Petrophysical properties of the root zone of sheeted dikes in the ocean crust: A case study from Hole ODP/IODP 1256D, Eastern Equatorial Pacific

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ABSTRACT

ODP (Ocean Drilling Program)/IODP (Integrated Ocean Drilling Program) Site 1256 is located on the Cocos Plate in the Eastern Equatorial Pacific Ocean, in a 15 Ma old oceanic lithosphere formed at the EPR during a period of superfast spreading (~200 mm/yr). ODP/IODP Hole 1256D reached for the first time the contact between sheeted dikes and underlying gabbros. It consequently offers a unique opportunity to study in situ, in present-day oceanic crust, the root zone of the sheeted dike complex. This root zone is a thin, 100 m thick boundary layer between the magmatic system (i.e., the axial melt lens, ~1100 °C), and the overlying high temperature hydrothermal system (~450 °C). The understanding of interactions within this boundary layer is critical to that of crustal processes along mid-ocean ridges. This work focuses on the petrophysical characterization of the root zone of the sheeted dike complex in order to further constrain the hydrothermal circulation system in the vicinity of the axial melt lens, as recorded in non-granoblastic dikes, granoblastic dikes, and varitextured gabbros. The petrophysical properties were determined from sample measurements in the laboratory and were compared to in situ downhole geophysical probing. The porosity structure is bipolar, depending on lithology, resulting in a layered system. Non-granoblastic dikes are generally altered in the greenschist facies (~250 °C) with relatively high and interconnected (cementation index m~1.7, electrical tortuosity τ~28.3) porosity (1.5%). In contrast, gabbros are retrogressively metamorphosed in the amphibolite (~450 °C) and greenschist facies, with lower porosity (1.3%) that involves numerous fissures and cracks, resulting in a more connected medium (m~1.58, τ~11.8) than non-granoblastic dikes. These cracks are more abundant but also tend to close with increasing depth as indicated in downhole geophysical data. Porosity and alteration, as viewed from surface electrical conductivity, appear to be directly correlated.

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1. Introduction

The mid-ocean ridge system, the largest single volcanic feature on Earth, is where new oceanic crust is produced, hence the locus of significant interactions between tectonic activity, volcanism, and seawater circulation. Hydrothermal circulation at mid-ocean ridges allows efficient and rapid cooling of the newly formed crust (e.g., Johnson et al., 1993; Phipps-Morgan and Chen, 1993; Stein and Stein, 1994). Seawater penetrates into the upper crust, and is heated as it moves down toward the base of the sheeted dike complex (e.g., Teagle et al., 1998). The discharge of hydrothermal fluids occurs along the ridge axis at temperatures up to ~400 °C (e.g., Koschinsky et al., 2008). The root zone of the sheeted dike complex, which represents a thin zone (~100 m thick) of extreme thermal gradient (up to 7 °C/m), is regarded as a thermal boundary layer between the convective magma chamber system below (~1100 °C) and the main convective hydrothermal system (~450 °C) above (e.g., Gillis, 2002; Honnorez, 2003; Gillis, 2008; Koepke et al., 2008; France et al., 2009).

Our current understanding of the sheeted dikes/gabbros transition was facilitated by i) geophysical observation at the East Pacific Rise (EPR) (e.g., Singh et al., 1999; Tolstoy et al., 2008), ii) structural/petrological studies of ophiolites (e.g., Nicolas and Boudier, 1991; Gillis, 2002, 2008; Nicolas et al., 2008; France et al., 2009), and iii) numerical modeling of hydrothermal circulation at the ridge axis (e.g., Cherkouki et al., 2003; Fontaine et al., 2008).

Regarding the magmatic system, geophysical imaging has revealed the occurrence, at the ridge axis, of a thin, narrow and nearly continuous melt lens that overlies the magma chamber at depth (e.g., Detrick et al., 1987,1993; Kent et al., 1993), and which feeds the upper crustal basaltic lid that is formed by the sheeted dike complex and the upper volcanic sequence (e.g., Phipps-Morgan and Chen, 1993; Singh et al., 1998; Macleod and Yaouancq, 2000). This upper melt lens is a key component of oceanic crust formation at fast spreading ridges. Interaction between the magmatic and the hydrothermal system can modify the composition of the melt lens and therefore influence the ocean crust composition (e.g., Coogan et al., 2003; France et al., 2009).
Several studies have focused on the dike/gabbro transition in ophiolites, and have lead to apparently contrasting models. Some authors propose that the melt lens is a steady state horizon (e.g., Nicolas et al., 2008) when others propose that it behaves as a dynamic horizon with upward and downward migrations of the top of the melt lens (e.g. Gillis, 2002, 2008, France et al., 2009).

Regarding the hydrothermal system, many studies show that faults and fractures play a major role in the localization and evolution of hydrothermal system at various scales on the upper crust (e.g. Norton and Knapp, 1977; Lowell et al., 1995; Singh et al., 1999; Tolstoy et al., 2008). However, geophysical studies also show that the deep hydrothermal circulation is located in a layer just above the magmatic lens where no seismic reflector is observed (Singh et al., 1999; Tolstoy et al., 2008). Numerical models are helpful to better understand the complexity of the hydrothermal circulation in the vicinity of the magma chamber. Models assuming hydrothermal flow in a porous medium with quasi isotropic permeability distribution allow estimate of heat and mass fluxes at the segment scale (e.g. Rosenberg and Spera, 1990; Travis et al., 1991; Wilcock, 1998, Cherkaoui, et al. 2003, Fontaine et al., 2008). The principal parameters that are tested are the permeability of the crust (linked to alteration, porosity; e.g., Fontaine et al., 2001), the basal temperature of the hydrothermal system, and the thermal perturbation related to a diking event (e.g., Cherkaoui and Wilcock, 1997).

For the first time in scientific ocean drilling, the 1507.1 m deep ODP/IODP Hole 1256D (Fig. 1; Wilson et al., 2003, 2006; Teagle et al., 2006; Alt et al., 2007) drilled a complete section of upper oceanic crust, from extrusive lava through dikes and into a few tens of meters of varitextured gabbros (Fig. 2). It is located in the Guatemala basin, and penetrates into 15 Ma crust that was formed at the EPR during superfast spreading period (>200 mm/yr) (Fig. 1; Wilson, 1996; Wilson et al., 2003). Hole 1256D consequently offers a unique opportunity to study the root zone of the sheeted dike complex in present-day in situ oceanic crust.

Permeability is the primary physical property controlling the localization, intensity and duration of the hydrothermal circulation in the upper crust. Despite its importance, we presently have a limited understanding of how permeability is distributed within oceanic basement. This understanding is largely based on a few direct measurements in boreholes (e.g. Becker, 1986) and a collection of indirect estimates (e.g.,Johnson, 1980; Hamano, 1980; Karato, 1983; Christensen and Ramananantsoandro, 1988; Fisher, 1998). At pore scale, permeability is a function of the connected porosity. In this paper, we present new data on the topology of the porous network of the root zone of the sheeted dike complex in Hole 1256D, from 1250 mbsf (Meters Below Sea Floor) to the bottom of the hole at 1507 mbsf (Fig. 2). It is based on the integrated analysis of downhole geophysical data, petrophysical measurements on core, and petrological observations (Fig. 2). The porosity is detailed for the three main lithologies of the root zone of the sheeted dikes (non-granoblastic dikes, granoblastic dikes, and varitextured gabbros) at microscopic (mini-core measurement) and mesoscopic scales (downhole measurements) from two independent methods (acoustical and electrical measurements), in order to i) to identify possible hydrothermal circulation channels, ii) to determine the porosity profile of the oceanic crust at the transition between dikes and gabbros, and iii) to discuss the possible constraints brought by this new data set on ocean crust hydrothermal circulation processes.

2. Lithostratigraphy of the root zone in Hole 1256D

The uppermost basement at Site 1256 consists of thin, sheet lava flows alternating with massive lava flows and pillow lava flows (from ~250 to 1004 mbsf). The lithological transition zone (~1004 to 1060 mbsf) that separate these flow units from the underlying sheeted dike complex is marked by breccias and highly fractured basalts. Dike chilled margins become common in recovered cores below 1060 mbsf. Alteration temperature increases downward, from lava into dike, with low temperature phases (~150 °C) giving way in partially altered dikes to chlorite, and hornblende below 1350 mbsf (400 °C) (Teagle et al., 2006). In the lower 60 m of the sheeted dikes (1348–1407 mbsf), doleritic basalts are partially recrystallized to distinctive granoblastic textures, which are interpreted as resulting from metamorphic overprint due to the vicinity of either a gabbroic sill intrusion (Wilson et al., 2006) or the axial melt lens itself (Koepke et al., 2008; France et al., 2009). The first gabbros recovered below the sheeted dikes were cored at 1407 mbsf. The lowermost rock recovered in the hole is a basaltic dike lacking granoblastic textures (Teagle et al., 2006; Fig. 2).

A series of 21 samples (mini-cores) were selected to represent the geological variability near the base of Hole 1256 D (Fig. 2; Table 1). They were generally taken away from alteration features such as veins and halos.

Primary and secondary mineralogical assemblages (Fig. 2), as well as textures (Fig. 3), were characterized on thin sections. These observations complement shipboard descriptions (Teagle et al., 2006), which are summarized herein. The root zone can be subdivided into three sections: non-granoblastic dikes, granoblastic dikes (i.e., dolerite dikes in which granoblastic textures were observed), and gabbros. The core recovery was 36% in dikes, 7% in granoblastic dikes, and 35% in gabbros, with yields an average recovery for the root zone of 18.8% (Fig. 2; Teagle et al., 2006).

Non-granoblastic dikes, which are away from the zone of interactions between melt and hydrothermal fluids, represent the petrological reference in this study. The primary assemblage of dike is: plagioclase, clinopyroxene, Fe–Ti oxides, and minor olivine (Figs. 2 and 3). Significant but local textural heterogeneities (cryptocrystalline dike margins, medium-grained with few large (1 mm) phenocrysts in dike cores) (Fig. 3) are observed. Dikes are altered in the greenstinet facies (~250 °C). Eight non-granoblastic dike samples (Teagle et al., 2006) from the 1250–1350 mbsf interval were studied (Fig. 2).

- Samples 173R-1-58 (core #—section #—distance to top of section in cm) and 173R-2-3 are fine grain basalts. Primary minerals are plagioclase, clinopyroxene, and titanomagnétite. Plagioclase occurs as long (0.5–1 mm) laths. Clinopyroxene is partly replaced by chlorite, and occasionally by actinolite. Disseminated
Fig. 2. Principal downhole lithological, petrophysical and petrological characteristics of Hole 1256D between 1250 and 1507 mbsf. a) Lithostratigraphic column of the root zone of the sheeted dike complex at ODP/IODP Site 1256 showing recovery, lithology, mini-core locations, resistivity (LLd), and downhole seismic measurements. b) Downhole distribution of secondary minerals observed during IODP Expedition 312 (Teagle et al., 2006). From actinolite to magnetite: hydrothermal alteration under greenschist facies conditions (\(-250^\circ C\)). From dusty clinopyroxene to hornblende: hydrothermal alteration under amphibolite facies conditions (\(-400^\circ C\)). From clinopyroxene to granoblastic texture: high temperature metamorphism (\(>850^\circ C\)).
In the lower part of the sheeted dike complex, from 1348.3 to 1406.6 mbsf, the dikes are strongly to completely altered, and locally recrystallized to granoblastic textures. Granoblastic dikes have the same primary parageneses as non-granoblastic ones, but also display irregularly distributed granoblastic patches, within which primary assemblages are recrystallized to secondary plagioclase and equant secondary clinopyroxene, magnetite, ilmenite, and rare orthopyroxene (Fig. 2; Teagle et al., 2006; Koepke et al., 2008). The mineralogy (Opx and Cpx) and textures indicate that recrystallization occurred at very high temperatures in granulite facies (>850 °C) (Teagle et al., 2006; Wilson et al., 2006; Koepke et al., 2008; France et al., 2009).

Because of the poor core recovery at the base of the sheeted dike complex (~7%), only two granoblastic samples could be studied. The first was sampled in the lower section of the sheeted dikes and the second in the granoblastic interval located between the lower and upper gabbro intervals.

- **Samples 202R-1-8.** A small proportion of this sample is recrystallized to microcrystalline (10 μm) aggregates of clinopyroxene, orthopyroxene and actinolitic hornblende (Fig. 3d). In non-recrystallized domains, hydrothermal alteration is poorly developed, and plagioclase is only slightly recrystallized. Hydrothermal veins of quartz surrounded by brown amphibole cross-cut recrystallized patches, which demonstrates that significant hydrothermal alteration postdates recrystallization and the development of the granoblastic texture.

- **Sample 227R-2-4.** The mineral assemblage is similar to the one in the previous sample, but recrystallized patches are larger, with coarser orthopyroxene (~100 μm) and actinolitic hornblende (Fig. 3d).

The plutonic section extends from 1406.6 mbsf to the bottom of the hole, at 1507.1 mbsf. This section is subdivided into the upper gabbro unit (1406.6–1458.9 mbsf) and the lower gabbro unit (1483.1–1507.1 mbsf) (Teagle et al., 2006). Gabbroic rocks are heterogeneous in grain size, mostly medium-grained (~500 μm) and textures are dominantly ophitic. Clinopyroxene is partially or completely altered by actinolitic hornblende rims. Along cracks, clinopyroxene is intensely altered to amphibole. Plagioclase is generally less altered than clinopyroxene, and partially replaced to titanomagnetite.
Fig. 3. Photomicrographs of sample thin sections (a-e and g-h: plane-polarized light; f: cross-polarized light). a) 173R-1, 58: non-granoblastic dikes with intergranular doleritic texture; b) 174R-1, 136: dolerite, with star plagioclase texture c) 176R-2, 58: non-granoblastic dike with plasmose plagioclase texture; d) 202R-1, 8: Dolerite partly recrystallized into a granoblastic assemblage of clinopyroxene (10 μm), orthopyroxene and actinolitic hornblende; e) 227R-2, 4: Dolerite partly recrystallized into a granoblastic assemblage of clinopyroxene (100 μm), orthopyroxene and actinolitic hornblende; f) 231R-2, 30: medium-grained gabbrro (500 μm) with ophitic texture partly recrystallized in amphibolites facies; g) 231R-2, 30: gabbrro with ophitic texture; h) 220R-1, 18: gabbrro with ophitic texture and heterogeneous grain size.
secondary plagioclase (albite), actinolitic hornblende, chlorite, prehnite, laumontite, and epidote (Teagle et al., 2006). A total of 6 samples from the upper gabbro unit and 5 samples from the lower gabbro unit were studied. The recovered gabbroic rocks display very heterogeneous textures, showing variable mineralogy and degree of alteration (Teagle et al., 2006); this heterogeneity is also encountered in our limited sampling (Fig. 3g–h).

- Samples 214R-2-62 and 216R-1-13 are olivine gabbros. Primary minerals are plagioclase, clinopyroxene, Fe–Ti oxides, and olivine. Homogeneous, ophitic texture is observed. Olivine is recrystallized to talc and chlorite. Plagioclase is largely recrystallized to albite and chlorite. Prehnite and epidote are present.

- Sample 215R-2-77 is an oxide-bearing gabbro. Primary minerals are plagioclase, clinopyroxene, and Fe–Ti oxides. The texture is homogeneously ophitic. Clinopyroxene and olivine are completely replaced by talc and amphibole. Disseminated titanomagnetite is common.

- Sample 220R-1-18 is olivine gabbro. Primary minerals are plagioclase, clinopyroxene, Fe–Ti oxides, and olivine. Ophitic texture with different grain size areas is present. Plagioclase is largely recrystallized to albite and prehnite. This assemblage indicates hydrothermal alteration at moderate temperatures.

- Sample 222R-1-125 is an oxide-bearing olivine gabbro. Primary minerals are plagioclase, clinopyroxene, Fe–Ti oxides, and olivine. Homogeneous, ophitic texture is observed. Olivine is present, with or without alteration halos of talc and amphibole. Plagioclase is only slightly recrystallized. Alteration is less abundant in this sample than in other gabbro samples studied herein.

- Sample 223R-2-36 and 230R-2-71 are oxide gabbros. Primary minerals are plagioclase, clinopyroxene, and Fe–Ti oxides. Ophitic texture with different grain size areas is present. The mineral assemblage is similar to the one in the previous sample but brown and green hornblende-rich amphibole becomes more abundant with depth, in association with secondary magnetite.

- Samples 231R-1-6, 231R-2-30 and 231R-2-126 are gabbronorites. Primary minerals are plagioclase, clinopyroxene, orthopyroxene, and Fe–Ti oxides. Ophitic texture with different grain size areas is present. Olivine is recrystallized to talc, amphibole and magnetite. Plagioclase is moderately to highly altered, appears dusty and corroded, is affected by albitionation, and contains specks of secondary magnetite.

- Sample 231R-4-51 is a gabbronorite. Primary minerals are plagioclase, clinopyroxene, orthopyroxene, and Fe–Ti oxides. The mineral assemblage is similar to the one in the previous sample but the proportion of olivine is higher.

3. Petrophysical methods

Petrophysical properties were determined from downhole geophysical measurements, and from sample measurements performed in the laboratory at atmospheric pressure and room temperature on a set of cm-scale cylindrical mini-cores (Figs. 4 and 5).

3.1. Density and porosity

Porosity plays a key role in fluid transport processes in the upper oceanic crust. It cannot be measured directly in situ from neutron activation in crystalline rock (Pezard et al., 1999) and can be dramatically overestimated when hydrous minerals or highly neutrophile trace elements (such as Gd, Eu, and Sm) are present (Harvey et al., 1996; Pezard et al., 1999, 2000; Helin-Clark, et al., 2004; Harvey and Brewer, 2005), yielding errors as an order of magnitude in porosity. As a consequence, whether from core or in situ, porosity has to be determined indirectly from petrophysical properties such as density, electrical resistivity or acoustic velocities. This way, in situ porosity profiles may be computed from density, velocity logs. (Fig. 4 column 2). The bulk density was measured in situ with the HLDS tool (Hostile Litho-Density tool of Schlumberger), which consists in a chemical radioactive Cs137 gamma ray chemical source associated to near and far gamma ray detectors (Ellis, 1987).

Symbol and index used in the equations listed below are summarized in Table 2.

In the laboratory, the porosity and grain density were computed using the triple weighing method where:

\[ \phi = \frac{M_{\text{sat}} - M_{\text{dry}}}{M_{\text{sat}} - M_{\text{im}}} \times 100 \]

and

\[ \rho_{\text{ma}} = \frac{M_{\text{dry}}}{M_{\text{dry}} - M_{\text{im}}} \rho_{\text{fluid}} \]

with \( \rho_{\text{fluid}} = 1.02 \text{g/cm}^3 \) for a 30 g/L salinity at 0.1 MPa and 20°C–25°C.

An alternative method (Melnyk and Skeet, 1986), consisting in the determination of the surface water weight of the mini-cores from drying curves, was also used for a more accurate estimate of the saturated sample weight, hence a better estimate of porosity and density (Fig. 6a).

In situ porosity profiles were recomputed by applying to downhole measurements (density and P-waves velocity) the relationship between porosity and P-wave velocities and between density and porosity measured in the laboratory, respectively. These computations take into account the influence of temperature and pressure on 1) the P-wave velocities in seawater 2) sea water density, for an average salinity of 35 g/L (Fig. 4 column 2). P-wave velocity in sea water and sea water density are computed from sea water equation of state of (Batzle and Wang, 1992) using the borehole fluid pressure and temperature measured by the TAP (Temperature Acceleration and Pressure) downhole tool (Teagle et al., 2006).

Porosity computed from downhole log density is given by Ellis (1987):

\[ \phi_{\text{insitu}} = \frac{a \rho_{\text{bulk}} + b}{a \rho_{\text{bulk}} + b} - \rho_{\text{fluid}} \]

(3)

with: \( a = 1/(1 - \phi_{\text{inc}}) \), \( b = (\phi_{\text{inc}} \cdot \rho_{\text{fluid}})/(1 - \phi_{\text{inc}}) \). \( a \) and \( b \) are derived empirically from mini-core analyses of density and porosity.

Porosity computed from the acoustic slowness “L” is given by Ellis (1987):

\[ \phi_{\text{insitu}} = \frac{(a' \cdot L_{\text{bulk}} + b')}{(a' \cdot L_{\text{bulk}} + b')} - L_{\text{fluid}} \]

(4)

\( a' \) and \( b' \) are also derived empirically from mini-core analyses, and refer to slowness (i.e., inverse of velocity).

3.2. Electrical properties

In a porous medium, comprising a matrix considered as electrically insulating, and a porous interconnected network saturated by a conductive electrolyte, two main types of electrical conduction can be distinguished: electrolytic conduction in the inner pore fluid present in the pore space, and surface conduction at the interface between minerals and the electrolyte (e.g., Waxman and Smits, 1968; Pezard, 1990; Revil and Glover, 1998). When the surface conduction component is negligible compared to the electrolytic component, the total conductivity of the porous media \( C_{\text{bulk}} \) can be regarded as directly proportional to that of the saturating fluid \( C_{\text{fluid}} \). This proportionality is the basis to define the electrical formation factor \( F \).
The occurrence of alteration mineral phases in an igneous rock can modify the expression of the total electrical conductivity. Waxman and Smits (1968) proposed for sand-clay mixtures a first order model that takes into account the excess conductivity \( C_s \) due to surface conduction processes:

\[
C_{\text{bulk}} = \frac{C_{\text{fluid}}}{F} + C_s
\]  

(5)

\( F \) is an intrinsic quantity that characterizes the 3D topology of the pore space, and describes the contribution of the pore space topology to the overall electrical resistivity of the fluid saturated media (Pape et al., 1985). The surface conduction \( C_s \) is mainly related to, and becomes significant in the presence of altered phases (Waxman and Smits, 1968; Pezard, 1990; Ildefonse and Pezard, 2001; Einaudi et al., 2005). Electrolytic conduction related to the formation factor \( F \) is mainly related to pore volume electrical transmissivity and, in the case of oceanic crystalline rocks to the presence of grain boundaries and micro-fractures. Electrical formation factor is related to sample porosity with two different, though similar approaches. The first one is based on Archie (1942) empirical equation:

\[
F = \frac{\tau}{\phi}
\]  

(7)

Where \( \tau \) relates to the geometrical complexity of the path followed by the electrical current in the fractured pore space or, in a more general sense, to the efficiency of electrical flow processes (e.g., Guéguen and Palciauskas, 1992; Clennell, 1997).

In order to evaluate the electrical properties and the geometry of the pore space for each of the samples, a series of electrical conductivity measurements were carried out after an initial phase of drying (vacuum setting during 48 ), then saturation with variable salinity NaCl solutions. Measurements were made at room pressure and temperature, with 7 different saturating fluid salinities (ranging from 0.008 S/m to 8 S/m; Fig. 6b). For reference, that of sea water at 20 °C is 5.0 S/m. While the Waxman and Smits (1968) model works well at high salinity, it tends to overestimate \( C_s \) at low salinity. In this domain, the non-empirical, statistical approach proposed by Revil and Glover (1998) is preferred, although initially derived for granular
sedimentary rocks Fig. 6b. The Revil and Glover (1998) model is based on the microgeometry of the porous space. The conductivity of the sample is then given by a complex model with two simpler forms at high and low salinities (Revil and Glover, 1998; Ildefonse and Pezard, 2001; Fig. 6b). The main electrical parameters were consequently deduced from the measurements processed using this more appropriate model.

In situ resistivity was measured in Hole 1256D at two different radial depths of investigation into the rock (Lld and LLs) with the Schlumberger Dual Laterolog (Ellis, 1987; Expedition 309/312 Scientists, 2006). The quality of the data is generally poor in the most resistive formations below 1373 mbsf (i.e., granoblastic dikes), with the highest resistivity values clipped for the Lld at 40 000 Ω.m, i.e., the maximum value that is allowed to be read by the downhole tool in order to prevent circuit damaging from excessive electrical tensions. Resistivity logging data have been corrected for borehole environmental effects such as borehole fluid resistivity and hole diameter from the Schlumberger charts (1988).

3.3. Acoustic velocities

In situ acoustic compressional and shear wave velocities were measured at 12 kHz with the Schlumberger dipole sonic tool (Expedition 309/312 Scientists, 2006; Guerin et al., 2008; Swift et al., 2008). In order to relate the 21 new mini-core to shipboard samples and downhole measurements, additional laboratory measurements of P and S-wave velocities were made, and used to evaluate elastic properties, porosity, alteration, and micro-cracking into the rock (e.g., Bourbié et al., 1986; Guéguen and Palciauskas, 1992; Mavko et al., 1998). In the laboratory, acoustic velocity measurements (Vp and Vs) were obtained at room pressure and temperature at 1 MHz and 500 KHz respectively, for both dry and saturated samples (Fig. 4, columns 4 and 5). A digital oscilloscope, a pulse generator and coupled piezoelectric transducers were used to measure Vp and Vs. In comparison, compressional velocities of shipboard samples were measured along three orthogonal directions from sea water-saturated mini-cubes at room pressure and temperature (Guerin et al., 2008; Swift et al., 2008).

3.4. Natural gamma radiation

Downhole natural gamma ray emission was measured in situ using the HNGS (Hostile Natural Gamma Spectroscopy) tool of Schlumberger. The spectral analysis of the gamma radiation received by the tool in five appropriate energy windows yields a total natural gamma count (in gAPI) plus that originating from

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<tr>
<td>M</td>
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potassium (in %) (Fig. 4 column 6), thorium, and uranium. A five-
window spectroscopy arrangement is used to determine the concentrac-
tions of radioactive potassium (in %), thorium (in ppm),
uranium (in ppm), and the total gamma ray emission (gAPI) (Ellis,
1987; Expedition 309/312 Scientists, 2006). Potassium, thorium,
and uranium contents in igneous rocks are generally related
directly to the presence of secondary minerals.

4. Results

Porosity is directly related to most petrophysical properties
including density, electrical resistivity, acoustic velocities, and
permeability (hence fluid–rock interactions). These physical proper-
ties tend to vary a lot in the root zone of the sheeted dikes. In this
context, porosity and permeability (largely controlled by fracturing
from thermal origin at the ridge axis) are key parameters. Porosity and
permeability are closely but not directly related, depending possibly
more on the shape and connectivity of the pore, than on the total
amount. As an illustration, pumice may have porosity values larger
than 0.5, but a very low permeability with micro-cracks only
connecting spherical pore between them. As a consequence,
permeability might be derived from porosity in very seldom cases.
In crystalline rocks from the root zone, the porosity signal is
dominated by micro-cracks from hydrothermal origin. In this case,
permeability appears to be controlled by a fraction of a percent of
the total volume of the rock. As a consequence, the microscopic shape
of the pore space deduced from laboratory measurements is discussed
here in great details from the different but complementary electrical
and acoustic approaches.

4.1. Porosity and density

Grain densities measured in the laboratory are not homogeneous
in the root zone, with an average of 2.94 g/cm³ ±0.03 g/cm³, a
maximum reached for oxide-rich samples (2.97 g/cm³ on average),
and a minimum related to altered non-granoblastic samples (2.92 g/
cm³ on average; Fig. 4; Table 1).

Porosity is, as expected for moderately altered igneous rocks, low
in most of the samples from the root zone (with an average of 1.33 ±
0.65% in non-granoblastic dikes, 0.92 ± 0.30% in granoblastic dikes,
and 1.43 ± 0.36% in gabbros). Highest porosities values (2.33% on
average) correspond to the most strongly altered (~80–90% replacement of primary phase) non-granoblastic samples (Samples
187R-1-1-114 and 187R-2-1) within a large-scale fault zone located
between 1325 and 1340 mbsf. This higher porosity values are
correlated over the same interval with high natural gamma
radioactivity values (up to 6 gAPI) in such rocks, and with low
electrical resistivity (~50 Ω m) (Teagle et al., 2006; Fig. 4).
The lowest porosity (0.70%) in dikes appears to correspond to the
recrystallized sample (sample 202R-1-8) with granoblastic textures.
Porosity at the top of the upper gabbro interval (1407–1425 mbsf) is
consistently higher (~1.8% ± 0.16%) than in the overlying granoblast-
ic dikes. In the gabbros, bulk density increases in relation with a
decrease in porosity, from a minimum of 2.91 g/cm³ at 1412 mbsf, to
a maximum of 2.99 g/cm³ at 1490 mbsf, within the second gabbro unit
(Fig. 4, Table 1). Porosity estimates derived from velocity and
density downhole geophysical measurements are higher than
porosities measured on discrete samples, particularly in the upper
50 m of the root zone, from 1250 to 1300 m, because samples were
generally taken away from veins and faults. The porosity maxima in
the downhole geophysical measurements are likely related to more
faulted and/or veined intervals.

4.2. Acoustic velocities

In general, the shipboard and laboratory post-cruise measure-
ments are not found to be in good agreement. For two samples
collected in the same core section, at nearby depth, compressional
velocities measured on post-cruise saturated sample are 300 to
500 m/s higher than those measured on shipboard saturated samples.
Compressional velocities of saturated samples measured onboard the
D/V JOIDES Resolution correspond to those obtained post-cruise for
dry samples (Fig. 4). We interpret this discrepancy as resulting from an
incomplete saturation of shipboard samples, likely due to the lack of
time often inherent to shipboard measurement procedures, which is
consistent with lower measured velocities.

In contrast, post-cruise laboratory measurements of saturated
samples are in good agreement with downhole velocities (Fig. 4).
Post-cruise Vp values for samples from the root zone range from 5300
to 6600 m/s, while Vp ranges from 3000 to 3700 m/s (Fig. 4). The
somewhat noisy Vp sonic log appears to be only moderately affected
by the presence of high or low porosity sections identified from other
in situ geophysical profiles or borehole wall images (at 1270 and
1300 mbsf, for example). This is also seen for core sample measure-
ments in the sheeted dike complex. Unfortunately, geophysical logs are
not available for most of the gabbroic intervals below 1407 mbsf.
Both shipboard and our measurements show that acoustic velocities

![Fig. 6. Porosity topology derived from acoustic measurements. a) Example of drying curve used to estimate saturated sample weight in crystalline rocks. The extrapolation of the linear part of the curve at t = 0 is the weight of the saturated sample (173-2-3). b) Core electrical conductivity as a function of the saturating fluid salinities. Curve fit: non-linear inversion of the experimental data using the Revil and Glover model (1998). The intrinsic electrical formation factor F and the surface conductivity Cs are extracted from the high- and low salinity parts of the curves, respectively.](image-url)
To a very significant and unexpected (Fig. 5). A similar, but less significant difference between the wet and dry S-wave velocities is observed. The large difference between wet and dry compressional velocities measured on mini-cores is discussed in the following.

4.3. Electrical measurements

From laboratory measurements made over several months at varying fluid salinity, the Revil and Glover (1998) model was used to compute the electrical formation factor \( F \) and the surface conductivity \( C_s \) for each sample (Table 1, Figs. 4 and 5). From top to bottom of the recovered basement section, the electrical formation factor is generally poorly correlated to the porosity (Fig. 5a). “Archie’s law” relationship between \( F \) and porosity \( \phi \) found for different types of rocks (Archie, 1942) is also not verified here, as found in previous studies of oceanic basalts and gabbros (Pezard, 1990; Ildefonse and Pezard, 2001; Einaudi et al., 2005).

Non-granoblastic dikes display high and variable electrical formation factors (920–6087), with the highest value corresponding to the freshest sample. Gabbros tend to have lower electrical formation factors (532–1064).

While electrical surface conduction is found over the entire dataset to be a direct function of porosity \( C_s = (0.19) \times \phi^{1.84} \), following a pattern already described for granites and oceanic basalts (Belghoul, 2007), the contribution of \( C_s \) to in situ electrical conduction remains small (from 1 to 6% in the root zone). \( C_s \) ranges from 0.06 to 0.42 mS/m (0.17 mS/m on average) in non-granoblastic dikes, and from 0.10 to 0.39 (0.21 mS/m on average) in gabbro, while electrical conduction \( C_{fluid}/F \) ranges in this interval from 3.5 to 35 mS/m for gabbros and basalts. Consequently, \( C_s \) might be neglected here in further analyses of porosity from in situ electrical resistivity.

\( F \) and \( \phi \) are also used to compute the electrical tortuosity \( \tau \) and cementation factor \( m \) (see above). While \( m \) tends to describe the non-uniformity of channels between pores, \( \tau \) relates to the 3D geometrical complexity of the flow path through the sample (e.g., Guéguen and Palciauskas, 1992; Cennell, 1997; Ildefonse and Pezard, 2001). When presented as a function of \( \phi \), these two parameters display a similar distribution (Fig. 5c and d) with 2 distinct groups of samples being linearly correlated with porosity. Dike and gabbro samples cluster along distinct trends, indicating different pore space topologies for these two types of rocks. Distinct formation factors \( F \) are also obtained for these two lithologies.

For a given porosity value, \( m \) and \( \tau \) are lower in gabbros (1.58 ± 0.06 and 11.9 ± 2.1 on average, respectively), than in non-granoblastic dikes (1.72 ± 0.19 and 28.3 ± 14.6 on average, respectively), suggesting a more fissural, and simpler pore space in gabbros than in non-granoblastic dikes. These results suggest that the chemical exchange related to alteration and recrystallization occur over somewhat larger distances in gabbro than in dikes. As observed for granite (Pezard et al., 1999), chemical exchange in gabbros appear to operate here in a more open system than in dikes, with an associated porosity increase corresponding to mass decrease from fluid circulation associated to the process. Because of the very low core recovery in granoblastic dikes, only two samples were available for this study. The electrical properties (Fig. 5a, b, c, and d) show that sample 2004R-1-8 has a poor space structure similar to that of non-granoblastic dikes, whereas sample 227R-2-4 has a poor space similar to that of varitextured gabbro samples, which is consistent with the thin section observations (cf section 2008R-1-8, 227R-2-4).

In complement to the electrical resistivity analysis, we also used acoustic velocity data to assess the small-scale porosity structure of our samples (Fig. 5f). The expected inverse correlation between \( V_p \) and \( \phi \) is observed. For P-waves, this inverse correlation is verified for all saturated samples and for dry non-granoblastic dikes, but not for dry gabbros (Fig. 5e).

5. Discussion

5.1. Origin of the difference in dry and wet P-wave velocities in gabbros

One of the main results from the joint investigation of electrical and acoustical properties of the root zone of the sheeted dikes in ODP/IODP Hole 1256D is that the porosity structure is bipolar, depending on lithology (i.e., gabbro, non-granoblastic dikes, and granoblastic dikes; Figs. 4 and 5). Within the same range of porosity, non-granoblastic dikes tend to display slightly higher electrical formation factor, cementation index and electrical tortuosity values than gabbros. The non-granoblastic dike and gabbro samples with relatively low electrical formation factors have generally low porosities, and are highly altered. This bipolarity can also be seen in the mini-core acoustic velocity data. The onset of the observed decoupling between \( V_p \) in dry and saturated samples is found to correspond to the transition from sheeted dikes to gabbroic lithologies at ~1407 mbsf. This relationship between pore space topology and lithology can be derived from a difference in mechanical properties, hence in cracking due to thermal stresses on axis. The more fractured type of rock, more permeable on axis, will then react more with hydrothermal fluids, hence lead to the more altered phase and, possibly, the more complicated pore space structure off-axis.

The difference in dry and wet P-wave velocities in gabbros is related to a contrast in porosity topology (e.g., Gueguen and Palciauskas, 1992). The pore space geometry has a large impact on acoustic velocities at low porosity values, as measured in the root zone of the sheeted dikes (1.3%). While shear moduli are generally considered to remain constant with respect to pore space saturation, incompressibility moduli are higher in a fluid saturated medium than in a dry medium, and bulk density increases linearly with saturation. At low porosity, the impact of density changes on velocity is almost negligible relative to moduli variations (e.g., Gueguen and Palciauskas, 1992). S-wave velocities depend mostly on shear modulus changes, hence remain unchanged in relation to saturation. Therefore, the S-wave velocity difference between dry and saturated media is small. In contrast, P-wave velocities depend also from incompressibility modulus. As a consequence, the difference between dry and saturated media P-wave velocities can be large. So, small changes in topology with increasing depth could be sufficient to explain this decoupling between dry and wet P-waves (Gueguen and Palciauskas, 1992).

All acoustic velocities were measured on mini-cores drilled horizontally in the core reference frame. While P and S-waves have the same propagation direction, their particle movement planes are orthogonal. In gabbros, the very low porosity relates mainly to open grain boundaries and triple junctions between minerals, and to micro-cracks, which is the dominating component. The Hudson’s model (1981a, b) was used to further test the influence of small cracks on acoustic waves propagation in our samples. This model leads to the derivation of an average pore space aspect ratio and to the average spatial density of cracks for each of the core samples. The model is based on a scattering theory analysis of the mean wave field in an elastic solid, with thin, penny-shaped ellipsoidal cracks or embedded inclusion (Mavko et al., 1998). David et al. (1999) adapted this model assuming that inclusions are much smaller than the acoustic wavelength. Crack density is obtained from the acoustic velocities and density of the rock, plus the Lamé compressional (\( \lambda \)) and shear (\( \mu \)) elastic constants of the solid phase.

\[
\varepsilon = \frac{\lambda}{\mu} + 2\mu \nu V^2 / \Gamma
\]
with

$$\Gamma = \frac{4}{27} \frac{(3\lambda_s + 2\mu_s)^2(\lambda_s + 2\mu_s)}{(\lambda + \mu_s)\mu_s} + \frac{64}{135} \frac{\mu_s(\lambda_s + 2\mu_s)(9\lambda_s + 10\mu_s)}{(3\lambda_s + 4\mu_s)(\lambda_s + \mu_s)}$$

(9)

The expression of the crack density for oblate ellipsoids can be derived from David et al. (1999):

$$\varepsilon = \frac{3}{4n} \frac{\phi}{\alpha}$$

(10)

with $\alpha$ representing the ratio of the short axis to the long axis ($b/a$) in the pore cross-section modelled as flat ellipsoids. The crack aspect ratio $\alpha$ is somewhat found to decrease with depth, indicating that cracks are more elongated as one gets closer to the base of the boundary layer (Fig. 7). The crack density increases linearly with increasing depth, from $0.03 \pm 0.01$ on average in non-granoblastic dikes to $0.08 \pm 0.01$ on average in gabbro (Fig. 7). The derived crack densities and aspect ratio are consistent with a dominantly fissural porous space, with becomes more prominent in gabbros and granoblastic dikes than in non-granoblastic dikes. This result is consistent with the hypothesis that small cracks are responsible for the large differences in P-wave velocities measured in gabbros between dry and saturated samples.

The geometry of the pore space at the mini-core scale has been constrained from independent electrical and acoustic petrophysical measurements. Electrical measurements (primarily controlled by topology and chemical nature of the liquid phase) and acoustic measurements (largely controlled by the solid phase) are found to provide a coherent microscopic description of the pore space topology, with positively correlated electrical tortuosity $\tau$ and aspect ratio $\alpha$ (Fig. 8). The electrical tortuosity, related to micro-cracks, is naturally expected to increase while the aspect ratio of these cracks approaches 1.

Porosity characterization in IODP Hole 1256D is biased by coring and sampling processes; core recovery is very low, especially in the granoblastic dikes (~8%), allowing to sample only the less porous rocks, and mini-cores were preferentially sampled away from veins and halos.

5.2. Role of the micro-porosity in the deep hydrothermal circulation

Recent geochemical studies of ophiolites (e.g. Bosch et al., 2004), show that very high temperature fluids can reach the magma chamber margins at depth, in the gabbro section. Fractures in the upper crust (layer 2a and 2b) constitute the bulk of the porosity and therefore play a fundamental role in circulation of these fluids (Norton and Knapp, 1977; Lowell et al., 1995). Geophysical studies at the EPR suggest that the deep hydrothermal system is structurally different. Closure of the hydrothermal system is located in a layer just above the magmatic lens where no seismic reflector is observed (Singh et al., 1999; Tolstoy et al., 2008), i.e., no major fracture is imaged in this zone, and micro-porosity is expected to play a significant role in controlling hydrothermal circulation. At 9°50’N on the EPR, beneath a well-studied hydrothermal vent field, the hypocenters of micro-earthquakes (Tolstoy et al., 2008) cluster within a 500 m thick layer above the melt lens, and are interpreted as indicating thermal cracking where cold sea water extracts heat from hot crustal rocks. No seismic reflector is observed within this layer. This is consistent with the seismic reflection study at 14°10’ S (Singh et al., 1999), which shows that hydrothermal circulation at depth occurs in a 200 m thick low velocity layer, with no seismic reflector, located immediately below or within the sheeted dike sequence.

5.3. Origin of micro crack porosity

The origin of the porosity in the root zone of the sheeted dikes is unknown, and could be ascribed to natural processes (e.g., brittle deformation related to regional tectonics, thermal stress related to cooling and/or warming). Mechanical stresses related to the off-loading encountered as a result of the coring process might also contribute to porosity (e.g., Pezard, 2000). However, several observations suggest that porosity is at least partly related to the high temperature hydrothermal circulation at the ridge axis.
- A positive correlation is observed between surface conductivity (that is in principle related to alteration; Waxman and Smits, 1968), and porosity for the three lithologies (non-granoblastic dikes, granoblastic dikes, and gabbros), suggesting that at least part of the measured porosity in our samples is related to hydrothermal circulation (and associated alteration). However, no correlation is found between porosity and the amount of alteration as described in thin sections. This apparent contradiction results from the difference in the way alteration is characterized, electrically or visually. Surface conductivity is only sensitive to the interface between alteration minerals and the electrolyte (e.g., Waxman and Smits, 1968; Pezard, 1990; Revil and Glover, 1998), whereas thin visual description of alteration, using both thin sections and hand samples, includes the estimate of groundmass alteration. Here we consider the correlation between Cs and porosity meaningfully; we discuss the possible role of the connected porous network (i.e., the one seen by electrical surface conduction) in hydrothermal fluid circulation.

- In granoblastic dikes, hydrothermal high temperature veins cross-cut recrystallized aggregates. Hence, late high temperature hydrothermal alteration postdates recrystallization and the development of the granoblastic texture. We therefore assume that at least part of the measured porosity in our samples is related to high temperature hydrothermal circulation at or close to the ridge axis. These observations are consistent with the recent study of crack porosity in the upper oceanic crust, in ODP holes 504B and 1256D (Carlson, 2010), which documents (see also Alt et al., 1996) a strong correlation between crack density, aspect ratio and patterns of alteration, suggesting that porosity is related to high temperature processes.

- Fissural porosity in gabbros and its association with high temperature alteration (cracks filled by brown amphibole in clinopyroxene; Teagle et al., 2006), suggest that the measured porosity in our samples is at least partially related to high temperature hydrothermal circulation at or close to the ridge axis. At the ridge axis, micro-cracking associated to high temperature hydrothermal fluid circulation can occur in varietextured, isotropic gabbros that are left above the melt lens as the latter migrates downwards (e.g. Gillis, 2008; France et al., 2009). In addition to this on axis contribution, cracking can also occur off-axis; Nicolas et al. (2003) observed in the Oman ophiolite high temperature alteration in lower crustal gabbros associated with a regular network of very thin (<1 mm), parallel micro-cracks. These micro-cracks are interpreted as preferential downflow channels for high temperature hydrothermal circulation, which can bring hot fluids down to the base of the crust (Bosch et al., 2004).

5.4. Constraints on the cracking front in the upper crust

Lister (1982) developed a conceptual model for the penetration of water in to cooling crust near oceanic ridges, in which fluids migrate via permeability generated by a cracking front that propagates downward into fresh, uncracked rock. Fracturing occurs when tensile stress due to cooling and contraction exceeds the lithospheric stress. Lister (1982) envisioned that the cracking front could affect the entire crustal section, and recognized that it would penetrate the dike-granobros transition only after the lithospheric section had moved off-axis. Our observations of porosity of hole IODP 1256D support the idea that water can penetrates, through micro-cracking, down to the dikes/gabbros transition at the ridge axis. Micro-cracks associated to high temperature assemblages are present in both dikes and gabbros, and are more abundant with depth (Fig. 7).

6. Conclusion

Porosity is one of the most important variables controlling fluid flow in hydrothermal systems. New laboratory data obtained from mini-core samples are presented in this work, and variability in the pore space geometry is demonstrated from two complementary methods, based on electrical resistivity, and acoustic velocity measurements, respectively. In the 15 Ma old cores recovered in ODP/IODP Hole 1256D, the investigated porosity structure results from the cumulative effect of ridge axis processes, off-axis processes, and possibly the coring process. However, petrological and petrophysical arguments show that at least part of the measured porosity in our samples is related to high temperature hydrothermal circulation at or close to the ridge axis. The porosity structure is found to be bipolar, controlled by lithology (i.e., non–granoblastic dikes, and gabbros), indicating that small-scale porosity of the ocean crust at the base of the sheeted dike complex is somewhat layered. Fluid circulation and the associated fluid–rock interaction play a significant role in the evolution of porosity of the oceanic crust. Non-granoblastic dikes are generally altered under greenschist facies conditions (>250 °C), and are marked by somewhat higher porosity values (1.5%), mostly interconnected (m=1.7, τ=28.3). In contrast, gabbros are recrystallized, after cooling at the ridge axis, in the amphibolite facies (>400 °C), and marked by slightly lower porosity values (1.3%) and fissural porosity (m=1.58, τ=11.8, α=0.025). The porosity topology of Granoblastic dikes is in between that of non-granoblastic dikes and that of gabbros. Based on core measurements from IODP Hole 1256D, this work sheds light on the vertical heterogeneity of the micro-scale porosity structure of the sheeted dikes root zone, in the upper oceanic crust.

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