LETTER

Evolution of permeability and fluid pathways in the uppermost oceanic crust inferred from experimental measurements on basalt cores

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To examine the systematic decrease in permeability with crustal age in the uppermost layer of the oceanic crust, we investigated the effect of mechanical compaction on the permeability of basalt cores from Hole 765D and Hole 456 of the Ocean Drilling Program. The results indicate that permeability systematically decreases with increasing confining pressure. However, the pressure effects observed in the laboratory experiments are insufficient to fully explain the result of in-situ measurements of permeability through the oceanic crust. The permeability of core samples is closely related to porosity, both of which reduce with increasing crustal age. We infer that the evolution of permeability of the oceanic upper crust could be controlled by a reduction in porosity due to carbonate mineral precipitation.

Keywords: Oceanic crust, Hydrothermal system, Permeability, Porosity, Carbonate precipitation

INTRODUCTION

Fluid flow transports heat and solutes within the oceanic crust, thereby modifying its physical state; this has a significant influence on the processes of the oceanic lithosphere as well as on the seismicity at subduction zones (e.g., Fisher, 1998). The lateral fluid flow within the oceanic upper crust is essential to explain seafloor heat flow anomalies and the apparent homogenization of temperature near the boundary between the sediment and basement (e.g., Wolery and Sleep, 1976). In this respect, permeability is an important parameter to consider while studying the circulation of seawater and heat flow within the oceanic upper crust. Fluids tend to circulate in the uppermost basalt layer (2A) far from the ridge, because the permeability of this layer is much higher than that of the overlying pelagic sediments and the underlying massive basalt layer (Fisher, 1998).

In-situ measurements of the permeability of the uppermost basalt layer reveal a systematic decrease with increasing crustal age, where permeability reduces by four orders of magnitude with crustal age (10^{-10} m² for 0.9 Ma

and 10^{-14} m² for 7.4 Ma) (Fisher and Becker, 2000; Becker and Davis, 2003). Laboratory measurements of basalt core samples yield a wide range of permeabilities (Karato, 1983; Christensen and Ramanantoandro, 1988); however, the mechanism of the permeability decrease is not well understood. In this study, we measured the permeability of basalt core samples from Ocean Drilling Program sites in the northeastern Indian Ocean (139 Ma) and east of the Mariana Trough (<1.8 Ma). In the laboratory measurements of permeability, we focused on the effect of confining pressure to test whether mechanical compaction is able to explain the age evolution of permeability of the uppermost layer of the oceanic crust.

SAMPLE DESCRIPTIONS AND EXPERIMENTAL METHOD

All basalt core samples were selected from the oceanic drilling cores stored at the Kochi Core Research Institute, Japan. We performed laboratory measurements of permeability and porosity of basalt samples from Hole 765D located in the northeastern Indian Ocean, and from Hole 456 located in the east of the Mariana Trough. We chose core samples located within the uppermost 200 m of the basalt layer, with samples from Hole 456 being particu-

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Figure 1. Photomicrographs of studied basalt samples under opened polarizers. The matrix of these samples consists mainly of plagioclase. Basalt samples from Hole 765D (a) contain abundant calcite veins and those from Hole 456 (b) contain several vesicles.

Table 1. Sample description and experimental results

No. Hole	Locality (Age)	Rock type	Depth (mbsf)	Bulk density (g/cm ³)	Grain density (g/cm ³)	Porosity (%)	Permeability (m ²)
1. 765D	Indian ocean (139 Ma)	Massive aphyric basalt	950-960	2.80	2.88	2.9	1.3×10^{-21}
2. 765D		Massive basalt	980-990	2.82	2.92	3.1	1.3×10^{-20}
3. 765D		Massive sparsely phyrlic basalt	1160-1170	2.75	2.80	2.0	1.4×10^{-20}
4. 456	East of Mariana Trough (<1.8 Ma)	Pillow basalt	133-135	1.98	2.75	28.1	1.1×10^{-16}
5. 456		Pillow basalt	133-135	2.18	2.89	24.6	6.9×10^{-17}
6. 456		Aphyric pillow basalt	152-153	2.26	3.01	25.2	4.8×10^{-17}

larly close to the top of the basalt layer. The uppermost oceanic crust is characterized by a uniform structure (Jonathan and Robert, 1994), and therefore these samples likely exhibited a similar pore structure. Samples from Hole 765D are as old as 139 Ma, and samples from Hole 456 are young, with ages <1.8 Ma. The samples from Hole 765D contain calcite veins and have relatively high bulk densities, whereas those from Hole 456 include a small amount of calcite and are characterized by relatively low bulk densities due to the presence of vesicles (Fig. 1). The mean grain densities of Hole 765D and Hole 456 samples are 2.87 and 2.88 g/cm³, respectively (Table 1). Thus, although the bulk density of the core samples differs between the two holes, the grain density is similar.

Cylindrical samples approximately 20 mm in diameter and 10–20 mm in length were prepared for permeability measurements. The two end-faces of each sample were polished to ensure parallelism. To remove any absorbed water, each specimen was dried at 70 °C for several days prior to measurement. To measure permeability, we used an inter-vessel deformation and fluid flow apparatus at Hiroshima University, Japan (Katayama et al., 2012). This apparatus is able to measure a wide range of permeability from around 10^{-12} to 10^{-22} m². Samples were covered with a polyolefin sleeve to isolate the pore fluid from the confining pressure fluid. To ensure complete saturation, sample assemblies with the rock core in place were evacuated before pore fluid was injected. Gas permeabilities were measured using a fluid flow method

at room temperature under a confining pressure that ranged from 5 to 50 MPa.

We used nitrogen gas instead of water as the pore fluid as it has several advantages, such as: (1) experimental times are often shorter; (2) nitrogen gas is chemically inert, so no chemical effects occur; and (3) the viscosity of nitrogen gas is less sensitive to changes in temperature than that of water, and hence the uncertainties in permeability measurements are much smaller than those when water is used as a pore fluid (Tanikawa and Shimamoto, 2009). Gas permeability is up to one order of magnitude higher than water permeability because of the slip flow of gases at pore walls (Klinkenberg effect). Although the absolute value of permeability measured using the gas flow method can differ from that measured using water, we were able to investigate the effect of confining pressure, which is not affected by the type of pore fluid. In comparison to the previous study, we used intrinsic permeability, including the Klinkenberg correction. The porosity was measured with a gas pycnometer, in which the pore volume was calculated from the pressure difference after injecting gas into the sample according to Boyle's law.

EXPERIMENTAL RESULT

We measured change in permeability during compression under confining pressures (P_c) of up to $P_c = 30$ MPa in Hole 765D and up to $P_c = 50$ MPa in Hole 456 (Fig. 2). Each sample shows three patterns of gas permeability,

Figure 2. Results of permeability experiments. Each sample shows three patterns of gas permeability, which correspond to three different upstream pore pressures ($P_p = 0.5, 1.0,$ and 2.0 MPa). In these experiments, fluid flow was driven by the pore pressure gradient, where the downstream pore pressure was set at atmospheric conditions. The errors on the values were estimated from the measurement uncertainty of the fluid flow meter.

which correspond to three different pore pressures—(P_p) of $P_p = 0.5, 1.0,$ and 2.0 MPa. Owing to the Klinkenberg effect, gas permeability is slightly dependent on pore fluid pressure.

In all experiments, gas permeability decreases with increasing confining pressure (Fig. 2). Permeabilities of Hole 765D samples are about three orders of magnitude lower than those of Hole 456 samples. This systematic difference in permeability matches the direction of the porosity differential between the two groups of samples, whereby porosity values are around 3% for Hole 765D and around 25% for Hole 456 (Fig. 3). This suggests that the variation in permeability of the samples could be controlled by the variation in porosity and hence, the evolution of permeability of the oceanic crust could be controlled by changes in porosity. It should be noted that porosity measured using the above mentioned method may include isolated pores, whereas only connected pores are responsible for fluid flow. The values of permeability in Figure 3 have been corrected to values of intrinsic permeability with the measured gas permeability at $P_c = 10$ MPa using the Klinkenberg correction. These data were compared with the experimental data for Hole 504B (Karato, 1983), in which permeabilities were measured at $P_c = 15$ MPa and upstream pore pressure = 7 MPa. Therefore, the effective mean pressure (~ 11 MPa) was similar to our result. In this experiment, porosity were calculated from the wet-bulk density. The data for Hole 504B plot between those for Hole 456 and Hole 765D (Fig. 3). As the ages of Hole 504B samples are close to 6 Ma, and therefore lie between the ages of samples from Holes 456 and 765D, permeability and porosity are systematically correlated to the age of the oceanic crust.

Figure 3. Relationship between permeability and porosity for Holes 765D and 456 (from this study) compared with Hole 504B data (Karato, 1983). The permeability and porosity measurements from Holes 765D and 456 have been corrected to intrinsic permeability values with a measured gas permeability of $P_c = 10$ MPa using the Klinkenberg correction.

The effect of pressure on permeability differs between samples from Holes 456 and 765D. The results show a weak pressure dependence for Hole 456 and a strong pressure dependence for Hole 765D (Fig. 4). The weak pressure dependence for the young oceanic crust (Hole 456) is consistent with previous laboratory measurements for samples from Hole 504B (Karato, 1983).

DISCUSSIONS

The laboratory permeabilities measured in this study are considerably lower than the permeabilities of the uppermost few hundred meters of the crustal basement measured using borehole packer methods. This disparity could reflect the difference in dimensions of fluid pathways between the field and laboratory scales. In the uppermost

Figure 4. Normalized permeability data. Hole 765D and Hole 456 show data of three samples described in Figure 2. The permeability of old crust (Hole 765D) decreases significantly with increasing confining pressure. In contrast, the permeability of young crust (Holes 456 and 504B) (Karato, 1983) decreases more gradually by less than one order of magnitude.

oceanic crust, large fractures can develop within pillow basalt, including hyaloclastite; fluids tend to flow into these large fracture zones. In contrast, such fractures are not included in hand specimen samples and therefore laboratory measurements do not record their influence. As a consequence, the permeability of the oceanic crust measured from an in situ borehole test is several orders of magnitude higher than that of a laboratory test (Brace, 1984). The absolute values of permeability between experiments and in-situ measurements are systematically different; however, this difference is probably caused by the hydraulic radius of porous media and hence, pressure dependence can be applied to natural systems (e.g., Karato, 1983).

The permeability of the uppermost 200–300 m of the basalt layer is relatively high but shows a systematic decrease with the oceanic plate age (10^{-10} m^2 for 0.9 Ma and 10^{-14} m^2 for 7.4 Ma) (Becker and Davis, 2003). This decrease in permeability is initially caused by mechanical compaction as the confining pressure of the basalt layers increases due to both the increasing thickness of pelagic sediments and the increasing depth of the ocean. Thus, we first estimated the confining pressure of these basalt layers as a function of pelagic sediment thickness and seafloor depth. In this calculation, we incorporated an accumulation rate of pelagic sediment of 5.0 m/Myr (Johnson and Pruis, 2003), and used a value for the increase in seafloor depth approximately equal to the square root of crustal age (Parsons and Sclater, 1977); a density of 2.6 g/cm^3 was used for pelagic sediment and 2.7 g/cm^3 for basalt. The results indicated that the confining pressure of the uppermost 300 m of the basalt layer ranges from 36.0 MPa for a crustal age of 1.0 Ma to 42.4 MPa for 7.0 Ma. Such a small change in confining pressure is insufficient to reduce permeability by four orders of magnitude (Fig. 4), indicating that mechanical compaction alone cannot explain the observed evolution of (i.e., the change in per-

meability with crustal age) in the uppermost oceanic crust. According to Figure 4, Hole 765D shows a strong pressure dependence, but Hole 456 shows a weak pressure dependence. This difference in permeability change likely reflects the pore structure. Only connected pores play an important role in permeability (Martin and Michael, 1999). Therefore, the geometry of a pore in each sample could be different.

Based on these experiments, factors other than mechanical compaction are required to explain the decrease in permeability with age in the young oceanic crust within the uppermost 300 m of the basalt layer. Carbonate veins are ubiquitous in the oceanic crust accreted in ophiolite sections and are precipitated within a few million years of the formation of the crust (Hart and Staudigel, 1978; Richardson et al., 1980; Staudigel et al., 1981). Calcite veins are more abundant in samples from Hole 765D (30 vol%) than in young samples from Hole 456 (5 vol%) (Fig. 1). This carbonate precipitation likely results in the reduction of porosity (Fig. 3) and consequently, the permeability also decreases with crustal age. Precipitation might be completed in the first 10 Myr of crustal evolution, since porosities of basalt core samples for Holes 504B and 765D are similar (Fig. 3), suggesting that fluid flow activity might be limited to the relatively young oceanic basements. We suggest that such CaCO_3 precipitation might be associated with biotic activity, because Staudigel and Furnes (2004) proposed that more than 50% of the alteration occurring in the uppermost 300 m of oceanic basalt is caused by biospheric processes.

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