Physical properties of upper oceanic crust: Ocean Drilling Program Hole 801C and the waning of hydrothermal circulation

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[1] The hydrologic evolution of oceanic crust, from vigorous hydrothermal circulation in young, permeable volcanic crust to reduced circulation in old, cooler crust, causes a corresponding evolution of geophysical properties. Ocean Drilling Program (ODP) Hole 801C, which obtained the world’s oldest section of in situ, normal oceanic crust, provides the opportunity to examine relationships among hydrologic properties (porosity, permeability, fluid flow), crustal alteration, and geophysical properties, at both core plug and downhole log scales. Within these upper crustal basalts, fluid flux in zones with high porosity and associated high permeability fosters alteration, particularly hydration. Consequently, porosity is correlated with both permeability and a variety of hydration indicators. Porosity-dependent alteration is also seen at the log scale: potassium enrichment is strongly proportional to porosity. We extend the crustal alteration patterns observed at Hole 801C to a global examination of how physical properties of upper oceanic crust change as a function of age based on global data sets of Deep Sea Drilling Project and ODP core physical properties and downhole logs. Increasing crustal age entails macroporosity reduction and large-scale velocity increase, despite intergranular velocity decrease with little microporosity change. The changes in macroporosity and velocity are significant for pillows but minor for flows. Matrix densities provide the strongest demonstration of systematic age-dependent alteration. On the basis of observed decreases in matrix density that are proportional to the logarithm of age, approximately half of all intergranular-scale crustal alteration occurs after the first 10–15 Myr. Apparently, crustal alteration continues, at a decreasing rate, throughout the lifetime of oceanic crust. INDEX TERMS: 3015 Marine Geology and Geophysics: Heat flow (benthic) and hydrothermal processes; 7220 Seismology: Oceanic crust; 3035 Marine Geology and Geophysics: Midocean ridge processes; 5102 Physical Properties of Rocks: Acoustic properties; 8135 Tectonophysics: Evolution of the Earth: Hydrothermal systems (8424); KEYWORDS: Ocean Drilling Program, basalt, oceanic crust, logging


1. Introduction

[2] An elegant conceptual model describes the hydrologic and geophysical evolution of oceanic crust. Hydrothermal circulation is vigorous in newly created oceanic crust, resulting in black smokers on the seafloor and alteration of the upper kilometer of crust. Alteration minerals fill macroporosity (cracks and interpillow voids), causing increased velocities and densities, decreased porosity and permeability, and oxidation of magnetic minerals [e.g., Jacobson, 1992]. Hydrothermal circulation on ridge flanks is damped by a relatively impermeable sediment cover [Anderson and Hobart, 1976], decreased crustal permeability, and decreased thermal buoyancy forces [Anderson et al., 1977]. Nevertheless, circulation on flanks is responsible for ~70% of the advective heat loss from oceanic crust [Stein and Stein, 1994] and substantial geochemical exchange with seawater [Mottl and Wheat, 1994; Elderfield and Schultz, 1996].

[3] To decipher the interactions among hydrogeology, structure, and physical properties in the aging of oceanic crust, we must examine at least four typical end-member situations: young crust created at slow and fast spreading rates and old crust created at slow and fast spreading rates.
The first three of these situations were partly addressed at Holes 395A (young and slow), 504B (young and moderate to fast), and 418A (old and slow). Legs 129 and 185 drilled the fourth end-member: Hole 801C, located in the Pigafetta Basin at 18°38.5'N, 156°21.6'E, northwest Pacific Ocean (Figure 1). Hole 801C contained the oldest in situ oceanic crust that has been sampled to date: 166.8 ± 4.5 Ma [Pringle, 1992]; its biostratigraphic [Bartolini and Larson, 2001] and radiometric basement ages closely match its predicted age on the basis of extrapolating magnetic anomal-}

Figure 1. Locations of DSDP and ODP sites with significant penetrations into normal oceanic crust and with either downhole logging (Table 2) or core physical properties (Table 3).
thermal deposit is highly fractured and extremely altered, with a late, oxidative type of alteration that is suggestive of proximity to a hydrothermal system [Alt et al., 1992].

Although Hole 801C was drilled during Leg 129, no downhole experiments were undertaken in the hole at that time. ODP Leg 144 returned to this site and completed both packer [Larson et al., 1993] and logging [Jarrard et al., 1995] measurements. A complete suite of in situ geophysical logs was obtained through the upper 100 m of basalt, including velocity, density, neutron porosity, resistivity, three-component magnetometer, geochemistry, and Formation MicroScanner® logs. ODP Leg 185 returned to this hole and deepened it by 339.3 m for a total of 474 m penetration onto oceanic crust [Shipboard Scientific Party, 2000a]. Recovered cores consisted almost entirely of MORB tholeiites, both flows and pillows, with rare hyaloclastites and sediments plus a second hydrothermal unit (Figure 2). Downhole logging extended from a bridge at 70–90 m above total depth, up to the bottom of casing at 483 m below seafloor (mbsf). This includes the zone previously logged on Leg 144. Nearly the same suite of logs was run as on Leg 144, except that geochemical logging was confined to the spectral gamma tool. Resistivity logging on Leg 185 used the dual laterolog rather than dual induction tool.

Physical properties of the upper hydrothermal zone and overlying alkali basalts have been presented and analyzed previously [Busch et al., 1992; Wallick et al., 1992; Larson et al., 1993; Jarrard et al., 1995] and are not presented here. Our analyses focus on the tholeitic section, first encountered on Leg 129 but mostly cored and logged during Leg 185.

2.1.2. Core Measurements of Physical Properties

Busch et al. [1992] and Wallick et al. [1992] measured core-based physical properties of upper Hole 801C...
basalts. These measurements included index properties (bulk density, matrix (or grain) density, porosity) and P-wave velocity at atmospheric pressure. Wallick et al. [1992] also measured velocity versus pressure. Our analyses exclude the data of Wallick et al. [1992], because of incomplete removal of interstitial water during drying [Jarrard et al., 1995]. Leg 185 basalt physical properties sampling used a much closer sample spacing than at any previous deep crustal site: approximately three samples per core. At each horizon, a wedge-shaped sample was taken for index properties [Shipboard Scientific Party, 2000a] and an adjacent cube was cut for three-axis measurement of velocity [Shipboard Scientific Party, 2000a], resistivity and X-ray computed tomography (CT) imaging [Hirono and Abrams, 2002]. An additional 235 measurements of horizontal P-wave velocity were made on split cores.

Alteration of the Leg 185 basalt cubes was estimated with light absorption spectroscopy (LAS). LAS measures the spectrum of light reflected from a rock surface. Light is absorbed by minerals at and near the rock surface due to both electronic and vibrational processes [Clark et al., 1990; Clark, 1995]. Types of identifiable minerals depend on the frequency band of the instrument. We use the FieldSpec Pro FR Portable Spectroradiometer because of its wide bandwidth (350–2500 nm). At near-infrared wavelengths (950–2500 nm), normal modes of characteristic vibrations of OH bonds occur; diagnostic absorption bands for water, Mg-OH, Al-OH, and Fe-OH are evident.

LAS has not previously been used as a technique for basalt analysis, yet it has several advantages over traditional DSDP/ODP analytical techniques: measurements take only seconds and are nondestructive, and LAS is particularly sensitive to hydration and smectite concentration, both of which have presented challenges to interlaboratory measurement consistency. The volumetrically dominant minerals in these basalts are pyroxene and plagioclase. Plagioclase is spectrally featureless and therefore undetectable by LAS. The dominant spectral signature in these basalts is from pigeonite, the pyroxene. Montmorillonites, with an OH absorption band at 1400 nm and a strong water absorption band at 1930 nm, can be detected at concentrations of only a few percent [Vanden Berg and Jarrard, 2002]. Trough depths at these two wavelengths, each normalized to adjacent wavelengths outside the absorption band, are highly correlated ($R = 0.882$), so we combine them into a single measure of relative abundance of montmorillonite among the Leg 185 basalts. Pore water has the same OH and water absorption bands as hydration minerals, so it is important to use LAS only on thoroughly dried samples. To assure complete drying, we evacuated the samples for 24 hours at a vacuum pressure of 9–11 Pa.

Permeabilities of 13 of these cube samples were measured by Terra Tek,[3] a commercial petrophysical analysis laboratory. Initial reconnaissance testing consisted of measurements of eight relatively porous samples using a low-pressure (300 psi or 2 MPa), steady state air flow technique on jacketed samples. Unfortunately, five of the eight samples were below the technique’s lower limit of $10^{-17}$ m$^2$ (0.01 mDarcy) (Table 1). Furthermore, of the three that were measurable, one developed a crack during testing and is therefore nonrepresentative of intergranular permeability, and one has a permeability that is barely resolvable by this technique. Our second stage of testing measured permeability of eight samples with a pulse decay technique, capable of accurate permeability measurements at levels as low as $10^{-20}$ m$^2$. At these very low permeabilities, even traces of residual pore water or surface humidity may bias measurements, so care was taken to minimize both.

### 2.1.3. Downhole Logs

Five downhole logs of the Hole 801C tholeitic interval were analyzed: density, sonic (velocity), resistivity, neutron, and potassium.

Density logging uses a pad-type tool that requires a good contact with the borehole wall for reliable performance. Most previous logging of oceanic crust used density tools that frequently lost pad contact with the borehole wall; consequently, these density logs were of generally poor quality [Carlson and Herrick, 1990]. Beginning on ODP Leg 125 the hostile environment lithodensity tool was used. Use of this tool at Hole 801C resulted in a density log that appears to be of generally fair-good quality. The caliper log indicates that hole size is nearly uniform throughout most of the hole, except for poor hole conditions in the interval 627–714 mbsf, a unit consisting of thin flows and pillows (Figure 2). Loss of pad contact and resulting low-density readings are most common in this interval. We used a combination of caliper, $\Delta p$, velocity, and photoelectric factor logs to identify zones with loss of pad contact and to guide deletion of associated unreliable density data. Approximately 21% of the data were deleted in this manner. As discussed in a subsequent section, however, interlog comparisons suggest that spurious values are common even in this edited density log. Figure 2 shows the edited density log; the unedited log is shown by Shipboard Scientific Party [2000a].

The basalt interval at Hole 801C was logged with the borehole-compensated sonic tool, which generates four source-receiver travel times at each measurement point. A few of the redundant measurements at each depth were obviously unreliable due to cycle skips and “road noise” (reverberations due to tool drag over rough borehole wall). We employed a simple reprocessing algorithm [Shipboard Scientific Party, 1987] that rejected the unreliable data, permitting calculation of a more reliable velocity log. The reprocessed velocity log of Figure 2 appears to be of good quality.

<table>
<thead>
<tr>
<th>Depth, mbsf</th>
<th>Steady State</th>
<th>Pulse Decay</th>
<th>Porosity, %</th>
</tr>
</thead>
<tbody>
<tr>
<td>606.12</td>
<td>&lt;10$^{-17}$</td>
<td>8.25 x 10$^{-19}$</td>
<td>18.0</td>
</tr>
<tr>
<td>607.22</td>
<td>&lt;10$^{-17}$</td>
<td>1.38 x 10$^{-18}$</td>
<td>13.4</td>
</tr>
<tr>
<td>607.74</td>
<td>2.3 x 10$^{-16}$</td>
<td>15.6</td>
<td></td>
</tr>
<tr>
<td>615.11</td>
<td>5.3 x 10$^{-16}$</td>
<td>7.5</td>
<td></td>
</tr>
<tr>
<td>616.91</td>
<td>&lt;10$^{-17}$</td>
<td>8.6</td>
<td></td>
</tr>
<tr>
<td>620.12</td>
<td>4.04 x 10$^{-19}$</td>
<td>10.4</td>
<td></td>
</tr>
<tr>
<td>673.35</td>
<td>6.66 x 10$^{-19}$</td>
<td>7.6</td>
<td></td>
</tr>
<tr>
<td>693.07</td>
<td>1.24 x 10$^{-19}$</td>
<td>6.1</td>
<td></td>
</tr>
<tr>
<td>720.68</td>
<td>&lt;10$^{-17}$</td>
<td>5.22 x 10$^{-19}$</td>
<td>10.5</td>
</tr>
<tr>
<td>757.00</td>
<td>1.2 x 10$^{-17}$</td>
<td>11.0</td>
<td></td>
</tr>
<tr>
<td>815.17</td>
<td>3.18 x 10$^{-17}$</td>
<td>5.9</td>
<td></td>
</tr>
<tr>
<td>900.32</td>
<td>&lt;10$^{-17}$</td>
<td>2.46 x 10$^{-19}$</td>
<td>8.1</td>
</tr>
<tr>
<td>901.07</td>
<td>&lt;10$^{-17}$</td>
<td>7.7</td>
<td></td>
</tr>
</tbody>
</table>

*Sample cracked during measurement.*
17] Leg 185 resistivity logging of Hole 801C [Shipboard Scientific Party, 2000a] used the dual laterolog, which has two depths of penetration (shallow and deep). This tool provides superior accuracy to the dual induction tool (used on Leg 144) in high-resistivity formations such as some basalts, because dual-induction response saturates at resistivities near ~2000 Ω·m. In general, resistivity logs have the highest signal-to-noise ratio of any log obtained in ODP. The shallower-penetration logs are, however, sensitive to changes in hole size: In enlarged portions of the hole, they underestimate formation resistivity. Consequently, we confine our analyses to the deep-resistivity log, which has a depth of penetration so great that it is insensitive to hole size variations.

19] Rock resistivity depends mainly on resistivity of the formation fluid, which is a function of both salinity and temperature. Apparent formation factor, the ratio of rock resistivity to fluid resistivity, removes this effect. We converted core-based resistivities to apparent formation factors, assuming a fluid resistivity of 0.2 Ω·m (seawater salinity and room temperature of 21°C). Assuming seawater salinity and the temperatures measured within the borehole at the time of logging [Shipboard Scientific Party, 2000a], log resistivities were converted to apparent formation factors (Figure 2). Because the time between logging and cessation of drilling is small compared to the time spent drilling and circulating drilling fluids, we suspect that near-borehole temperatures are similar to those measured within the borehole. If the zone reached by the deep resistivity tool attained equilibrium temperature by the time of logging, then temperatures are ~10° higher than we assumed and formation factors are ~25% lower than we calculated. This uncertainty is relatively small, compared to the 2 orders of magnitude variations in formation factor observed within the logged interval.

20] An alternative approach to Leg 185 formation temperatures is to compare the upper part of the Leg 185 resistivity log to the log obtained on Leg 144. If the dual laterolog and induction tools have identical log responses, then the ratio of these logs can be converted to a log of temperature change. We found, however, that the differences between these two logs were too great to be attributable to temperature change; tool response is probably responsible.

21] ODP neutron porosity logs consistently overestimate the porosity of basalts [Anderson et al., 1985; Broglio and Moos, 1988; Lysne, 1989; Moos, 1990]. Broglio and Ellis [1990] utilized logs from five sites to analyze the relative contributions of several factors to this systematic bias. They concluded that the most important factor is the ODP technique of running the neutron porosity tool without appropriate eccentralization. At the one site (Site 735) that was logged with a bowspring-eccentralized neutron tool, this bias was absent. Although neutron tools are “compensated” in the sense of applying some correction for tool standoff from the borehole wall, Broglio and Ellis [1990] found that this compensation is insufficient for correction of neutron logs that lack eccentralization.

21] The Hole 801C neutron logs run on Legs 144 and 185 were corrected for standoff. Jarrard et al. [1995] compared Leg 144 density-based porosity to neutron porosity and found that the neutron tool overestimated porosity by 5–25%. A similar cross plot for the Leg 185 logs (not shown) exhibits a nearly identical pattern to that for Leg 144. This discrepancy, which is higher than those identified by Broglio and Ellis [1990] in other ODP crustal sites, probably results from a combination of interlayer water in alteration minerals and calibration to limestone matrix density rather than to basalt matrix density. Broglio and Ellis [1990] corrected neutron logs from several crustal sites for both effects. For Hole 801C, however, too few H2O+ analyses are available for alteration correction, which compromises the usefulness of the Hole 801C neutron logs.

22] From the DSDP and ODP Initial Reports and their associated CD-ROM databases, we have extracted four types of data: velocity logs, core plug velocities, core index properties (bulk density, matrix density, and porosity), and core-based identification of volcanic style.

23] From ~1200 sites drilled by DSDP and ODP, we selected those that fulfill the following criteria: (1) velocity logging of at least 40 m of basement and (2) basement composed of normal oceanic crust formed in an open ocean environment. This second criterion excluded sites on crust formed by back arc spreading, oceanic plateau or seamount volcanism, or at a heavily sedimented spreading center; all three environments are likely to differ from normal oceanic crust in both volcanic style (especially proportion of intrusives versus extrusives) and in hydrothermal alteration. Thirteen holes met both criteria: 395A, 396B, 417D, 418A, 504B, 556, 558, 564, 750D, 801C, 843B, 896A (Table 2 and Figure 1). Basement logs for these sites [Kirkpatrick, 1978; Salisbury et al., 1979; Cann and Von Herzen, 1983; Hill and Cande, 1985; Broglio and Moos, 1988; Moos, 1990; Jarrard and Broglio, 1991; Goldberg and Moos, 1992; Jarrard et al., 1995; Shipboard Scientific Party, 1978a, 1978b, 1979a, 1979b, 1983, 1985a, 1985b, 1985d, 1990a, 1990b, 1990c, 1993, 1998, 2000a] vary in length, from a minimum of 41 m at Site 843 to 1827 m at Hole 504B. Most sites have 90–200 m of velocity log (Figure 3), and only three (395A, 418A, and 504B) have more than 300 m. We confined our analyses to the top 300 m of basement, thereby preventing our intersite comparisons from being biased by intrasite velocity gradients. Except for a possible velocity gradient within the top 30 m of basement, no systematic depth-dependent changes in velocity are evident within the top 300 m of basement (Figure 3).

24] Most DSDP and ODP velocity logging has utilized the Schlumberger borehole-compensated sonic tool, which determines P wave travel time by a first-break thresholding criterion: the first energy exceeding a preset threshold is assumed to be that of the initial P wave peak [e.g., Serra, 1984]. Raw velocity logs for basalts generally have many unreliable intervals, because cycle skips and “road noise” (reverberations due to tool drag over rough borehole wall) induce failure of the threshold technique. For most sites, we employed a simple reprocessing algorithm [Shipboard Scientific Party, 1987] that we have developed to take advantage of redundancy of measurements, reject the unreliable data, and use retained data to calculate a more reliable velocity log. Original first-break travel time logs were not available for Sites 396 and 417, so their velocity logs could
Table 2: Sites With Significant Downhole Logging of Normal Oceanic Crust

<table>
<thead>
<tr>
<th>Site</th>
<th>Age (Ma)</th>
<th>Spreading Rate (mm/yr)</th>
<th>Basement Depth (m)</th>
<th>Log Interval (mbsf)</th>
<th>Macroporosity (%)</th>
<th>Velocity (klm/s)</th>
<th>Potassium (K, %)</th>
</tr>
</thead>
<tbody>
<tr>
<td>395A</td>
<td>7.0</td>
<td>17</td>
<td>92.0</td>
<td>92–393</td>
<td>0.9 ± 3.1</td>
<td>4.55 ± 0.03</td>
<td>0.40 ± 0.006</td>
</tr>
<tr>
<td>396B</td>
<td>9.4</td>
<td>20</td>
<td>150.5</td>
<td>214–336</td>
<td>3.0 ± 3.0</td>
<td>4.25 ± 2.2</td>
<td>0.56 ± 0.011</td>
</tr>
<tr>
<td>417D</td>
<td>10.8</td>
<td>15</td>
<td>343</td>
<td>343–448</td>
<td>0.4 ± 3.2</td>
<td>3.3 ± 1.7</td>
<td>0.22 ± 0.008</td>
</tr>
<tr>
<td>418A</td>
<td>11.0</td>
<td>15</td>
<td>324</td>
<td>324–624</td>
<td>2.5 ± 3.2</td>
<td>3.7 ± 1.7</td>
<td>0.35 ± 0.007</td>
</tr>
<tr>
<td>556</td>
<td>34</td>
<td>10</td>
<td>284</td>
<td>284–336</td>
<td>2.2 ± 1.9</td>
<td>4.4 ± 2.5</td>
<td>0.54 ± 0.01</td>
</tr>
<tr>
<td>558</td>
<td>35.5</td>
<td>11</td>
<td>408</td>
<td>408–543</td>
<td>3.2 ± 1.6</td>
<td>3.8 ± 1.8</td>
<td>0.40 ± 0.012</td>
</tr>
<tr>
<td>770C</td>
<td>42</td>
<td>45</td>
<td>423.2</td>
<td>427–519</td>
<td>3.0 ± 3.0</td>
<td>4.7 ± 2.5</td>
<td>0.40 ± 0.01</td>
</tr>
<tr>
<td>801C</td>
<td>167</td>
<td>90</td>
<td>530</td>
<td>530–820</td>
<td>2.8 ± 1.9</td>
<td>4.4 ± 2.5</td>
<td>0.54 ± 0.01</td>
</tr>
<tr>
<td>843B</td>
<td>110</td>
<td>20</td>
<td>242.5</td>
<td>257–297</td>
<td>2.2 ± 1.9</td>
<td>4.7 ± 2.5</td>
<td>0.40 ± 0.012</td>
</tr>
<tr>
<td>896A</td>
<td>5.9</td>
<td>38</td>
<td>179</td>
<td>179–403</td>
<td>1.9 ± 3.1</td>
<td>4.7 ± 2.5</td>
<td>0.54 ± 0.01</td>
</tr>
</tbody>
</table>

For Sites 418 and 504, we used velocity logs calculated by semblance analysis of waveforms [Moos et al., 1990]. Velocity and resistivity are both controlled primarily by porosity, but resistivity measurements are less affected by poor borehole conditions. We removed velocity values in intervals where velocity spikes to low values were inconsistent with resistivity character.

Core velocity measurements used here (Figure 3) were shipboard measurements, made at atmospheric pressure and reported by the shipboard scientific parties. For some sites, postcruise studies determined velocity as a function of pressure, but we have chosen not to mix the different data types.

Volcanic style is evaluated from core descriptions by Shipboard Scientific Party [1978a, 1978b, 1979a, 1979b, 1983, 1985a, 1985b, 1985d, 1990a, 1990b, 1990c, 1992, 1993, 2000a]. When core recovery is high, distinguishing pillow basalts from sheet flows is relatively straightforward. Often, however, low core recovery lends ambiguity to this discrimination. For example, Hole 843B yielded only low-recovery spot cores that were all described as flows, though Formation MicroScanner logs show that some pillows are present [Goldberg and Moos, 1992]. If core recovery is sufficient to discriminate flows from pillows, then the continuous records provided by logs usually permit a refinement of unit boundaries because geophysical properties generally change much more at the boundaries of such units than within them. These slight revisions of boundaries, either shipboard or shore-based [e.g., Broglio and Ellis, 1990], are not always applied, so we have redone them for all logged sites. Similarly, we have confirmed or modified the log depth shifts for all sites; in some cases, cable stretch has necessitated up to several meters constant depth shift of the “final” logs available from CD-ROM and the Lamont-Doherty Earth Observatory log data repository. Although nearly all penetrated basement rocks are either pillows or massive sheet flows, several other kinds of igneous rocks are encountered occasionally. Brecias are the most common of these; we group breccias with pillows, because the boundary between the two is gradational in morphology and probably also in extrusive intensity. At the shallow subbasement depths considered here, sills are rare. Recovery of sediment interlayers is usually less than a few centimeters, partly because of core recovery bias. Each of the other encountered rock types is present at only one site: basaltic sand and gravel at Hole 396B, serpentinized gabbro at Site 556, serpentinite at Site 558, and hydrothermal deposits at Hole 801C.

A second search of the DSDP and ODP databases and publications (Initial Reports and Scientific Results) focused on index properties of basalts (Table 3). Several DSDP sites have measurements of basalt bulk density but not porosity or matrix density; results from these are tabulated by Johnson and Semyan [1994] but excluded from our analyses. Again, only sites on normal oceanic crust were considered; data from sedimented ridges (e.g., Juan de Fuca and Gulf of California) were excluded. Data from multiple holes at the same site were combined, except for two sites (417A&D, 1149B&D) at which the holes were widely spaced or clearly in contrasting hydrothermal settings (e.g., altered basement high versus less altered basement low). Useful index data are available for 25 sites.
Table 3 and Figure 1), almost double the number of sites with velocity logs (Table 2).

3. Site 801 Analysis

3.1. Permeability

Permeability is the architect of the crust’s hydrothermal systems. Nearly all measurements of the permeability of upper oceanic crust are averages for portions of a borehole tens to hundreds of meters in extent. Fisher [1998] provides an excellent review and synthesis of these bulk permeability measurements and their implications. This large-scale crustal permeability is a critical control on hydrothermal circulation patterns and associated heat flux because the volumetrically dominant flux is channelized within the most permeable zones, particularly large open fractures [Fisher, 1998; Fisher and Becker, 2000].

Bulk permeabilities for the top 1200 m of oceanic crust suggest two main layers: the top ~500 m has permeabilities of $\sim 10^{-14} - 10^{-13}$ m$^2$, and the next 700 m has permeabilities of $\sim 10^{-17}$ m$^2$ [Fisher, 1998]. These bulk permeabilities are all from <8 Ma crust. A single measurement from older crust is that at Hole 801C, where testing of a 93-m interval indicated a bulk permeability of $\sim 8 \times 10^{-14}$ m$^2$; more likely, flow is confined to an 18-m hydrothermal zone at the top of the tholeiites, and bulk permeability is accordingly higher, $\sim 4 \times 10^{-13}$ m$^2$ [Larson et al., 1993]. Drilling-induced temperature anomalies measured on
Table 3. Average Index Properties of Basalts for Sites on Normal Oceanic Crust, With 95% Confidence Limits

<table>
<thead>
<tr>
<th>Site</th>
<th>Hole</th>
<th>Age, Ma</th>
<th>Matrix Density, g/cm³</th>
<th>Matrix Density, % Porosity, g/cm³</th>
<th>Bulk Density, g/cm³</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td></td>
<td></td>
<td>&gt; 5%</td>
<td>95% CL</td>
<td>95% CL</td>
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<tr>
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<td></td>
<td></td>
<td></td>
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<td></td>
</tr>
<tr>
<td>332</td>
<td>A&amp;B</td>
<td>3.5</td>
<td>2.963 ± 0.015</td>
<td>2.942 ± 0.016</td>
<td>2.800 ± 0.018</td>
</tr>
<tr>
<td>333</td>
<td>A</td>
<td>3.5</td>
<td>2.974 ± 0.015</td>
<td>2.974 ± 0.015</td>
<td>2.754 ± 0.013</td>
</tr>
<tr>
<td>334</td>
<td></td>
<td>8.9</td>
<td>2.992 ± 0.032</td>
<td>4.75 ± 3.13</td>
<td>2.897 ± 0.088</td>
</tr>
<tr>
<td>395</td>
<td>&amp;A</td>
<td>7.0</td>
<td>2.954 ± 0.018</td>
<td>2.955 ± 0.010</td>
<td>2.852 ± 0.016</td>
</tr>
<tr>
<td>396</td>
<td>B</td>
<td>9.4</td>
<td>2.847 ± 0.086</td>
<td>2.855 ± 0.030</td>
<td>2.752 ± 0.039</td>
</tr>
<tr>
<td>409</td>
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<td>2.4</td>
<td>2.970 ± 0.022</td>
<td>2.970 ± 0.022</td>
<td>2.650 ± 0.028</td>
</tr>
<tr>
<td>410</td>
<td>&amp;A</td>
<td>9</td>
<td>2.903 ± 0.035</td>
<td>2.903 ± 0.035</td>
<td>2.456 ± 0.038</td>
</tr>
<tr>
<td>412</td>
<td>A</td>
<td>1.6</td>
<td>3.010 ± 0.013</td>
<td>3.005 ± 0.011</td>
<td>2.897 ± 0.013</td>
</tr>
<tr>
<td>417</td>
<td>A</td>
<td>108</td>
<td>2.835 ± 0.028</td>
<td>2.852 ± 0.026</td>
<td>2.698 ± 0.081</td>
</tr>
<tr>
<td>417</td>
<td>D</td>
<td>108</td>
<td>2.924 ± 0.018</td>
<td>2.925 ± 0.013</td>
<td>2.796 ± 0.022</td>
</tr>
<tr>
<td>418</td>
<td>A</td>
<td>110</td>
<td>2.878 ± 0.017</td>
<td>2.899 ± 0.016</td>
<td>2.742 ± 0.042</td>
</tr>
<tr>
<td>470</td>
<td>A</td>
<td>15</td>
<td>2.940 ± 0.050</td>
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<td>2.865 ± 0.037</td>
</tr>
<tr>
<td>504</td>
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<td>5.9</td>
<td>2.981 ± 0.007</td>
<td>2.987 ± 0.005</td>
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</tr>
<tr>
<td>505</td>
<td>B</td>
<td>4</td>
<td>2.988 ± 0.033</td>
<td>2.986 ± 0.023</td>
<td>2.864 ± 0.038</td>
</tr>
<tr>
<td>543</td>
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<td>80</td>
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<td>2.898 ± 0.052</td>
<td>2.788 ± 0.072</td>
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<td>556</td>
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<tr>
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<td>2.995 ± 0.023</td>
<td>2.839 ± 0.029</td>
</tr>
<tr>
<td>597</td>
<td></td>
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</tr>
<tr>
<td>648</td>
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<td>2.999 ± 0.041</td>
<td>2.998 ± 0.015</td>
<td>2.897 ± 0.023</td>
</tr>
<tr>
<td>765</td>
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<td>128</td>
<td>2.835 ± 0.031</td>
<td>2.867 ± 0.016</td>
<td>2.792 ± 0.023</td>
</tr>
<tr>
<td>770</td>
<td>B&amp;C</td>
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<td>2.839 ± 0.042</td>
<td>2.845 ± 0.025</td>
<td>2.747 ± 0.031</td>
</tr>
<tr>
<td>801</td>
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<td>167</td>
<td>2.880 ± 0.013</td>
<td>2.900 ± 0.016</td>
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</tr>
<tr>
<td>896</td>
<td>A</td>
<td>5.9</td>
<td>2.936 ± 0.016</td>
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<td>2.842 ± 0.027</td>
</tr>
<tr>
<td>1149</td>
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<td>2.796 ± 0.074</td>
<td>2.796 ± 0.074</td>
<td>2.616 ± 0.134</td>
</tr>
<tr>
<td>1149</td>
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<td>132</td>
<td>2.829 ± 0.027</td>
<td>2.832 ± 0.026</td>
<td>2.682 ± 0.038</td>
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</table>

Leg 185 clearly delineates this high-permeability zone, as well as a thinner permeable zone at 705–712 mbsf [Shipboard Scientific Party, 2000a]. This deeper zone is very porous, with the lowest formation factors and velocities of the entire tholeitic interval (Figure 2).

[30] Bulk permeability measurements cannot address the problem of intergranular-scale permeability and the associated fluid flow that is responsible for pervasive intergranular-scale crustal alteration. Intergranular permeabilities of most rocks are controlled by pore geometry, which in turn is dependent on porosity and lithology. This dependence commonly results in a linear relationship between porosity and ln(permeability) for all rocks from the same formation [Nelson, 1994]. Very few core plug measurements of basalt permeability have been undertaken. Young basalts from DSDP Leg 70 (0.5–2.7 Ma) and Hole 504B (5.9 Ma) have permeabilities of mostly 10−20 to 10−18 m² that are weakly dependent on porosity [Karato, 1983a, 1983b]. Permeabilities of 110 Ma crust at Holes 417D and 418A are generally 10−21 to 10−17 m² [Karato, 1979a; Hamano, 1979] and moderately correlated with porosity. These intergranular permeabilities are ~3–8 orders of magnitude lower than bulk permeabilities for the upper layer, which in turn are less than regional-scale permeabilities [Fisher et al., 1994; Fisher and Becker, 2000]. Other basalt permeability measurements may not be representative of open ocean crust: Juan de Fuca sediments [Christensen and Ramanantoandro, 1988], Tonga-Kermadec [Christensen and Ramanantoandro, 1988], and Arctic Ocean [Aksyuk et al., 1992].

[31] Our samples from Hole 801C (Table 1), along with mostly lower-porosity samples from other sites on normal oceanic crust [Johnson, 1979a; Hamano, 1979; Karato, 1983a, 1983b], exhibit a linear relation between porosity and ln(permeability) (Figure 4a). Compared to typical trends for sedimentary rocks, however, the basalt scatter is much higher and permeabilities are lower by ~3 orders of magnitude. Both differences are attributable to the isolation of most vesicles from available flow paths. The relationship between permeability and porosity, though noisy, indicates that porosity can be used as an indirect indicator of permeability and therefore of relative fluid flux. Permeability controls alteration by determining water-rock ratio. Consequently, measurements of LAS-based alteration index for Hole 801C are correlated with both permeability (Figure 4b) and porosity (Figure 5c). On the basis of our small sample from Hole 801C, alteration does not appear to cause a major change in the porosity/permeability relation.

[32] On the basis of the freshness of most non-801C samples in Figure 4a and the regression trend of Figure 4b, the minimum permeability for sufficient water-rock ratio to generate significant alteration is ~10−19 m², which occurs at a porosity of ~5%. Flows, with porosities of mostly <5%, consequently undergo much less alteration than more porous pillows. However, these threshold permeability and porosity values are only generalizations, because water-rock ratio and alteration depend not only on local intergranular permeability but also on proximity to high-permeability cracks.

3.2. Alteration

[33] Basalt physical properties at the centimeter scale measured in core plugs may differ from those at the meter scale measured by logs, because of differences in type of porosity and in alteration history. At plug scale, porosity consists mainly of vesicles and microcracks, whereas log-scale porosity also includes fractures. At plug scale, alteration converts glass, olivine, and plagioclase mainly to clay minerals, thereby decreasing matrix density and matrix velocity. Porosity may increase due to this hydration, or
decrease due to precipitation of carbonates or zeolites. At log scale and larger, porosity is thought to decrease due to filling of cracks and interpillow voids by alteration minerals, carbonates, quartz/chalcedony, or Fe-oxyhydroxides.

Basalt core descriptions provide only a subjective description of variations in alteration. The $\text{H}_2\text{O}^+$ of basalts, as measured by CHN analyzer, does provide a semiquantitative measure of the extent of alteration-induced hydration [Alt et al., 1992]. Typical $\text{H}_2\text{O}^+$ contents of 3–6 wt % for the upper extrusives contrast with initial MORB $\text{H}_2\text{O}^+$ of only 0.12 wt % in fluid inclusions [Sobolev and Chausssion, 1996]. For the portion of Hole 801C cored during Leg 129, the index property measurements of Busch et al. [1992] were paired with geochemical analyses of Alt et al. [1992], permitting demonstration of a close correlation between increasing hydration (greater $\text{H}_2\text{O}^+$) and decreases in velocity, matrix density, and matrix velocity [Busch et al., 1992; Jarrard et al., 1995]. An association between alteration and porosity was also evident, probably because of two factors: (1) alteration increases porosity and (2) high porosity promotes alteration because of high permeability and therefore enhanced fluid flow. Figure 5b confirms that matrix densities of Hole 801C tholeiites decrease with increasing porosity, although the relationship is weaker than was observed for the short tholeiite interval cored during Leg 129.

CHN-based hydration measurements were not available for the Leg 185 index property samples, so we determined relative variations in hydration with LAS. Figure 5a shows that matrix density for these basalts is only subtly correlated with LAS-based hydration, in contrast to the much better association between matrix density and hydration for the shorter interval analyzed by Jarrard et al. [1995]. Increasing hydration is associated with increasing porosity (Figure 5c), with a higher correlation ($R = 0.69$) than observed by Jarrard et al. [1995].

### 3.3. Density

Formation bulk density ($\rho_b$) depends on fractional porosity ($\phi$), matrix (or grain) density ($\rho_{ma}$), and fluid density ($\rho_f$), according to the relationship $\rho_b = \rho_{ma} (1 - \phi) + \rho_f \phi$. Matrix density is the average density of the minerals forming the solid part of the rock, including any alteration minerals. This definition and an analogous one for matrix velocity are universal in physical properties analyses but not log analyses. Many log interpretations, particularly hydrocarbon-related ones, confine matrix density and matrix velocity to the “clean” portion of a rock, often with corrections for any clay content. Fractional porosity is generally much more variable than matrix density and fluid density. Thus the density log is a relatively straightforward porosity log, and both $\rho_{ma}$ and $\rho_f$ often can be assumed to be constant. For upper crustal basalts, however, matrix density often decreases with increasing porosity, because of greater alteration in the more porous and therefore more permeable rocks. This pattern has been observed in several other studies [Hamano, 1979; Christensen et al., 1980; Carlson and Herrick, 1990; Jarrard and Broglio, 1991], including prior analyses of the upper basalts from Hole 801C [Busch et al., 1992; Jarrard et al., 1995]. Jarrard et al. [1995] modified the equation above to include this effect, based on the upper 801C basalts, and we used their equation to convert the new Hole 801C density log into a porosity log.

### 3.4. Velocity

Porosity is the most important variable controlling both density and velocity in the 801C basalts. This common control imparts a strong correlation between density and velocity, routinely noted for individual sites and summarized for core plugs from multiple sites by Busch et al. [1992]. Figures 6a and 6b compare velocities to densities for the 801C tholeiites on the basis of core and log data.

The relation between density and velocity can be predicted on the basis of a combination of the theoretical density equation above and the Wyllie et al. [1956] equation: $V^{-1} = (1 - \phi)V_{ma} + \phi/V_f$, where $V$ is whole rock compressional wave velocity, $\phi$ is fractional porosity, $V_{ma}$ is matrix velocity, and $V_f$ is pore fluid velocity. This equation
is only an empirical approximation; theoretical equations [e.g., Gassmann, 1951; Kuster and Toksoz, 1974] require elastic moduli that are rarely measured. Matrix velocities for basalts [e.g., Serra, 1986] vary due to changes in alteration and composition (particularly proportion of mafic minerals). The Site 801 basalts cored during Leg 129 have a zero-porosity matrix velocity of 6.5–6.8 km/s [Jarrard et al., 1995], whereas those cored during Leg 185 have a matrix velocity of 6.0–6.3 km/s.

Assuming a matrix velocity of 6.3 km/s and matrix density of 3.0 g/cm³, the predicted relationship between velocity and density at Site 801 is shown by a curve on Figures 6a and 6b. This curve fits the lowest-porosity core and log data reasonably well, but it systematically mismatches higher-porosity data. We attribute this mismatch to porosity-dependent variations in matrix density and matrix velocity, caused by increased alteration at higher porosities. Jarrard et al. [1995] noted a similar pattern for core plug data from upper 801C. They found, in contrast, that a predicted curve incorporating regression-based porosity-dependent matrix densities and matrix velocities fit not only the 801C core plug data but also the synthesis by Busch et al. [1992] of core plug measurements from various sites. Figures 6a and 6b confirm that this pattern holds throughout the deepened tholeiitic interval at 801C. The pattern holds for both core and log data, despite differences in type of porosity.

Several other studies have examined the relationship between basalt porosity and velocity, based on core or log data. Those that have examined variations in matrix velocity have noted reduced matrix velocities at high porosity, attributed to alteration [Carlson and Herrick, 1990; Jarrard and Broglia, 1991]. A velocity/porosity cross plot for Hole 801C (not shown) indicates that low-alteration samples (based on LAS) are only subtly higher in velocity, by ~200 m/s, than more altered samples.

3.5. Formation Factor

The deep and shallow laterologs differ not only in depth of penetration but also in overall current geometry: the shallow measurement path is predominantly vertical and the deep path is more horizontal. Pezard and Anderson [1989] and later Pezard et al. [1996] used this distinction to identify basalt intervals at Hole 504B with dominantly

![Figure 5. Relationships among porosity, alteration (based on light absorption spectroscopy, or LAS), matrix density, and apparent formation factor, based on core plug measurements for tholeiites from Hole 801C. In general, higher porosities promote hydration reactions generating clays, thereby increasing LAS-based alteration intensity and lowering matrix density and formation factor. Solid dots: basalts; open circles: interpillow material.](image)
horizontal ($R_{\text{deep}} < R_{\text{shallow}}$) or vertical ($R_{\text{deep}} > R_{\text{shallow}}$) fractures, and a similar approach has been used at nearby Hole 896A [de Larouzière et al., 1996]. Comparison of the deep and shallow laterologs for Hole 801C tholeiites shows generally very close agreement [Shipboard Scientific Party, 2000a]. A cross plot, not shown here, indicates a nonlinear relationship: Shallow resistivity is systematically lower than deep resistivity at the highest and lowest resistivities. This behavior implies that differences in tool response (internal calibrations?) overwhelm any possible influence of resistivity anisotropy on the logs.

[42] Our analyses use formation factor ($F = R_o/R_m$) rather than the resistivity log ($R_o$), thereby removing the influence of variations in fluid resistivity ($R_m$). If clays have a minor

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Figure 6. (left) Core-based and (right) log-based relationships among velocity, bulk density, and apparent formation factor for Hole 801C tholeiites. Log porosities are calculated from the bulk density log. Also shown for comparison to the data of Figures 6a and 6b are predicted relationships assuming either constant matrix values or porosity-dependent matrix values. Note that porosity-dependent matrix values give the most successful prediction of observed variations. Open circles: basalts; crosses: interpillow material.
contribution to formation factor, then \( F \) is related to fractional porosity by the Archie [1942] equation, as generalized by Winsauer and McCardell [1953]: \( F = \alpha \phi^{-m} \). Surface conduction on clays can have a significant effect on basalt resistivity, particularly at elevated temperatures [Olhoeft, 1981]. Consequently, calculations of porosity from resistivity for Hole 504B are based on log-based estimates of both pore fluid and clay conduction [Pezard, 1990; Pezard et al., 1996]. At Sites 768 and 770, in contrast, clay conduction was detectable but small enough to neglect without substantially biasing porosity determination [Jarrard and Schaar, 1991]. Our analyses for Hole 801C do not include estimates of clay conduction, so we calculate apparent formation factors rather than intrinsic, clay-free formation factors.

[41] Creation of hydrous minerals during alteration is expected to decrease both apparent formation factor and matrix density. As previously mentioned, this process is often most advanced in the highest-porosity, most permeable zones. The combination of high pore fluid conduction and high clay conduction in porous, altered zones generates a correlation between apparent formation factor and matrix density, as seen in Figure 5d.

[42] Several previous studies of basalt cores or logs have used a cross plot of either apparent or intrinsic formation factor versus porosity to estimate \( a \) and \( m \). For example, Pezard [1990] found from cores at Hole 504B that \( F = 10 \phi^{-1.15} \). For core and log data from Sites 768 and 770, estimates of \( a \) range from 2.2 to 6.5, and estimates of \( m \) range from 1.2 to 1.6 [Jarrard and Schaar, 1991]. For the very old crust of Hole 418A, Broglio and Moos [1988] found from logs that \( F = 11.5 \phi^{-1.35} \) in the uppermost altered interval, and \( F = 29.5 \phi^{-1.16} \) in the lower portion of that site. Similarly, the three major lithostratigraphic subdivisions (alkalic basalts, hydrothermal zone, and tholeiitic basalts) of Hole 801C have subtly different relations between apparent formation factor and porosity [Jarrard et al., 1995]. For the tholeiites, the highest-porosity rocks indicate a linear trend with \( a \) of \( \sim 4 \rightarrow 10 \), but low-porosity zones exhibit little apparent correlation of formation factor to porosity, probably because of porosity errors.

[43] Figures 6c and 6d compare Hole 801C apparent formation factors to porosity for both core plug and log data. Compared to the close relationship between these parameters shown by core data, log data have high scatter. We attribute this scatter to porosity errors, as was the case for Leg 144 logs, due to inadequate contact of the density tool pad with the borehole wall. If equilibrium temperatures had been assumed for the calculation of apparent formation factors from resistivities, rather than assuming some formation cooling by drilling circulation, the pattern in Figure 6d would be offset slightly upward; this uncertainty is minor in comparison to the range of formation factors and their scatter. For the core data, linear regression indicates that \( F = 2.6 \phi^{-2.1} \), within the range of relationships observed at other basalt sites. The log-based pattern, though high in scatter, appears to be systematically offset to the left of the core-based pattern (lower porosities for a given formation factor).

[44] Figures 6e and 6f compare apparent formation factors to velocity, for both core plug and log data. Because both depend on porosity, a positive correlation is expected. Both core plug and log data show a good correlation between these parameters. Comparison of the two trends indicates generally lower formation factors, for a given velocity, for log data than for core data. Velocity difference between in situ and laboratory pressures is unlikely to be responsible for this offset between trends, because rebound here is small [Wallick et al., 1992] and reduces both velocity and formation factor.

[47] The cross plots of Figure 6 indicate that apparent formation factor is the geophysical property that differs most between core and log scales. This conclusion is confirmed by overlays of core and log measurements for velocity, density, and formation factor versus depth (Figure 2). Apparent formation factors are lower for logs than for core plugs. Either subvertical cracks (which are low resistivity but not seen by velocity logs) or crack filling by clays (which are conductive yet increase velocity) may explain this difference.

[48] The relationships among log-based velocity, formation factor, and density seen in Figure 6 differ substantially in data scatter. Velocity is much better correlated with formation factor than either of these parameters is correlated with density. Comparison of depth plots of the three logs (Figure 2) confirms this conclusion. We attribute the density discrepancies to artifacts within the density log, caused by poor pad contact. Although the Hole 801C density log is one of the best quality density logs obtained by either DSDP or ODP from oceanic crust, its reliability is marginal. Fortunately, velocity and resistivity logs are just as sensitive to porosity variations as is the density log, and they have much higher signal-to-noise ratios than does density logging in basalts.

[49] Velocity/formation factor relationships for oceanic basalts have not been analyzed previously, but some porosity/formation factor log analyses have used sonic-based porosities because density-based porosities were poor or unavailable [Jarrard and Broglio, 1991; de Larouzière et al., 1996]. At log scale, the velocity/formation factor relation may depend on alteration intensity [de Larouzière et al., 1996]. For Hole 801C core plugs, however, the least and most altered samples lie along the same velocity/formation factor trend.

3.6. Potassium Enrichment

[50] Is the plug-based correlation between alteration and porosity (Figure 5c) also evident in the continuous records available from in situ logs? We investigated this question by comparing log measurements of velocity and potassium (Figure 7).

[51] The potassium content of altered basalts is often an order of magnitude higher than that of fresh basalts, due to potassium absorption from seawater during low-temperature alteration [Hart, 1969]. Potassium-rich saponite and celadonite are commonly formed during low-temperature diagenesis [e.g., Alt and Honnorez, 1984]. Consequently, Broglio and Moos [1988] found that both the potassium log and total gamma ray log at Hole 418A were highly correlated with qualitative ratings of core alteration, and both indicate variation in content of potassium-rich alteration minerals at that site. A relationship between gamma ray and either velocity, resistivity, or fracture intensity logs has been observed at Holes 395A [Moos, 1990], 765D [Shipboard Scientific Party, 1990a], and 896A [de Larou-
...such zones are less likely to be recovered by coring. This strong log-based correlation between porosity and potassium-bearing alteration minerals is consistent with a core-based correlation for the upper tholeiites [Busch et al., 1992] and our core-based correlations between porosity and hydrous minerals. Although the log-based pattern could be generated by fluid flow and associated precipitation either in cracks or at an intergranular scale, the core-based patterns require intergranular flow and alteration.

3.7. Synthesis

[54] Hole 801C tholeiites demonstrate the relationships among hydrologic properties, geophysical properties, and crustal alteration (Figure 8a). At an intergranular scale, high porosity generates permeability that promotes fluid flux and resulting alteration (Figure 8a). Alteration, in turn, may increase intergranular porosity and permeability (Figure 8a). Greater alteration, as indicated by H$_2$O$^+$ [Jarrard et al., 1995] or LAS, is associated with both greater porosity (Figure 5c) and permeability (Figure 4b). Alteration also reduces matrix densities and matrix velocities by up to 10%. The relationship between velocity and density is best fit by a model in which higher porosities are associated with alteration-induced reductions in matrix velocity and matrix density (Figure 6a).

[55] Whereas plugs reveal intergranular patterns, coring misses cracks and crack filling that potentially control macroporosity and large-scale velocity [Hyndman and Drury, 1976]. Log analysis, in contrast, allows us to determine whether the laboratory sample-scale observations also hold at the interflow scale. Our comparison of Hole 801C log velocity to density shows a pattern (Figure 6b) that agrees with core data (Figure 5a). In contrast, the relationship of log formation factor to velocity is offset from that seen at the intergranular scale (Figures 6e and 6f), possibly because of conductive clays in cracks. The potassium log confirms that greatest alteration is correlated with highest porosities and lowest velocities (Figure 7). The net effect is that the most altered zones remain lower in velocity than the lower-porosity, lower-permeability, unaltered zones. An extreme example is at 711 mbsf (Figures 2 and 7): this probable crack still has the highest permeability, lowest velocity, and lowest formation factor within the tholeiites, yet it has the most hydrous minerals (based on the potassium log).

[56] These observations at Hole 801C lead to the following tentative predicted effects of age-dependent crustal alteration on evolution of geophysical properties: (1) ongoing alteration causes significant decreases in intergranular (core plug) matrix density, matrix velocity, bulk density, and velocity, possibly accompanied by subtle increase in intergranular porosity; (2) alteration reduces macroporosity; (3) the alteration impact on log-scale velocity involves the competing effects of fracture filling and intergranular velocity reduction; (4) alteration increases potassium, at both core and log scales; and (5) these patterns will be more evident in pillows than in flows because of permeability differences.

4. Temporal Variations in Physical Properties of Upper Oceanic Crust

[57] The evolution of hydrothermal circulation is well established qualitatively. Hydrothermal circulation is most...
Figure 8. Flowcharts of the interrelations among hydrologic and geophysical parameters during the aging and alteration of oceanic crust. (a) Observed pattern at Hole 801C, based on core plug measurements. (b) Generally accepted “standard model.” (c) Expected relationships at the interflow scale, based on the standard model; some links have been observationally verified, whereas others are only assumed. Note that alteration increases large-scale velocities by reducing macroporosity, whereas it decreases intergranular velocity. Modified from Jarrard et al. [1995].

4.1. Previous Evidence for Timing of Waning of Hydrothermal Circulation

[58] The temporal evolution of hydrothermal circulation in oceanic crust is an outstanding problem. Different geophysical techniques suggest that ridge flank hydrothermal circulation terminates at ~5–10 Ma, >10 Ma, ~10–40 Ma, ~65 Ma, ~100 Ma, or beyond the Cretaceous.

[60] Global data syntheses of upper crustal seismic velocities [Grevelmeyer and Weigel, 1996; Carlson, 1998] indicate that velocities increase approximately linearly with the logarithm of time from ~0.1 Ma to ~5–10 Ma, but changes after that are subtle or absent. Grevelmeyer et al. [1999] demonstrate this systematic increase within a single 0–8 Ma heat flow and seismic velocity transect on the west flank of the East Pacific Rise.

[61] Bulk permeabilities decrease exponentially from 1 to 8 Ma and may continue to decrease, though subsequent changes are undetermined [Fisher and Becker, 2000]. Most of these permeability data are from sedimented ridges, in crustal permeability due to crustal alteration [Anderson et al., 1977; Stein and Stein, 1994]. Of these three mechanisms by which hydrothermal circulation is reduced with time and distance from the ridge, crustal alteration has received the most attention, perhaps because only it can account for the age dependence of upper crustal velocities [Schreiber and Fox, 1976, 1977]. Honnorez [1981], Thompson [1991], and Alt [1995] provided comprehensive reviews of low-temperature crustal alteration processes. Alteration minerals eventually fill cracks and interpillow voids, thereby reducing porosity and permeability and increasing velocity (Figure 8b). This model suggests that hydrothermal circulation is ultimately self-limiting. It is, however, opposite to the pattern observed at Hole 801C (Figure 8a) in a fundamental respect: greater alteration at Hole 801C is associated with high porosity and low velocity. This contradiction may be only apparent: the highest-porosity zones probably experience the most alteration and associated porosity reduction, yet they remain more porous than fresher, more massive intervals.

[58] Of the many associations that are expected to result from crustal aging, some are only assumed rather than directly observed (Figure 8c), and others have been observed only in young crust. The correlation between greater alteration and greater velocity, assumed to be present at the multiflow scale (Figure 8c), has only been demonstrated at one site, Hole 418A [Broglia and Moor, 1988]. Alteration, though extremely heterogeneous, certainly does generally increase with crustal age very near the ridge crest, and abundance of crack-filling carbonate veins is higher in old crust than in 8 Ma crust [Alt and Teagle, 1999]. Hole 504B provides direct evidence for alteration-induced permeability reduction [Anderson and Zoback, 1982; Anderson et al., 1985; Becker, 1989]. Bulk permeability decreases systematically with crustal age, for the first 8 Myr [Fisher and Becker, 2000], but packer experiments at Hole 801C found some of the highest crustal permeability ever observed [Larson et al., 1993]. To evaluate both the generality of Hole 801C observations and the inferred patterns of crustal aging, it is necessary to undertake a global examination of temporal variations in physical properties of upper oceanic crust.
which may have alteration histories that are not representative of open ocean crust.

[62] Dating of alteration minerals within upper oceanic crust suggests that alteration terminates by ~10–40 Ma [Peterson et al., 1986; Gallahan and Duncan, 1994; Hart et al., 1994]. Most of these dates are <15 Myr younger than the age of surrounding crust, but some indicate that formation of alteration minerals continues for ~40 Myr [Gallahan and Duncan, 1994].

[63] The global heat flow synthesis of Stein and Stein [1994] shows that the heat flow deficit (theoretical minus observed) gradually decreases with age, disappearing at about 65 Ma. The deficit indicates advection via open circulation between seawater and the crust, with narrow unsampled zones of upwelling heat loss. Beyond 65 Ma, heat flow is dominantly conductive. The scaling age depends primarily on crustal age and only secondarily on sediment thickness, implying that it is caused mainly by decreased crustal permeability and/or reduced temperature contrast [Stein and Stein, 1994; Stein et al., 1995]. Independent evidence of hydrothermal circulation in 15–70 Ma crust comes from geochemical gradients within pore fluids of equatorial Pacific sediments [Baker et al., 1991].

[64] Velocities and densities of basalt core plugs decrease as crust ages, until ~100 Ma [Johnson and Semyan, 1994]. This analysis of global DSDP (and minor ODP) data from the top 50 m of oceanic crust confirmed the much earlier Christensen and Salisbury [1973] velocity results for the uppermost 1–2 m of basalts.

[65] Hydrothermal circulation may terminate beyond the Cretaceous, if at all, based on heterogeneity within some regional heat flow surveys on Cretaceous crust [e.g., Embley et al., 1983; Noel and Hounslow, 1988]. Average heat flow in these regions is conductive, but variations within each region suggest hydrothermal circulation. If global heat flow data are binned by crustal age, percentage standard deviation decreases systematically with age, even within the Mesozoic [Stein and Stein, 1994]. This observation indicates that waning hydrothermal circulation is still present in some old crust [Stein and Stein, 1994]. Although heat flow data cannot unambiguously determine whether this hydrothermal circulation in old crust is closed cell (no interchange with seawater) or slow open cell [Fisher, 1998], the two differ crucially in geochemical mass balance. For closed cell fluid flow, precipitation equals solution, and crustal geophysical properties may be unchanged by ongoing alteration.

[66] These conflicting indications of when off-axis hydrothermal circulation wanes have been partially reconciled. Apparently, any crustal alteration that may occur beyond ~5–10 Ma is so subtle that some geophysical methods may be incapable of resolving it. Fisher and Becker [2000] suggest that the rapid initial increase in velocity and decrease in permeability are not necessarily incompatible with heat flow evidence for persistence of open cell convection to ~65 Ma. They hypothesize that hydrothermal circulation continues to 65 Ma, in localized channels within the crust that are too small a percentage of total upper crust for significant effects on average crustal velocity. If the heat flow evidence for continued fluid flow could be coupled with a demonstration of the magnitude of ongoing crustal alteration, estimates of global geochemical fluxes into the oceans [Kadko et al., 1995; Elderfield and Schultz, 1996] and subduction zones could be refined. Sections 4.2–4.5 synthesize core, log, and seismic refraction evidence of whether physical properties of upper oceanic crust continue to evolve beyond ~5 Ma.

4.2. Crustal Velocity Versus Age

[67] Hoult and Ewing [1976] first demonstrated that sonobuoy refraction velocities for layer 2A increase systematically with increasing crustal age. This observation was attributed to lowering of uppermost crustal macroporosity by hydrothermal alteration products (Figure 8b). Subsequent refraction experiments using improved techniques demonstrate that seismic velocities of the upper oceanic crust increase rapidly for the first 5–8 Myr, from an average of 2.3 km/s at the spreading center to 4.3–4.4 km/s at 5–10 Ma [Grevemeyer and Weigel, 1996; Carlson, 1998]. Velocity modeling suggests that the large near-ridge velocity increases may be accomplished by changing the shape of pore spaces via secondary mineralization with relatively small overall porosity reduction [Wilkens et al., 1991; Shaw, 1994; Moos and Marion, 1994]. Beyond 10 Ma, no significant change in upper crustal velocity is evident.

[68] The uppermost layer of the seismic modeling studies, which is ~200–500 m thick, is the portion of oceanic crust that exhibits the greatest age-dependent change [White, 1984]. Thus the depth interval of the DSDP and ODP sites should provide a sensitive barometer of crustal aging for all crustal ages. Figure 9 plots average velocity versus logarithm of crustal age, based on averaging each of the 13 velocity logs in Figure 3. The velocity averages shown are actually based on averaging transit times (the inverse of velocity) rather than velocity because the seismic method similarly averages transit times rather than velocities. The 95% confidence limits shown are optimistic, because the assumption that each log value is independent is invalid in depth series. Intersite variation is high because of local variations in degree of alteration, initial porosity, and volcanic style.

[69] Average velocity for the flows ranges from 4.6 to 5.7 km/s, with no age dependence of velocity evident (Figure 9). We attribute this lack of age dependence to initial porosities and permeabilities too low for significant fluid flow and associated macroporosity reduction. Average velocity for the pillows appears to be 4.25–4.7 km/s initially, increasing with age to 4.8–5.0 km/s. Scatter at intermediate ages encompasses nearly the full range of observed velocities, precluding detection of any changes after 30 Ma. Overall average velocity, like pillow velocity, is initially 4.2–4.7 km/s and increases with age to 5.0–5.1 km/s. This change appears to continue beyond 30 Ma, but the evidence is not compelling. Because pillows are generally much more abundant than flows, the patterns for pillows and total basalts are mostly similar. Hole 504B (the youngest site), however, has many high-velocity flows and so is anomalously fast in total average velocity.

[70] The age dependence of total average velocity shown here (Figure 9) extends the already well-established age dependence of uppermost crustal velocities for <5–10 Myr based on seismic studies [Hoult and Ewing, 1976; White, 1984; Grevemeyer and Weigel, 1996, 1997; Carlson, 1998; Grevemeyer et al., 1999]. Individual seismic studies, though
subject to dip and other analytical uncertainties, are both more abundant and average larger basement volumes than the individual sites of this study. Nevertheless, the logs of this study augment the seismic results in two critical ways. First, this study is able to link velocity change to volcanic style and macroporosity reduction. Second, the logs appear to indicate systematic velocity increase with increasing age (top) for total basement rocks results from (middle) the age dependence for pillows. (bottom) In contrast, flows exhibit no age-dependent change.

Figure 9. Log-based velocity as a function of crustal age, for the 13 sites of Table 2, compared to the seismic velocity synthesis of Carlson [1998]. The moderate log-based trend toward velocity increase with increasing age (top) for total basement rocks results from (middle) the age dependence for pillows. (bottom) In contrast, flows exhibit no age-dependent change.

4.3. Macroporosity Versus Age

[71] Plugs reveal intergranular patterns, but they tend to miss cracks and crack-filling that potentially are major controls on large-scale velocity [Hyndman and Drury, 1976; Anderson and Zoback, 1983]. Estimation of crack porosity and its effect on large-scale physical properties using incomplete core recovery is difficult [Johnson, 1979b]. This crack filling may be more important to large-scale crustal velocity than the opposing intergranular effects of alteration. If much of basalt porosity is in the form of large-scale voids and fractures that are not sampled by 10-cm³ core plugs, then comparison of core and log measurements has the potential of detecting this macro-porosity. We use velocity logs rather than density logs for macroporosity detection because even the best density logs have abundant artifacts due to tool sensitivity to borehole irregularities (e.g., compare the Hole 801C density, velocity, and formation factor logs in Figure 2). Comparisons of core and log velocities are degraded, however, by low core recovery and nonrepresentative core recovery. Core recovery in basalts varies from ~10% to 80%, and the coring process can bias basalt core recovery toward fresher, less altered basalts. This reduced recovery leads to 0.5–9.0 m of depth uncertainty in attempts to match core and log measurements for the same depth.

[74] Many studies have compared basalt core and log velocities at individual sites. Such comparisons at several younger sites [Kirkpatrick, 1978; Moos et al., 1990; Jarrard and Broglio, 1991] concluded that macroporosity missed by the plugs caused lower velocity and density values for logs than for plugs. For old sites, in contrast, the standard model predicts that macroporosity is reduced or absent and therefore that plug and log velocities will be similar. This prediction is confirmed at Holes 765D [Shipboard Scientific Party, 1990a] and 801C (Figure 2), refuted at Hole 417D [Salisbury et al., 1979], and dominantly refuted at Hole
Figure 10. Apparent macroporosity (defined as the difference between plug-scale and log-scale porosities calculated from velocity measurements) as a function of crustal age for the 13 sites of Table 2. The pattern of macroporosity reduction with increasing age (top) for all basement rocks results from (middle) the age dependence of macroporosity values could be evaluated by comparing results of Schlumberger and multichannel sonic logs: calculated macroporosities differed by $+2.0\%$, $-0.6\%$, and $+0.1\%$, respectively.

[76] Flows indicate little or no systematic change of macroporosity with increasing crustal age. Macroporosity is so low (usually $<4\%$) for all crustal ages that only half of the site averages are significantly greater than zero at the 95% confidence level. Core descriptions routinely refer to flows as massive, with porosity generally confined to abundant cracks immediately adjacent to the flow boundaries. Undoubtedly this crack porosity does undergo some filling during alteration, but this surface change probably has minimal effect on average porosity of a flow.

[77] Pillow macroporosities are substantially higher than flow porosities, and they exhibit a decrease with age ($R = -0.89$) that is significant at the 99% confidence level, dropping from an initial value of $\sim 10\%$ to $0\%$–$4\%$ beyond 30 Ma (Figure 10). These data do not demonstrate con-

Figure 10 compares core plug and log velocities for the 13 sites of this study. Because a systematic difference between the two is thought to result primarily from the influence of macroporosity on logs but not on cores, we have expressed the difference between each core/log velocity pair as an apparent porosity difference $\Phi_L - \Phi_C = 100*(\Delta t_L - \Delta t_C)/(\Delta t_L - \Delta t_{ma})$, where $\Phi_L$ is percentage porosity calculated from log velocity, $\Phi_C$ is percentage porosity from core plug velocity, $\Delta t_L$ is measured log transit time (inverse of velocity), $\Delta t_C$ is measured core transit time, $\Delta t_f$ is fluid transit time (0.62 s/km), and $\Delta t_{ma}$ is matrix transit time. Matrix transit time is assumed to be 0.147 s/km, the value for fresh basalt. Alteration actually reduces $\Delta t_{ma}$ by up to 10% [e.g., Jarrard et al., 1995], but the resulting macroporosity error of $<3\%$ of calculated values is minor in comparison with the order of magnitude macroporosity variation among sites. Apparent macroporosity ($\Phi_L - \Phi_C$) is plotted as a function of age in Figure 10, for pillows, for flows, and for all basement rocks combined. For Holes 395A, 418A, and 504B, the robustness of macroporosity values could be evaluated by comparing results of Schlumberger and multichannel sonic logs: calculated macroporosities differed by $+2.0\%$, $-0.6\%$, and $+0.1\%$, respectively.

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vincingly that porosity reduction continues beyond 30 Ma, though the two oldest sites have the lowest pillow macroporosities. The pattern of total macroporosities is very similar to that from pillows, because pillows are the dominant component at most sites. Total macroporosities also exhibit a decrease with increasing age ($R = -0.85$) that is significant at the 99% confidence level, and this decrease appears to persist rather than wane at 30 Ma. Initial macroporosities of 8–10% eventually are reduced to near zero. This pattern is a strong confirmation of the standard model, which predicts age-dependent macroporosity reduction.

For the 13 sites of Table 2, core plug velocities exhibit an inverse correlation with logarithm of age that is significant at the 95% confidence level, both for total basalts ($R = -0.57$) and for pillows ($R = -0.68$). Core plug velocities of flows are not significantly correlated with logarithm of age ($R = -0.26$). Temporal increase in overall large-scale velocity due to fracture and void filling occurs despite alteration-induced decrease in core plug velocity. We conclude that the gradual convergence of core and log velocities (Figure 3) is not simply a macroporosity decrease (Figure 10) but a combination of fracture and void filling (increased log velocity) and decreased core plug velocity. Consequently, the age dependence of apparent macroporosity (Figure 10) is stronger than that for either log velocity (Figure 9) or core plug velocity, despite depth and volume mismatches inherent in comparing log and core plug velocities to calculate macroporosity.

4.4. Potassium Versus Age

Low-temperature crustal alteration processes include oxidation, hydration, and alkali fixation. The latter two generate clay minerals, the most abundant alteration products [e.g., Donnelly et al., 1979; Gillis and Robinson, 1988; Alt and Honnorez, 1984; Alt et al., 1986]. Consequently, altered basalts are generally higher in potassium than unaltered basalts [Hart, 1969; Hart et al., 1974]. The extent and timing of alteration-induced potassium enrichment of oceanic crust has implications for the potassium budgets of the ocean and subduction zones. Available data, however, have been generally sparse and nonrepresentative. Most major elements analyses of DSDP and ODP basalts select the least altered samples, because the geochemical objective is original composition rather than alteration-induced changes. Basalt core recovery is biased toward less altered materials, with underrepresentation of hyaloclastites and interpillow alteration minerals.

Spectral gamma ray logs, which provide continuous records of potassium concentration, offer additional perspectives on alteration-induced potassium enrichment. The potassium log at Hole 418A is highly correlated with qualitative ratings of core alteration [Broglia and Moos, 1988], whereas potassium logs for Sites 768 and 770 are weakly correlated with alteration indicators [Jarrard and Broglia, 1991]. We interpreted the strong inverse correlation between potassium and velocity logs within Hole 801C tholeites (Figure 7) as indicating production of high-K clays and celadonite primarily in the most porous and therefore most permeable zones, possibly augmented by alteration-induced increase in intergranular porosity and decrease in matrix velocity.

4.5. Core Alteration, Bulk Density, Porosity, and Matrix Density Versus Age

Basalt core descriptions routinely include subjective rankings of alteration intensity. Criteria differ from leg to leg, and no systematic pattern of increasing alteration rank with increasing age is evident. Progressive CO$_2$ enrichment of oceanic crust has been quantified by analysis of five DSDP and ODP sites [Staudigel et al., 1996; Alt and Teagle, 1999]. Alt and Teagle [1999] used these results to demon-
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![Figure 12. Potassium as a function of crustal age, for the seven sites of Table 2 that have spectral gamma ray logging. Age-dependent crustal alteration may cause progressive potassium enrichment, but the pattern is not statistically significant.](image)

strate that carbonate precipitation in veins continues beyond the 6 Ma age of Sites 504 and 896, resulting in enrichment of total CO₂ content of the upper extrusive section.

[88] Johnson and Semyan [1994] examined the possibility of age-dependent changes in velocity, porosity, and density, based on a compilation of core plug measurements from DSDP and ODP basalts. Far more drill holes into basalt had velocity (65 holes) or density (50 holes) data than porosity data (25 holes), and only two of these, including Site 801, were ODP sites. Johnson and Semyan [1994] found that average velocity and bulk density decreased systematically from 0 to 100 Ma, but data older than 100 Myr were inexplicably highly scattered and often higher than the trend for younger crust. Porosities exhibited a possible increase with age, but the change was not statistically significant. Note that these patterns of age-dependent change at the plug scale are the opposite of those seen at log scale (e.g., Figures 9 and 10).

[87] Johnson and Semyan [1994] also compiled averages of iron oxidation ratio and of H₂O⁺ from major element analyses. They found that oxidation occurs mainly within the first 5 Myr and exhibits no detectable increase beyond 20 Ma. Zhou et al. [2001] measured oxidation state of titanomagnetites in 97 MORB samples and demonstrated that small grains exhibit a strong linear pattern of increasing oxidation with logarithm of age. Their independent data confirm the findings of Johnson and Semyan [1994]: age-dependent oxidation is clearest for 0.1–5 Ma and is not resolved beyond ~10–30 Ma. The H₂O⁺ compilation of Johnson and Semyan [1994] confirms earlier indications [Hart, 1970, 1973; Donnelly et al., 1979; Muehlenbachs, 1979] that hydration increases with increasing age. Unfortunately, the timing of this change is ill-determined: H₂O⁺ correlates better with linear age (significant at 95% confidence level) than with logarithm of age (not significant), whereas oxidation correlates better with logarithm of age (significant at 99% confidence level) than with linear age (not significant). Both velocity and density decrease with increasing H₂O⁺ (both significant at 99% confidence level) [Johnson and Semyan, 1994], probably because of the combination of porosity-dependent alteration and alteration-induced decrease in matrix density and matrix velocity.

[88] Dating of alteration minerals, particularly celadonite, commonly yields ages 8–16 Myr younger than the age of the crust [Peterson et al., 1986; Gallahan and Duncan, 1994; Hart et al., 1994]. The comprehensive dating of celadonites within the Troodos ophiolite indicates that half of these alteration minerals formed more than 12 Myr after the crust, and alteration continued for at least 40 Myr [Gallahan and Duncan, 1994], particularly in the more permeable units. However, discrepancies between Rb/Sr and K/Ar ages and between core and outcrop ages suggest that these durations could be overestimates [Booij et al., 1995]. Furthermore, ophiolite alteration is nonrepresentative: it is generally more intense, with much higher fluid flux, than normal crustal alteration [Alt and Teagle, 2000].

[88] Our compilation of temporal changes in core plug physical properties draws on almost the same DSDP and ODP databases as those used by Johnson and Semyan [1994]; only five additional ODP basalt sites have obtained significant penetration of normal oceanic crust and associated index property measurements. Both compilations are confined to normal, open ocean oceanic crust. Johnson and Semyan [1994] also excluded sites with anomalous depths, whereas we exclude data from off-axis volcanism and sedimented ridges. They included sites with any basement penetration and data from only the top 50 m of basement. We only include sites with >40 m of basement penetration, in order to obtain a more representative crustal sample, and we average results from the top 300 m of basement. Their minimum number of measurements per site was two, but ours is five, again to obtain a more representative sample. They considered multiple holes at a site as independent samples and calculated an average for each, whereas we combine data from adjacent holes, except in rare cases where the holes were drilled into different hydrothermal environments (e.g., topographic high versus saddle). Both compilations are subject to similar biases: preferential core recovery and changing measurement techniques. For example, incomplete drying biases matrix densities and porosities downward [Jarrard and Broglia, 1991; Jarrard et al., 1995], whereas the routine technique of drying samples by heating to 100°–110°C biases them upward by driving off interlayer water from clays.
Figure 13. Bulk density, porosity, and matrix density as a function of crustal age based on core index property measurements for the sites of Table 3. The weak pattern of gradual density decrease results from systematic decrease in matrix density, not from porosity increase. Arrows associated with the youngest site indicate that its actual age is less than shown; its true age of 0.0 Ma cannot be plotted on a logarithmic scale nor be used in the regressions.

Unlike Johnson and Semyan [1994], we have not compiled core plug velocities, except for logged sites as described earlier. Our major focus is on matrix density, not compiled by them, because our analyses of Site 801 and other basalt sites led us to hypothesize that matrix density measurements may provide sensitive and abundant indicators of crustal hydration. Figure 13 shows variations of bulk density, matrix density, and porosity as a function of age, based on site averages of core plug analyses (Table 3).

No systematic change in porosity is evident, probably because of the competing effects of porosity filling by alteration minerals and porosity generation during alteration of minerals such as olivine. This observation at the plug scale does not imply lack of age-dependent decrease in macroporosity.

Bulk densities may decrease with increasing age ($R = 0.22$; not statistically significant) (Figure 13). The correlation is much higher ($R = 0.41$, significant at 95% confidence level) if low-porosity, low-density Site 410 is excluded. The density decrease appears to continue throughout the sampled age range of 0–167 Ma. This observation is consistent with the systematic decrease found by Johnson and Semyan [1994] for 0–100 Ma, but their observations of anomalously high scatter and high densities beyond 100 Ma are not confirmed by these data. The oldest three points on this plot are new data from ODP Leg 185: Hole 801C, Hole 1149B, and Hole 1149D. The pattern of bulk density decrease without systematic porosity change implies that the source of density decrease is matrix density.

Matrix density exhibits a strong pattern ($R = -0.57$; significant at 99% confidence level) of decrease with increasing age (Figure 13). The simplest interpretation of this pattern is that matrix density continues to decrease, at a rate proportional to logarithm of age, throughout the period 0–167 Ma. However, data scatter in the time period 10–70 Ma is high, and the hypothesis of no change beyond about 40–60 Ma cannot be rejected. Index property sampling at Site 597, which has an anomalously high average matrix density (Table 3), is described as concentrating on a massive flow [Shipboard Scientific Party, 1985c]; excluding this point raises the correlation coefficient to $-0.70$.

The matrix density plot of Figure 13 fails to remove a major source of variance: Intersite variations in basalt porosity affect permeability and therefore degree of alteration. Massive flows are usually much less altered than porous pillows, and several studies of individual sites have found a correlation between increasing porosity and decreasing matrix density [e.g., Christensen et al., 1980; Carlson and Herrick, 1990; Jarrard and Broglio, 1991; Busch et al., 1992; Jarrard et al., 1995]. Figure 14 plots site-average matrix densities for porosities of <5%, 5–10%, 10–15%, and >15%. Sites less than 10 Myr old exhibit little or no evidence of porosity-dependent reduction in matrix density, for these grouped data. The same may be true for 15–42 Ma sites, but near absence of porosities greater than 10% limits resolution in this time bracket. Older sites, 80–167 Ma, have matrix densities that clearly decrease with increasing porosity; among these are Leg 185 Holes 801C, 1149B, and 1149D.

To reduce the influence of porosity on the pattern of matrix density change due to crustal aging, we confine the averaging of matrix densities to samples with a porosity of >5% (Figure 15). Most massive flows, which are too impermeable to share in progressive alteration of oceanic crust, are thereby excluded. The pattern is quite similar to that for all porosities, but the dependence on logarithm of age is stronger ($R = -0.72$). The hypothesis of negligible crustal alteration beyond 10 Ma [Carlson, 1998; Grevermeyer et al., 1999] can be rejected at the 99% confidence level ($R = -0.73$ among sites >10 Ma).

4.6. Local Heterogeneity

Superimposed on the age-dependent trends in log-scale velocity (Figure 9), macroporosity (Figure 10), potassium (Figure 12), density (Figure 13), and matrix density (Figures 13 and 15) is substantial local heterogeneity. This heterogeneity, evident as scatter on these plots, is particularly evident at intermediate ages. Alteration appears to occur earlier at some sites than others, though eventually all are altered (Figures 9 and 13). Most of our 15–40 Ma points come from Leg 82 drilling in an area of the North Atlantic where both alteration and geophysical properties are clearly
Heterogeneous. ODP Leg 187, which obtained basement penetration of 7–67 m at 13 sites south of Australia, is not included in our figures because no logging or physical properties measurements were undertaken. Visual core descriptions of basalt alteration for these sites, ranging from 14 to 28 Ma, show high intersite heterogeneity and no correlation between alteration and crustal age [Shipboard Scientific Party, 2001]. The complexity of both vertical and horizontal variations in alteration intensity is best documented in ophiolites, particularly the Troodos ophiolite [Gillis and Robinson, 1988, 1990]. Basement topography often controls the geometry of hydrothermal circulation, concentrating outflow at topographic highs [Fisher et al., 1990, 1994]. This pattern is evident for paired holes on adjacent topographic highs and saddles, at 417A/417D/418A and at 504B/896A [Donnelly et al., 1979; Shipboard Scientific Party, 1993; Alt et al., 1996].

Heterogeneity of crustal alteration intensity hampers efforts to determine the pattern of waning hydrothermal circulation. Lack of logged sites prior to 6 Ma is not a serious disadvantage, because ample seismic experiments demonstrate the rapid velocity increase between 0 and about 5–10 Ma [Carlson, 1998; Grevemeyer et al., 1999]. Paucity of logged sites with ages between 15 and 100 Ma is particularly unfortunate: changes in velocity (Figure 9), macroporosity (Figure 10), and potassium (Figure 12) within this time interval are poorly determined, and only macroporosity suggests that upper crust of this age is systematically less altered than crust >100 Ma. Fortunately, this time interval is better represented by index property measurements. Matrix densities of high-porosity basalts provide the most comprehensive demonstration that crustal alteration continues, albeit at a decreasing rate, throughout the period 0–167 Ma, corresponding to virtually the entire range of extant oceanic crust.

4.7. Rejuvenation of Hydrothermal Circulation at the Outer Rise?

The geochemical signature of age-dependent crustal alteration can provide an essential starting point for calculations of geochemical mass balances associated with subduction. Implicit to such calculations, however, is the assumption that normal oceanic crust is not modified by the subduction process itself. This assumption may be invalid.

The flexure of lithosphere prior to subduction generates an outer rise. Extension at the top of the lithosphere causes normal faulting, horst-and-graben topography [Ludwig et al., 1966], and associated seismicity with tensional focal mechanisms [Stauder, 1968; Isacks et al., 1968; Christensen and Ruff, 1988]. These phenomena imply development and opening of near-vertical cracks. Within ocean basins, off-ridge open cell hydrothermal circulation appears to be limited largely by vertical permeability; horizontal permeability is still relatively high [Fisher and Becker, 2000]. Normal faulting at the outer rise might rejuvenate hydrothermal circulation by generating vertical permeability, particularly when combined with the strong

Figure 14. Relationships between porosity and matrix density based on core index property measurements for the sites of Table 3. High porosity permits increased fluid flow and resulting increased alteration, but this effect is only evident at older sites. See color version of this figure at back of this issue.

Figure 15. Matrix density as a function of crustal age, based on core index property measurements for the sites of Table 3. Because alteration is generally most intense in more porous basalts (Figure 14), site-average matrix density is calculated for samples with porosities of >5%.
topographic gradient from the crest of the outer rise to the trench floor. Strain cycling induced by cyclic seismic flexure may also contribute to fluid flow [Silver et al., 2000]. The very low heat flows measured immediately seaward of the Costa Rica and Peru trenches may be attributable to this mechanism [Langseth and Silver, 1996; Silver et al., 2000].

In addition to deepening Hole 801C, ODP Leg 185 drilled a deep crustal site on the outer rise of the Izu-Bonin Trench (Figure 1). Holes 1149B and 1149D, though widely spaced for holes from the same site, both encountered severely altered basalts. Indeed, the two lowest matrix densities on Figure 15 are these 132 Ma holes. Geochemical gradients in interstitial waters just above basement at Site 1149 demonstrate that fluids are currently circulating within this crust and, by geochemical mass balance, that this crust is still undergoing alteration [Shipboard Scientific Party, 2000b]. Historic seismicity on the Izu-Bonin outer rise is tensional [Christensen and Ruff, 1988; Kao and Chen, 1996]. Figure 16 shows that the crust surrounding Site 1149 is cut by abundant normal faults, as is typical for outer rises.

This correlation of crustal extension and faulting with extreme alteration and active hydrothermal flow may be coincidental. Alternatively, it may suggest that Site 1149 is representative of the final alteration state of subducting normal oceanic crust. If so, estimates of geochemical mass balance at subduction zones need to incorporate the effects of hydrothermal rejuvenation within oceanic crust immediately seaward of the trench axis. Matrix densities for Site 1149 imply ~25–30% more hydration than is predicted by the regression trend of Figure 15, but scatter from the trend is high. Unfortunately, no other crustal sites have been drilled on flexural outer rises. Additional heat flow surveys might resolve this uncertainty, by testing the generality of the pattern seen offshore Costa Rica by Langseth and Silver [1996].

5. Conclusions

ODP Hole 801C, which obtained the world’s oldest section of in situ, normal oceanic crust, provides the opportunity to examine relationships among hydrologic properties (porosity, permeability, fluid flow), crustal alteration, and geophysical properties, at both core plug and downhole log scales [Busch et al., 1992; Jarrard et al., 1995]. Within these upper crustal basalts, higher porosity is responsible for higher permeability and therefore higher fluid flow rates. High fluid flux, in turn, fosters alteration, particularly the hydration reactions that generate clays. Consequently, porosity is well correlated with both LAS-based hydration intensity (significant at 99% confidence level; Figure 5c) and matrix density (significant at 99% confidence level; Figure 5a).

Geophysical properties such as formation factor are impacted by this enhanced alteration in high-porosity zones, resulting in a correlation between core plug formation factor...
and matrix density (significant at 99% confidence level; Figure 5d). Porosity-dependent alteration is also seen at the log scale: the excellent character match between potassium and velocity logs (Figure 7) demonstrates that potassium enrichment is proportional to porosity. Core plug results indicate that hydration not only reduces density by reducing matrix density, but also reduces velocity by reducing matrix velocity. These combined effects are evident on velocity/density cross plots, for both core plug (Figure 6a) and log (Figure 6b) data.

[104] The similarity of core plug and log patterns at 801C is significant, because the core plugs indicate intergranular-scale properties, whereas the logs also respond to macroporosity. The log-based relationships of velocity to potassium (Figure 7) and of velocity to bulk density (Figure 6b) could result from either intergranular-scale alteration or hydration of cracks and interpillow voids. These two alteration scales cannot be isolated from the Hole 801C data alone; a global examination of temporal changes in hydrologic, alteration, and geophysical properties is needed.

[105] To examine the evolution of physical properties for normal oceanic crust, we used global data sets of DSDP and ODP core physical properties and downhole logs (Figure 1 and Tables 2 and 3). These core and log data suggest that crustal alteration continues, at a decreasing rate, throughout the lifetime of oceanic crust. Crustal aging is accompanied by increased hydration, increased log-scale velocity, and decreased macroporosity.

[106] Matrix densities provide the strongest demonstration of systematic increase in alteration versus logarithm of time (Figure 13). If variance associated with porosity-dependent alteration is reduced by confining the matrix density analysis to porosities greater than 5% (Figure 15), the ongoing evolution of matrix density is statistically significant not only for the complete age interval but also beyond 10 Ma (99% confidence level) and beyond 30 Ma (95% confidence level).

[107] Large-scale velocity of the upper oceanic crust increases with age. Seismic experiments demonstrate a systematic increase of average velocity from ~2.3 km/s at the spreading center to 4.3–4.4 km/s at 5–10 Ma [Grewe-meyer and Weigel, 1996; Carlson, 1998]. Beyond ~5–10 Ma, in contrast, seismic velocities detect no significant age dependence [Carlson, 1998], but DSDP and ODP log velocities for pillows indicate that crustal velocity increase continues (Figure 9; significant at 95% confidence level). Both velocity increases are attributed to decrease in macroporosity; the former may be more rapid because of velocity sensitivity to initial change in pore aspect ratio. This gradual increase in large-scale velocity occurs despite systematic velocity decrease at the intergranular scale, as indicated by core plug velocities for 0–100 Ma [Christensen and Salisbury, 1973; Johnson and Semyan, 1994] and for the 13 logged basement sites. Precipitation in cracks and interpillow voids must be dominant over intergranular-scale alteration in controlling the evolution of large-scale crustal velocity.

[108] Macroporosity can be estimated based on the difference between log and core plug velocity at the same depth. Macroporosity decreases with increasing crustal age, with a logarithm of age dependence that is evident in overall site averages (significant at 99% confidence level; Figure 10) but even better demonstrated among the pillow basalt (significant at 99% confidence level; Figure 10). This systematic macroporosity decrease contrasts with the absence of significant age-dependent porosity change among core plugs, based on the compilations of Johnson and Semyan [1994] and Figure 13.

[109] The time-dependent alteration of upper oceanic crust, as well as associated changes in geophysical properties, can be locally overwhelmed by two sources of variance: local topography, which controls hydrothermal circulation intensity, and initial permeability, which is sensitive to porosity and therefore also to volcanic style. Flows are generally much more massive and impermeable than pillows, so they are much less subject to alteration. Consequently, systematic age-dependent changes in log-scale velocity, core plug velocity, and macroporosity are much stronger for pillows (significant at 95%, 95%, and 99% confidence level, respectively) than for flows (none is significant). The observation of intense intergranular alteration of pillows and flow margins, despite intergranular permeabilities that are more than 4 orders of magnitude lower than the fracture and interpillow permeabilities measured by packers, requires that fluid flow and attendant alteration can proceed even at these exceedingly low permeabilities. The lower limit for fluid flow and associated alteration appears to be ~10⁻¹⁹ m², based on the observation that lower intergranular permeabilities are encountered mostly in fresh, massive basalts with porosities of <5% [Karato, 1983a, 1983b; Johnson, 1979a; Hamano, 1979]. Supporting evidence for this permeability threshold comes from the lack of matrix density reduction for porosities of <5% (Figure 13) and the lack of age dependence for both velocity and macroporosity of massive flows (Figures 9 and 10).

[110] Examples of ongoing hydrothermal circulation in very old crust, such as the present-day flow within basement at Site 1149 [Shipboard Scientific Party, 2000b], microbial alteration patterns within Hole 801C [Shipboard Scientific Party, 2000a], and local heat flow anomalies in several regions [e.g., Embley et al., 1983; Noel and Hounslo, 1988], do not appear to be confined to isolated instances. Persistence of detectable hydrothermal circulation throughout the lifetime of oceanic crust is suggested by age-dependent change in heat flow variance [Stein and Stein, 1994], velocity logs, macroporosity data, and matrix densities. This conclusion is not incompatible with previous results suggesting lack of detectable change in seismic velocity beyond 5–10 Ma [Carlson, 1998], preponderance of alteration mineral ages < 40 Ma [e.g., Gallahan and Duncan, 1994], and convergence of theoretical and observed heat flows at 65 Ma [Stein and Stein, 1994]. The majority of velocity change and alteration is relatively young, and termination of the heat flow deficit signifies a reduction of advection to a level minor in comparison to conduction, rather than an ending of hydrothermal flow [Anderson and Skilbeck, 1981; Jacobson, 1992; Stein and Stein, 1994]. On the basis of observed decreases in matrix density that are proportional to the logarithm of age (Figures 13 and 15), approximately half of all intergranular-scale crustal alteration occurs after the first 10–15 Myr.

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References


Figure 7. Comparison of velocity and potassium logs for Hole 801C tholeiites. The strong inverse correlation between these logs indicates that precipitation of potassium-bearing clays and celadonite occurs mainly in the most porous (lowest-velocity) zones.

Figure 14. Relationships between porosity and matrix density based on core index property measurements for the sites of Table 3. High porosity permits increased fluid flow and resulting increased alteration, but this effect is only evident at older sites.