) second phase of the intrusion, (4) zone of brittle/plastic transition, (5) volcanic breccia in the zone of boiling hydrothermal fluids, (7) boundary of this boiling zone, (8) piezometric level of hydrothermal solutions, (9) supra-intrusion zone saturated with steam and mineralized water (brine), (10) directions of fluid flows and water flows.

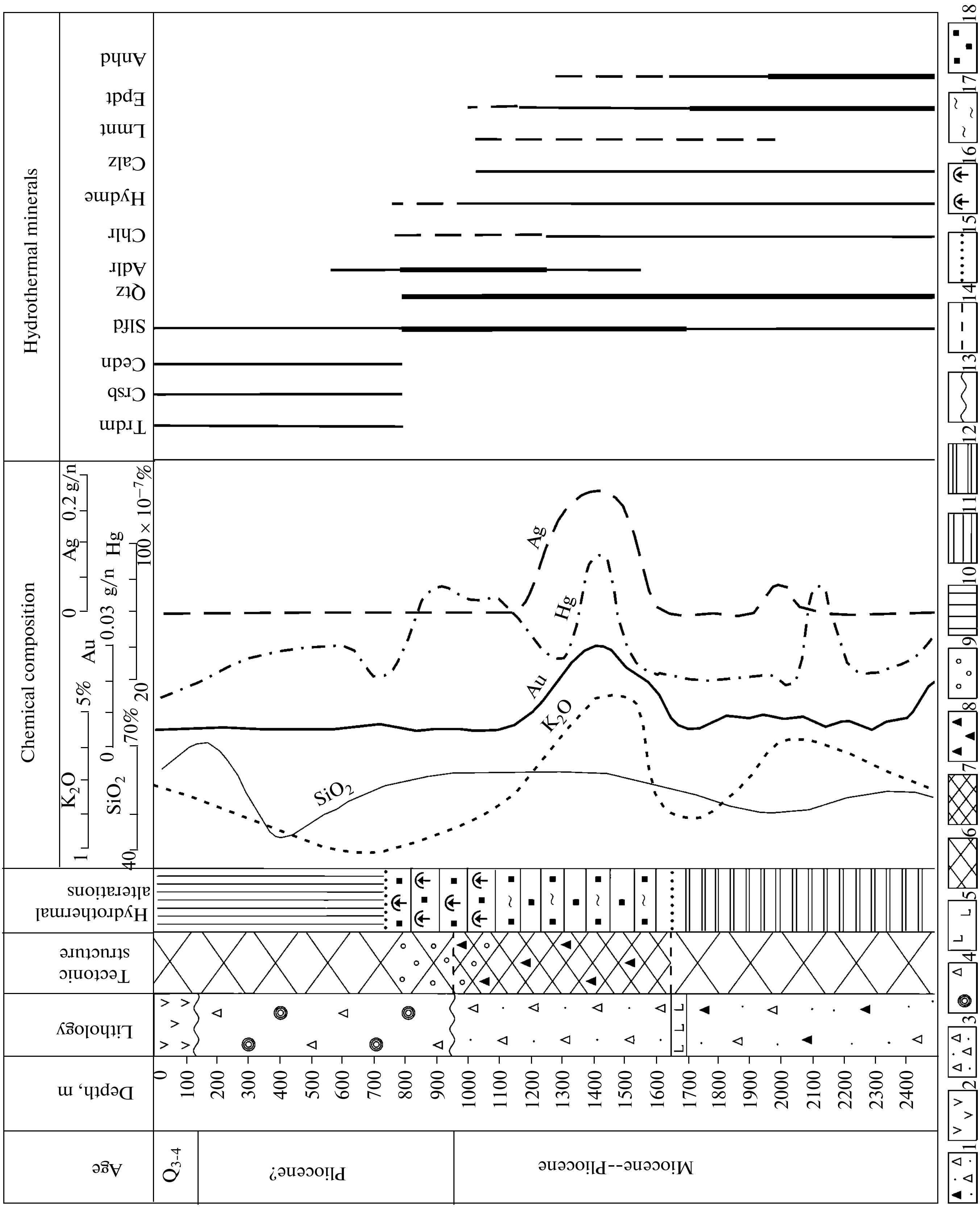
Hydrothermal minerals
Anhd
Epdt
Lmnt
Calz
Hydme
Chlr
Adlr
Qtz
Slfd
Cedn
Crsb
Trdm
Chemical composition
Ag 0 0.2 g/n
Hg 0.03 g/n
Au 20 100 × 10−7%
K2O 1 5%
SiO2 40 70%
Ag
Hg
Au
K2O
SiO2
Hydrothermal alterations
Tectonic structure
Lithology
Depth, m
Age
Q3-4
Pliocene?
Miocene--Pliocene

Fig. 8. Deep geological and geochemical section of the North Paramushir hydrothermal magmatic system, column of well GP-3 [Rychagov et al., 2002].

(1) lithocrystalloclastic, psepho–psammitic intrusive tuffs (intrusive breccias), (2) andesitic lavas, (3) lithocrystalloclastic, psepho-psammitic, variegated tuffs, (4) fine-fragmental tuffites with relict microfauna forms, (5) gabbro–porphyrites (a flattish body), (6) tectonic fissuring of rocks, (7) intensive tectonic fissuring, (8) tectonic breccias, (9) zone with high open porosity, (10) zone of low-temperature, opal–cristobalite–tridymite–chalcedony–quartz mineralization involving hydromica, (11) medium-temperature, quartz–chlorite–epidote–muscovite propylite, (12) secondary quartz–adularia–hydromica quartzites, (13–15) bounbdaries: (13) lithologic, (14) tectonic, (15) metasomatic, (16) zone of liquid–steam transition with quartz–adularia metasomatites, (17) quartz–chalcedony streaks, (18) ore mineralization.

penetrates from the plastic to the brittle medium, reduction in the dissolvability of fluid components, and the deposition of material around the magma chamber and at the boundary of brittle/plastic deformation (Fig. 5), just as it occurs in the earth. The above versions of the model cover the case of one-pulse emplacement of magma into the reservoir. Multi-pulse emplacement possibly involves a superposition of these scenarios.

ELASTO-PLASTIC TRANSITIONS IN THE STRUCTURE OF ISLAND-ARC HYDROTHERMAL MAGMATIC SYSTEMS

Drilling to depths of 5 km within long-lived volcanogenic ore centers, geothermal areas, and on active andesitic volcanoes situated at the Kuril–Kamchatka, Japan, Indonesian and other island arcs in the Pacific yielded reliable thermodynamic and physico-chemical parameters peculiar to the zone of interaction between hydrothermal systems and magma intrusions. Especially interesting results for estimating the interaction between themselves and the influence of the conductive heat flow and convective magmatic and hydrothermal cells on the physico-chemical properties of host rocks were obtained during exploration and extraction of geothermal and epithermal gold–base-metal and Au–Ag–Cu–Mo–... porphyry fields at the Philippine island arc [Corbett and Leach, 1998]. In spite of the fact that the poly-phase intrusion is emplaced within the upper crust there, one can see scenarios 1 and 2 in action, possibly also scenario 3, owing to the long-continued evolution of a magma body with a complex structure (Fig. 6). Under island-arc conditions, a powerful conductive heat flow is provided by the occurrence of mantle plumes, while the convective flows are produced by major poly-phase intrusions (ultramafic to acidic in composition) that are emplaced into the lower crust and by their subsequent long-continued evolution. A zone of brittle/plastic transition arises in the apical parts of subvolcanic magma bodies (depths of 1–3 to 5–7 km have been studied) at contacts with host rocks (Fig. 7). That zone has a complex structure owing to interaction of the magmatic fluid with hydrothermal solutions and host alumosilicate rocks, the generation of brines and zones of cooling magma, the generation of breccia bodies and of a special type of hydrothermal metamorphic alteration, etc. The magmatic system periodically interacts with the hydrothermal system owing to physical fracturing that affects the zone of brittle/plastic transition.

We showed previously that the evolution of an island-arc ore-forming hydrothermal magmatic system involves self-isolation of the system from other regional geologic structures around it owing to outward transport of many major and trace components from the zone of ascending heat flows toward the boundaries of the system [Belousov et al., 1998; Rychagov, 2003]. The components that are actively involved in the self-isolation of a hydrothermal magmatic system include colloidal compounds of silicic acid and the silica minerals that derive from these, viz., opal, tridymite, cristobalite, chalcedony, and quartz. It is this mechanism of water–rock interaction that is realized in the geothermal processes that are occurring in the interior of the North Paramushir hydrothermal magmatic system [Rychagov et al., 2002]. One of the geologic blocks of this structure was penetrated by the GP-3 well to a depth of 2.5 km. A zone of boiling metal-bearing hydrothermal fluids formed above the intrusive body with peculiar geochemical and mineralogical ore zonalities. The boiling of hydrothermal fluids results in the precipitation of silicate gel in porous volcanogenic–sedimentary rocks, with the gel subsequently crystallizing to produce silica minerals (Fig. 8). The latter process gave rise to a secondary (an additional upper aquiclude and heat-insulating) horizon up to 300–500 m and greater in thickness. Such a horizon which is found in a block of the North Paramushir geothermal magmatic system, is 750 m thick [Rychagov et al., 2002]. The zone of brittle/plastic transition from the cooling magma body to host rocks is also the so-called "breccia mantle," which is a set of various breccia formations (from intrusive and automagmatic to hydrothermal and metasomatic breccias) in endo-contact and exo-contact parts of the intrusions (Fig. 9). The visible thickness of the breccia mantle in this geologic feature is 400–500 m, but it can also be as thick as 800–1000 m [Rychagov, 1989]. We studied the Vychenkiya, southern Kamchatka paleovolcanic complex to map transitions from intrusive

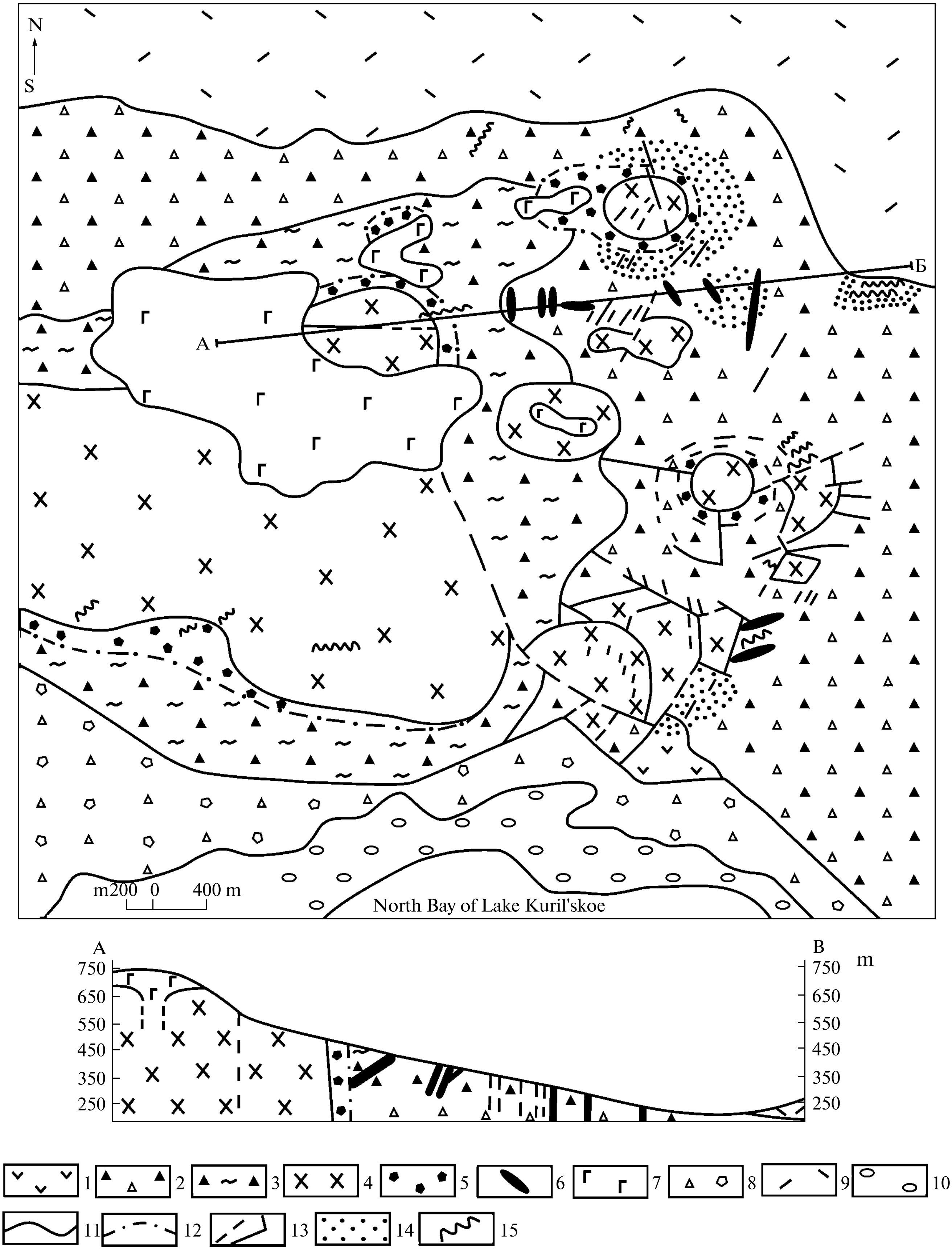
N
S
A
Б
m200 0 400 m
North Bay of Lake Kuril'skoe
A
B
m
750
650
550
450
350
250
750
650
550
450
350
250
1
2
3
4
5
6
7
8
9
10
11
12
13
14
15

Fig. 9. Structure of the Vychenkiya, southern Kamchatka ore-bearing volcano-plutonic complex.
(1) andesite lavas at the bottom of the Paleogene–Early Miocene(?) section, (2) andesite and dacitic andesite tuffs and welded tuffs, (3) sequence of interstratified, acidic lavas, welded tuffs, tuffs, and ignimbrites, (4) subvolcanic basaltic andesites, (5) intrusive breccias: giant-, coarse- and fine-fragmental auto-magmatic breccias with fragments and blocks of host rocks, (6) andesite and basaltic andesite dikes, (7) extrusive lava complex of dacitic andesites from the Pliocene (?) phase of magmatism, (8) Quaternary tuff conglomerates, (9) Holocene pumice deposits, (10) rewashed pumice deposits and lacustrine (lacustrine–marine?) sand, (11) stratigraphic and intrusive boundaries, (12) boundary to the breccia mantle in intrusive bodies, (13) tectonic faults, inferred (dashed) and certain (continuous line), (14) fields of hydrothermally altered rocks, (15) quartz–ore zones and veins.

bodies via various "intermediate" rock types to ore-forming structures at different erosional truncations. The depth of erosional truncation for this geologic structure is estimated here as 800–1000 m for the southwestern part and 300–500 m for the northeastern and eastern parts of the complex [*Struktura* ..., 1993].

CONCLUSIONS

(1) Special rheologic conditions are created in the structure of long-lived ore-forming hydrothermal magmatic island-arc systems in the form of elasto-plastic transitions in the geological medium. The elastic-to-plastic transitions regulate the generation of hydrothermal magmatic systems, their supply sources (magma chambers and intrusive bodies), and the occurrence of physico-chemical processes in the transition zone between a near-surface hydrothermal and a deep-seated magmatic convective cell.

(2) Two scenarios (models) were suggested for the generation of hydrothermal magmatic circulation systems to incorporate the location of the magmatic heat source in the plastic or brittle zones of the Earth's crust. The ascent of magma melt and fluids along vertical fissure systems occurs, when an excess pressure is available in the head parts of the fissures. The excess pressure arises from the difference in the densities of the fluid (or magma) and the host rocks. It depends on the vertical extent of cavities that are filled with fluid or magma [Zhatnuev, 2005]. The transition from lithostatic to hydrostatic pressure as a fissure penetrates a brittle rock results in an adiabatic expansion of the fluid, a sharp decrease in its density, and the cooling of the solutions (often accompanied by boiling), as well as in mineral generation. For this reason the transition zones from lithostatic to hydrostatic pressure can be recognized from intensive secondary mineral generation, in particular, silicification and filling of the fissure pore space with tridymite, cristobalite, opal, chalcedony, and quartz in association with adularia. The quartz–adularia metasomatites and argillization zones contain gold–sulfide mineralization, native metals and intermetallic compounds.

(3) Multidisciplinary geological, geophysical, hydrogeological, mineralogical, geochemical, and other research in the hydrothermal magmatic systems in southern Kamchatka and on Paramushir Island demonstrated the high consistency between the model that envisages the emplacement of magma melts into a brittle crust and the structure of present-day, high-temperature, hydrothermal magmatic systems (at the progressive phase of their evolution) at depths of over 1.0–1.5 km and Miocene to Pliocene ore-bearing volcano-plutonic complexes that have been eroded to different depths in different geologic blocks. The part of the brittle/plastic transition zone in the structure of volcano-plutonic complexes is played by the breccia mantle of intrusive gabbro–diorite bodies, which controls the flows of high-temperature metal-bearing gas–water fluids.

ACKNOWLEDGMENTS

We express our deepest gratitude to academician S.A. Fedotov who expressed interest in this paper and discussed models for the mechanisms that lift basaltic melt from deeper crustal horizons to the ground surface.

This work was supported by the Russian Foundation for Basic Research (project nos. 09-05-00022a and 10-05-00009a), by the Presidium of the Siberian Branch of the Russian Academy of Sciences (project no. 117), and by the Presidium of the Far East Branch of the Russian Academy of Sciences (project nos. 09-II-SO-08-006 and 09-II-SO-08-004).

REFERENCES

Aver'ev, V.V., The Hydrothermal Process in Volcanic Areas and Its Relationship with Magmatic Activity, in *Sovremennyi vulkanizm* (Present-Day Volcanism), Moscow: Nauka, 1966, pp. 118–128.

Bazylev, B.A., Metamorphism of Hyperbasites from Atlantis Fracture Zone, the Atlantic Ocean: Evidence for Deep Penetration of Water into Oceanic Lithosphere, *Dokl. Akad. Nauk SSSR*, 1992, vol. 323, no. 4, pp. 741–743.

Belousov, V.I., Rychagov, S.N., Kuz'min, Yu.D., et al., Silica in High-Temperature Hydrothermal Systems in Areas of Present-Day Volcanism, *Ekologicheskaya Khimiya*, 1998, vol. 7, no. 3, pp. 200–216.

Corbett, G.J. and Leach, T.M., Southwest Pacific Rim Gold–Copper Systems: Structure, Alteration and Mineralization, *Special Pub. Society of Econ. Geol. Ins.*, 1998, no. 6.

Efimov, A.B. and Ershova, T.Ya., Thermodynamics around a Magma Conduit, *Volcanology and Seismology*, 1999, vol. 20, nos. 4/5, pp. 479–496, Gordon and Breach Science Publishers.

Fedotov, S.A., Major Fissure Eruptions: A Theory. The Mechanism of the Great Tolbachik Fissure Eruption, in *Bol'shoe treshchinnoe Tolbachinskoe izverzhenie, 1975-*

1976 gg., Kamchatka (The Great Tolbachik Fissure Eruption, 1975–1976, Kamchatka), Ch. XVI, Fedotov, S.A., Ed., Moscow: Nauka, 1984, pp. 537–575.

Fournier, R.O., Hydrothermal Processes Related To Movement of Fluid from Plastic Into Brittle Rock in the Magmatic–Epithermal Environment, *Economic Geology,* 1999, no. 94, pp. 1193–1212.

Fyfe, W.S., Price, N.J., and Thompson, A.B., *Fluids in the Earth's Crust: Their Significance in Metamorphic, Tectonic and Chemical Transport Processes*, Amsterdam: Elsevier, 1978.

Gidrotermal'nye sistemy i termal'nye polya Kamchatki (Hydrothermal Fields and Thermal Fields of Kamchatka), Sugrobov, V.M., Ed., Vladivostok: DVNTs AN SSSR, 1976.

Ivanov, S.N., The Limiting Depth of Open Fissures and the Hydrodynamical Zonality of the Earth's Crust, in *Ezhegodnik-1969,* (Annual Issue-1969), Sverdlovsk: Izd-vo In-ta geologii i geokhimii UF AN SSSR, 1970, pp. 212–233.

Ivanov, S.N., Zones of Plastic and Brittle Deformation in the Vertical Section of the Lithosphere, *Geotektonika*, 1990, no. 2, pp. 3–14.

Khristoforova, N.N., Khristoforov, A.V., and Muslimov, R.Kh., Low-Density Zones in the Crystalline Basement, *Georesursy*, 1999, no. 1(1), pp. 4–15.

Kissin, I.G., On the Sources and Migration Pathways of Fluids That Are Involved in the Generation of Conductive and Low Velocity Crustal Zones, *Dokl. Akad. Nauk*, 2001, vol. 380, no. 6, pp. 800–804.

Kol'skaya sverkhglubokaya. Issledovanie glubinnogo stroeniya kontinental'noi kory s pomoshch'yu bureniya Kol'skoi sverkhglubokoi skvazhiny (The Kola Deep Well: Studies of Deep Crustal Structure by Drilling the Kola Deep Well) Kozlovskii E.A., Ed., Moscow: Nedra, 1984.

Kononov, V.I., *Geokhimiya termal'nykh vod oblastei sovremennogo vulkanizma (riftovykh zon i ostrovnykh dug)* (The Geochemistry of Thermal Waters in Areas of Present-Day Volcanism: Rift Zones and Island Arcs, *Trudy GIN*, no. 379, Moscow: Nauka, 1983.

Nikolaevskii, V.N., Crustal Faults and Tectonic Waves, *Elektronnyi nauchno-informatsionnyi zhurnal "VECTNIK OGGGN RAN,"* 2001, no. 1(16). www.scgis.ru/russian/cp1251/h_dgggms/1-2001/nikolaevsky.htm#begin

Pavlenkova, N.I., Crustal and Upper Mantle Structure and the Mechanism That Drives the Motion of Deep-Seated Material, *Elektronnyi nauchno-informatsionnyi zhurnal "VECTNIK OGGGN RAN,"* 2001, no. 4(19). www.scgis.ru/russian/cp1251/h_dgggms/4-2001/pavlenkova.htm#begin

Rychagov, S.N., *Brekchievaya struktura geologicheskoi sredy* (The Breccia Structure of the Geologic Medium), Available from VINITI, Petropavlovsk-Kamchatskii, 1989.

Rychagov, S.N., *The Evolution of Hydrothermal Magmatic Systems at Island Arcs*, Extended Abstract of D-r Sci. (Geol.–Mineral.) Dissertation, IGEM RAN, Moscow, 2003.

Rychagov, S.N., Belousov, V.I., Glavatskikh, S.F., et al., The North Paramushir Hydrothermal Magmatic System: The Deep Geological Section and a Model of Present-Day Mineralization in Its Interiors, *Vulkanol. Seismol.*, 2002, no. 4, pp. 3–21.

Struktura gidrotermal'noi sistemy (The Structure of Geothermal Systems), Moscow: Nauka, 1993.

Vashchilov, Yu.Ya., *Blokovo-sloistaya model' zemnoi kory i verkhnei mantii* (A Block-Layered Model of the Earth's Crust and Upper Mantle), Moscow: Nauka, 1984.

Zhatnuev, N.S., Fissure Fluid Systems in a Zone of Plastic Deformation

DAN, 2005, vol. 404, no. 3, pp. 380–384.

Zhatnuev, N.S., The Dynamics of Deep-Seated Magmas, *DAN*, 2010, vol. 430, no. 6, pp. 787–791.

Zhatnuev, N.S., Mironov, A.G., Rychagov, S.N., and
Gunin, V.I., *Gidrotermal'nye sistemy s parovymi rezer-* 1
vuarami (Hydrothermal Systems with Steam Reservoirs), Novosibirsk: Izd-vo SO RAN, 1996.

SPELL: OK