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DEPTH OF MAGMA CHAMBER DETERMINED BY EXPERIMENTAL PETROLOGIC METHODS

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ABSTRACT

Depth of magma chamber (or young intrusive body) is one of the most important parameters that determine the thermal structure and the circulation pattern of fluid around a geothermal area. The depth is also important when we recognize the magma chamber as a deep-seated geothermal resource itself. Here, we introduce an experimental petrologic method for determining the depth of magma chamber. There are several methods in order to determine the depth (pressure) of a magma chamber. Among them, a melting experiment for the rock erupted from the magma chamber is the most reliable method. The method consists of the following procedure: (1) Select a sample (volcanic rock) which is representative of magma in the chamber; (2) Give a petrographic description of the sample; (3) Try to reproduce the petrographic description experimentally by melting and re-equilibrating the sample at various conditions (pressure, temperature, oxygen fugacity, etc.) using a high-temperature, high-pressure apparatus. If the trial results in success, the conditions of the experiment that reproduces the petrographic description are considered to be those of the magma chamber. In this way, we determined the depth of magma chamber beneath Usu Volcano, Japan. We selected Us-b pumice (the major product of the 1663 A.D. eruption; homogeneous rhyolitic pumice with small amount of phenocrysts) as the sample. We conducted a series of melting experiments for the sample using internally heated pressure vessels. As a result, the petrographic description (pl+opx+mt/ilm, 95% melting, An=42-44, about 780 °C) of the pumice was reproduced at about 250 MPa. Thus, the depth of the magma chamber at the 1663 A.D. eruption is estimated to be about 10 km. We also determined the depth at other eruptions, using the dependence of plagioclase composition on temperature and water pressure. After detailed petrographic descriptions for the eruptive products and application of our experimental results, we revealed that the 250-MPa magma chamber still existed after the 1663 A.D. eruption and that another magma chamber formed at about 4 km in depth after the 1663 A.D. eruption. Using the above method, we can estimate depths of magma chambers (or intrusive bodies) beneath volcanoes where their eruptive products are available.

1. INTRODUCTION

Depth of magma chamber (or young intrusive body) is one of the most important parameters that determine the thermal structure and the circulation pattern of fluid around geothermal area (Norton and Knight, 1977; Hanson, 1992; Tomiya, 1995). The depth is also important when we recognize the magma chamber as a deep-seated geothermal resource itself. From a petrologic point of view, depth (pressure) of magma chamber is also a key parameter for magma evolution because it determines the species of minerals crystallizing from the magma and, therefore, the composition of the residual melt, that is, the evolved magma.

Thermodynamic conditions of a magma chamber can be estimated using petrographic information (e.g., compositions and modal fractions of phenocrysts and groundmass) from a representative rock erupted from the chamber. In order to estimate the pressure of magma chamber, the following methods are generally used; (1) geobarometer, (2) water content, and (3) melting experiment. One of the examples of the first method is an amphibole geobarometer (e.g., Hammarstrom and Zen, 1986) where pressure is estimated from the aluminum content of the amphibole crystallized from the magma. This method, however, is difficult to apply to natural magma because the geobarometer should be applied only when the mineral assemblage of the magma is the same as that of experimental products for which the geobarometer was calibrated. The second method is to measure water (and carbon dioxide) contents of melt inclusions and to estimate the saturation pressure (e.g., Wallace et al., 1995). This method is simple but there is a problem that the time of the formation of melt inclusions is often unknown. The last method is the most reliable one and will be explained in the next chapter in detail.

In this paper, an example of estimation of the depth of magma chamber by melting experiment is shown. The studied field is Usu volcano, one of the most active volcanoes in Japan. The starting material we used is Us-b pumice, the largest plinian product of the volcano. The evolution of the magma-feeding system will be also discussed.

2. THE MELTING EXPERIMENT FOR AN ERUPTIVE PRODUCT FROM USU VOLCANO

2.1 The Melting Experiment

The most reliable method in order to estimate the depth (pressure) of a magma chamber is to conduct a series of melting experiments with a representative rock erupted from the chamber. This method is based on the following assumptions; (1) the sample rock selected for the experiment was in chemical equilibrium in the magma chamber; (2) the phenocrysts were the solid phase and the groundmass was the melt phase in the magma chamber, respectively. In each experiment, the sample is kept at some conditions (temperature, pressure and oxygen fugacity) and chemically re-equilibrated. If the conditions are different from those of the magma chamber, the run product will present different features (compositions of crystals and melt, degree of partial melting, etc.) from the natural rock. However, if the conditions are identical to those of the magma chamber, the run product will present the same features as the natural rock, and we can safely conclude that the experimental conditions are the same as those of the magma chamber (e.g., Rutherford et al., 1985).

The procedure of the experiments is as follows.

Preparation:

- Grind the rock to powder less than 30 microns.
- Dry out the powder overnight with furnace at 1 atm, 900 °C, and about QFM (Quartz - Fayalite - Magnetite) buffer.
- Pack the weighed powder (typically 20 mg) and weighed water (typically 2 mg) in a capsule (See section 2.3 for detail of the capsule.) and seal it.

Experiments:

- Set the capsule into a high-pressure apparatus. (See section 2.2 for detail of the apparatus.)
- Keep the capsule at desired conditions for a period of several tens to hundreds hours (dependent on experimental temperature) and then quench it. Such an experiment is conducted for various conditions; in this study, 100 to 400 MPa in pressure, 700 to 1000 °C in temperature, and NNO (Ni metal - NiO) to MNO (MnO - Mn_3O_4) in oxygen fugacity. In most of the experiments, preheating was applied where the capsule was kept at the temperature of 25 °C higher than the desired one for several hours before the main heating.

Analyses:

- Check the processed capsule for the following points: the capsule is not damaged; both the buffering materials and water are not exhausted; and the water does not escape from the capsule (the weight of the capsule is not changed) during the run.
- Polish a cross section of the capsule and analyze it by electron microprobe for the texture and the chemical composition of the crystals and the glass (melt).

Before this procedure, detailed petrographic study for the rock (the starting material for the experiment) must be done to investigate whether it is suitable for the experiment. (See section 2.4 for detail of the petrographic description.) This is indeed the critical point to have a success of the experiment because if the rock is not representative of the magma that was in chemical equilibrium in the magma chamber, the experimental results will be meaningless. In this sense, our sample (Us-b pumice) is suitable (see section 2.4).

2.2 The High-Pressure Apparatus

Our experiments were conducted with two types of internally heated pressure vessels. For the 100-MPa and 200-MPa experiments, an internally heated pressure vessel at the Tokyo Institute of Technology, Japan, was used. This apparatus has a drop-quenching system, with which sample capsules are dropped from heated area into cold area at the end of each run so that the samples are quenched at a cooling rate of about 100 °C/sec (Takahashi and Tomiya, 1992; Holloway et al., 1992). The 400-MPa experiments were carried out with an internally heated pressure vessel at the Chiba University, Japan. This apparatus does not have the drop-quenching mechanism, so that samples are cooled by turning off the furnace at the end of each run. Typical cooling rate is about 10 °C/sec in this apparatus. Even at this cooling rate, however, no quenched crystals appeared in our run products because of high silica content of our starting material (74 wt.%; Ôba et al., 1983). For both pressure vessels, Ar was used as a pressurizing gas medium.

2.3 The Sample Containers

The sample was packed in a noble-metal capsule in each experiment. We used both oxygen-buffered capsules (e.g., Eugster and Skippen, 1967) and oxygen-unbuffered capsules. The NNO oxygen-buffered experiments were performed with a crimped Ag70-Pd30 inner tube (2.3 mm in diameter), in which the sample powder and excess water was contained, inside a sealed Pt (3.0 mm in diameter) and a sealed Au outer tube (5.0 mm in diameter), in which the inner tube and the buffer material were contained (triple capsule). The MNO oxygen-buffered experiments were performed with a sealed Ag70-Pd30 inner tube and a sealed Au outer tube (double capsule). Oxygen-unbuffered experiments were performed with single capsules of Au or Ag70-Pd30 (2.3 mm in diameter).

2.4 The Sample ~ Usu Volcano and Us-b Pumice

We selected Usu volcano as a study field. It is one of the most active volcanoes in Japan, and is widely known by the spectacular eruption of Showa-Shinzan lava dome (1943-45 A.D.). The volcano has been geologically well studied (e.g., Ôba, 1966; Ôba et al., 1983). Recently its magma-feeding system is being revealed by detailed petrographic studies (e.g., Tomiya and Takahashi, 1995a).

As a starting material for the melting experiment, Us-b pumice was selected. It is the product of the large plinian eruption in 1663 (2 km^3 in tephra volume or 0.3 km^3 in dense rock equivalent; Ôba and Kondo, 1964). The pumice is the largest and the first product of the silicic stage of the volcano, which includes seven eruptions (1663, 1769, 1822, 1853, 1910, 1943-45, 1977-78 A.D.) and started at 1663 A.D. after a long (thousands of years) dormancy. The SiO_2 content of the pumice, 74 wt.% (Table 1; Ôba et al., 1983), is the highest value in the volcano. Petrographic description of the pumice was done by Okumura et al. (1981) and Ôba et al. (1983), and recently revised by Tomiya and Takahashi (1995a). Their descriptions are summarized as follows. The phenocryst content is only about 5 vol.%, that is, the melt fraction of the magma is about 95 %. Most of the phenocrysts are plagioclase (pl) of An = 42 ~ 44 (An = 100*Ca/(Ca+Na)). The others are orthopyroxene (opx), magnetite (mt) and ilmenite (ilm). Because these phenocrysts are quite homogeneous (no chemical zoning), we conclude that the magma was in chemical equilibrium in the magma chamber. The temperature and oxygen fugacity (fO_2) of Us-b pumice are estimated with the Fe-Ti oxide geothermometer by Ghiorso and Sack (1991) to be T = 780 °C and $\log_{10}fO_2 = -15.0$, respectively.

3. THE DEPTH OF USU MAGMA CHAMBER

In order to determine the depth of the Usu magma chamber using the experimental results, we use two types of approaches. One is the use of phase diagram (section 3.1) and the other is the use of plagioclase composition (section 3.2). In any case, we investigate whether the petrographic features of Us-b pumice (section 2.4) are represented in the run products.

3.1 Result From Phase Equilibria

At first, we use the phase diagram of Us-b pumice, obtained by the experiments (Fig. 1a), where the beginning of crystallization of each solid phase (pl: plagioclase, opx: orthopyroxene, qz: quartz, hb: hornblende, mt: magnetite/

ilmenite) is shown by solid curve. Because the fO_2 of Us-b pumice ($\log_{10}fO_2$ = -15.0 at 780 °C) is near the NNO buffer, only the results of NNO-buffered runs are shown here. (See Tomiya (1999) for the effect of fO_2 on the phase equilibria.)

The pressure of the magma chamber where the Us-b magma was in equilibrium is determined as follows. Because its temperature is 780 °C and its phenocryst assemblage is plagioclase + opx + Fe-Ti oxides (no quartz), the water pressure is within a range of from 150 to 250 MPa (Fig. 1a). In addition, because its melting degree is about 95 % (Fig. 1b), the water pressure is determined to be close to 250 MPa. Since a magma chamber just prior to a plinian eruption is thought to be saturated with water (Wallace et al., 1995), the water pressure is considered to be equal to the total pressure. Accordingly, the pressure corresponds to about 10 km in depth, assuming the average density of the crust is 2500 kg/m^3.

3.2 Result From Plagioclase Compositions

Second, we use the plagioclase composition (An content), which is quite sensitive to water pressure (e.g., Kudo and Weill, 1970; Housh and Luhr, 1991).

The dependence of An content on temperature and water pressure, obtained by our experiments, is summarized in Fig. 2. Since the pl phenocryst in Us-b pumice crystallized at 780 °C and its composition is An = 42 ~ 44, the water pressure of the Us-b magma was about 200 MPa. This is consistent with the estimate of 250 MPa from the phase equilibria (Fig. 1a).

3.3 About the Other Eruptions Following the Us-b eruption

The estimated depth (about 10 km) of the Usu magma chamber is the value at the time of the 1663 Us-b eruption. The depths at the other (following) eruptions can also be estimated by the experimental results.

We estimated the depth of the magma chamber(s) at the following eruptions, using the compositions of magnetite, ilmenite and plagioclase as follows. First, using the compositions of magnetite and ilmenite, we estimate the magmatic temperature by the Fe-Ti oxide geothermometer. Second, we measure the composition (An) of plagioclase that was in equilibrium with the Fe-Ti oxides. From the temperature and the An, we can estimate magmatic pressure using Fig.2. (Because the melt compositions of Us-b and other eruptive products are similar (Tomiya and Takahashi, 1995a), Fig.2 can be used with consistency.) We note that in this method we must use only pumice which was a quenched droplet of magma during its eruption. On the other hand, lava flows (domes) should not be used because they were cooled slowly during and after their eruption so that chemical diffusion and overgrowth of crystals modified the petrographic features from those of the magma chamber. This modification is significant for magnetite and ilmenite because their diffusion coefficients are large.

In this way, we estimated the depth at three eruptions (1769, 1822, 1853). The depth at the other eruptions was not estimated because of the lack of ilmenite (1977), the lack of pumice (1943-45), and the lack of essential product (1910). Each product (pumice) of the three eruptions (1769, 1822, 1853), except for the 1769 one, shows evidences that just prior to the eruption mixing of dacitic and rhyolitic magmas occurred (Tomiya and Takahashi, 1995b). For example, there are two types of compositions of magnetite/ilmenite phenocrysts, that is, dacitic type and rhyolitic type. Using the Fe-Ti oxide geothermometer, we estimated temperature of each type of magma. In addition, we distinguished the two types of plagioclase which coexisted with the two types of magnetite, respectively, using aggregates of magnetite and plagioclase phenocrysts. Accordingly, we obtained two sets of temperature and plagioclase composition from each eruptive product. Plotting these results on Fig.2, we concluded that the pressures of the rhyolitic-type magmas are identical to Us-b, about 250 MPa, and that those of the dacitic-type magmas are about 100 MPa. The latter pressure (100 MPa) corresponds to about 4 km in depth.

4. DISCUSSION

We found that beneath Usu volcano there are two magma chambers, the deeper (rhyolitic) one at about 250 MPa (10 km) and the shallower (dacitic) one at about 100 MPa (4 km) (Fig. 3). On the basis of the following discussion, we propose that the deeper one is a main chamber and the shallower one is a sub chamber in this volcano. First, the deeper one had already existed prior to the 1663 (Us-b) eruption, but, on the other hand, the shallower one formed after the 1663 eruption. Second, Tomiya and Takahashi (1995a) concluded that prior to the 1663 eruption there was a basaltic magma just below the rhyolitic (Us-b) magma, and that the dacite magmas erupted after Us-b were formed by mixing of the rhyolitic and the basaltic magmas. Third, the volume of the Us-b is much larger than those of the other products.

Takada (1994) discussed the depth of silicic magma chamber by compilation of existing data and by physical modeling of magma ascending. He suggested that there are two levels for silicic magma chambers; the deeper one to be the depth where the silicic magma was generated and the shallower one to be the depth where the densities of the silicic magma and the crust were equal (the level of neutral buoyancy). This may also be the case in Usu volcano.

From a geothermal point of view, the deeper chamber may be difficult to exploit. On the other hand, the shallower chamber beneath Usu volcano will be available in the near future. To find such shallow (several kilometers) magma chambers will be important at that time.

5. CONCLUSIONS

We estimated the depth of Usu magma chamber by a series of melting experiments for Us-b pumice. At the time of the 1663 (Us-b) eruption, the depth of the chamber was 10 km. After the eruption, a new magma chamber formed at the depth of about 4 km. Since then, beneath Usu volcano there have been two magma chambers, the deeper rhyolitic chamber and the shallower dacitic chamber.

We can estimate by this method the depths of magma chambers (or intrusive bodies) beneath volcanoes where their eruptive products are available.

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Table 1. Chemical compositions of Us-b pumice (Ôba et al., 1983). Recalculated in order that the non-volatile total is 100 wt. %.

(wt. %)	SiO_2	TiO_2	Al_2O_3	Fe_2O_3	*FeO*	*MnO*	*MgO*	*CaO*	Na_2O	K_2O	P_2O_5	*Total*
Us-b	73.89	0.21	14.78	0.65	1.53	0.16	0.26	2.26	4.93	1.23	0.10	100.00

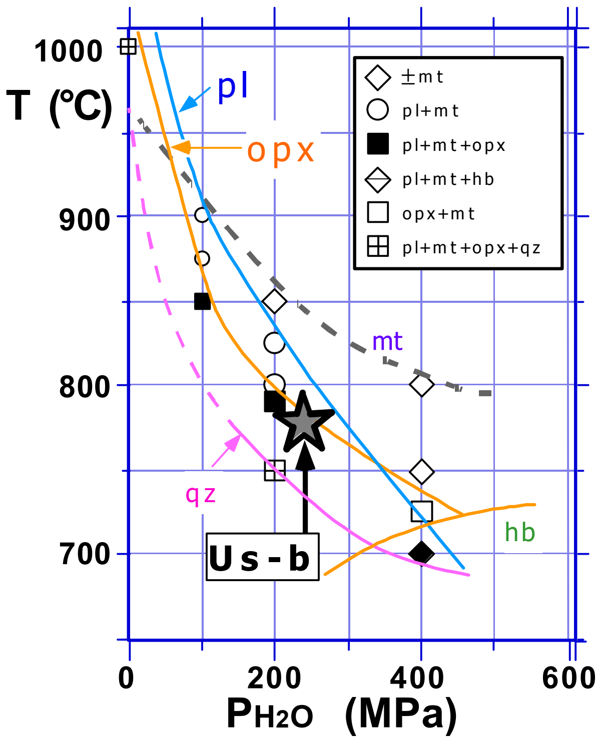

Figure 1a. The phase diagram for Us-b pumice bulk composition, obtained by our experiments (NNO buffered runs). The beginning of crystallization of each solid phase is shown by a solid curve. The obtained Us-b condition is indicated by the star (T = 780 °C, P_{H2O} = 250 MPa).

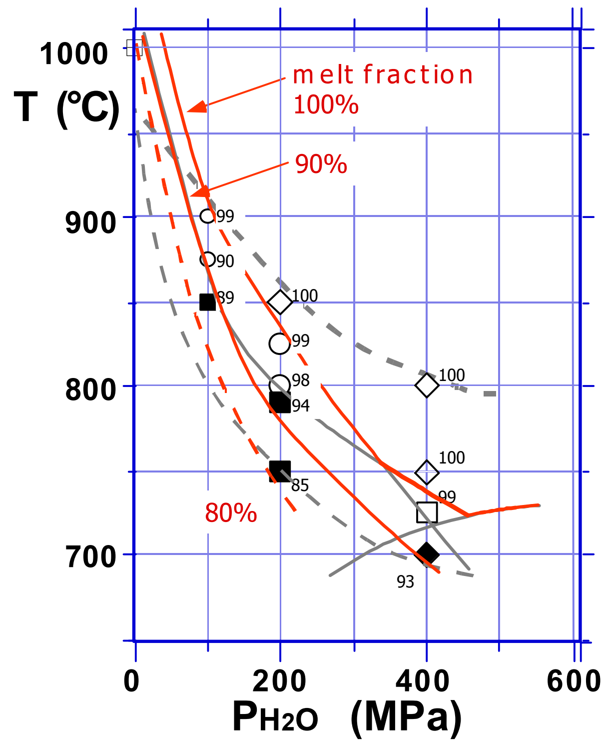

Figure 1b. The melt fractions for the run products. As the "100 % melt" curve, the beginning of silicate crystallization is used and the small amount of mt crystallization (order of 0.1 %) on the curve is ignored. The beginnings of other crystallization are also shown (gray curves). Note that the melt fraction of Us-b is 95 %.

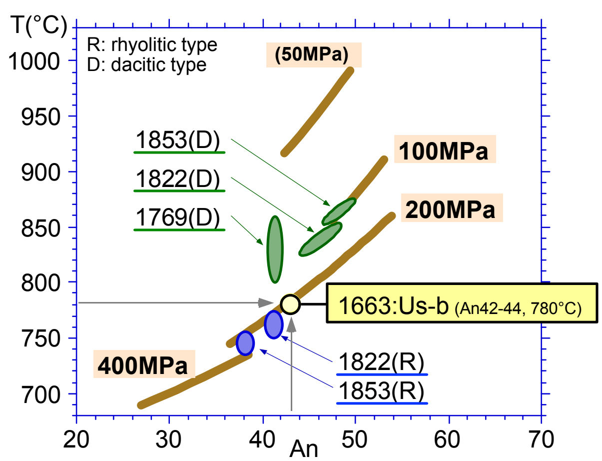

Figure 2. The dependence of An content of plagioclase on temperature and water pressure, obtained by our experiments. The water pressure of the Us-b (rhyolitic: R) magma chamber is estimated to be about 200 MPa, consistent with the estimate of 250 MPa from the phase equilibria. After the 1663 eruption, a shallower (dacitic: D) magma chamber formed at about 100 MPa.

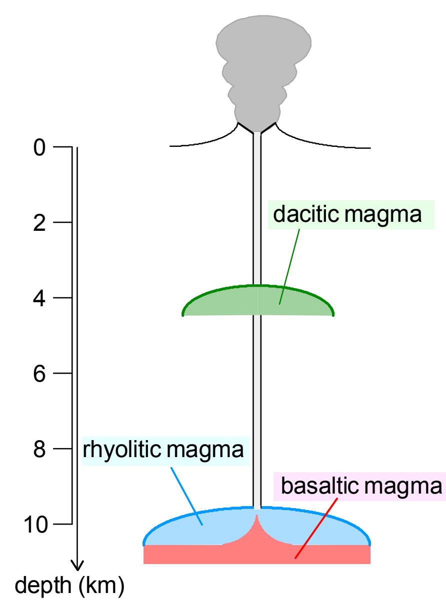

Figure 3. A schematic cross section of the Usu magma-feeding system. The main chamber is about 10 km deep and the later chamber at about 4 km deep. The latter formed after the 1663 eruption.