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Seismic Properties, Density, and Composition of the Icelandic Crust Near Reydarfjördur

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Compressional and shear wave velocities, densities, and porosities have been measured to confining pressures of 6 kbar for 135 samples from a deep borehole in Eastern Iceland. The measurements are from a random sampling of dikes and sills through the 2 km hole. Velocities within the flow units are relatively low at the tops, reach maximum values within central portions, and decrease near the bases. This systematic variation correlates well with variations in porosity and bulk density. The dike units, on the other hand, have relatively uniform velocities, which are approximately equivalent to maximum velocities measured within flow units. Below 700 m the dike and flow velocities show systematic increases with depth which are related to changes in mineralogy and independent of porosity. The velocity gradient $(dVp/dZ)_T$ within this region is well defined by the laboratory velocities and approximates 0.6 s^{-1}, in good agreement with field refraction measurements. High velocities near the base of the hole are related to increasing alteration with depth and, in particular, an increase in epidote content. The abundance of epidote in the rocks recovered from the lower portion of the drillhole supports a metamorphic origin for the layer 2–layer 3 boundary in this region.

INTRODUCTION

Information on the structure of the Icelandic crust has been derived principally from seismic refraction studies carried out over the past two decades. The first seismic profiles reported by *Båth* [1960] produced a model consisting of a three-layered crust with a total crustal thickness of approximately 28 km. Compressional wave velocities for the layers based on two profiles across Iceland were 3.69, 6.71, and 7.38 km/s with layer thicknesses of 2.1, 15.7, and 10.0 km, respectively. A somewhat different average crustal structure was presented by *Pálmason* [1971] based on a total of over 80 refraction profiles distributed throughout Iceland. Pálmason recognized five layers, termed 0 through 4. The shallowest seismic layer, layer 0, found only in the neovolcanic zone, has thicknesses between 0 and 1.0 km and an average compressional wave velocity of 2.75 km/s. Layers 1 and 2 have average velocities of 4.14 and 5.08 km/s and average thicknesses of 1.04 and 2.15 km, respectively. The average velocity of layer 3 is approximately 6.50 km/s and thicknesses are in the range of 4–7 km, except in northern Iceland where they reach 12–13 km. Many of Pálmason's profiles were not of sufficient length to give information on layer 4; however, the average velocity of this layer based on five profiles is 7.19 km/s.

Recently, *Flovénz* [1980] reinterpreted Pálmason's data using synthetic seismograms. He concluded that the Icelandic crust consists of two distinct regions: an upper section in which velocity increases continuously with depth from approximately 2.0 to 6.5 km/s (layers 1 and 2) and a lower region with nearly constant velocity (layer 3). Within the upper section, the gradient is approximately 0.57 s^{-1}. Compressional wave velocities in the lower crust are 6.5 km/s, in good agreement with *Pálmason*'s [1971] model. In regions in which there has been little erosion, the depth to the 6.5 km/s layer is typically 5–6 km.

The seismic structure of Iceland is anomalous in many ways. Total crustal thickness appears to be much greater than that of normal oceanic regions and possibly thinner than the crust underlying the continents. Typical upper mantle velocities in the range of 7.8–8.2 km/s have not been reported under Iceland. Furthermore, it is unclear whether the 7.38 km/s layer of *Båth* [1960] and the 7.19 km/s layer (layer 4) of *Pálmason*'s [1971] represent anomalous mantle material possibly affected by high temperature, partial melting, and/or alteration, or correlate with high velocity basal oceanic crustal layers observed by *Sutton et al.* [1971].

The nature of Pálmason's layer 3, with a velocity of approximately 6.5 km/s, is also highly uncertain. Pálmason [1971] concluded that the layer 2–3 boundary is a metamorphic boundary formed at the transition of greenschist facies mineral assemblages to amphibolite facies assemblages and fossilized upon cooling of the crust. Evidence supporting this interpretation includes a correlation of depth to layer 3 with temperature as deduced from borehole data and the observation that the layer 2–3 boundary is essentially horizontal in regions where the basalt flows are dipping at relatively high angles. *Walker* [1975], on the other hand, has postulated that Icelandic layer 3 consists of a swarm of basic intrusive sheets similar to those exposed in southeastern Iceland. According to Walker's interpretation, the intrusive sheets penetrate higher into the crust in regions with relatively low density rocks such as hyaloclastites or acid volcanics, thus explaining the relatively shallow depths to Icelandic layer 3 in southeastern Iceland and near volcanic centers.

During the summer of 1978 a drilling operation was launched in eastern Iceland near Reydarfjördur, with the major objective of penetrating Icelandic layer 3. One of Pálmason's refraction lines, located approximately 12 km west of the drill site, gave depths to layer 3 of only 1.5 km. The drilling penetrated 1.9 km into the Icelandic crust, supposedly within the reach of layer 3. However, a seismic refraction experiment conducted near the drill site during the

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drilling operation indicated that the layer 2–layer 3 transition occurs at a depth of 3.0–4.5 km beneath the drill site [*MacKenzie et al.*, this issue; *Thomson et al.*, this issue]. This suggests a substantial eastward dip of the layer 2–layer 3 transition in this region.

In this paper we present the densities, porosities, ultrasonic velocities, and elastic constants of rocks recovered from the drillhole. The relationships between these parameters are discussed in detail. The precise depth locations of the samples allow velocity-depth profiles to be constructed from velocities measured in the laboratory and compared with the seismic refraction profiles. Finally the seismic properties obtained from several of the rocks from the drillhole provide information as to the possible nature and depth of Icelandic layer 3 underlying this region.

SAMPLE SELECTION AND EXPERIMENTAL DETAILS

Ninety samples were taken from flow basalts and forty-five from intrusive dike units for physical properties determinations, the locations of which are illustrated in Figure 1. Visual inspection of the flow units revealed large variations in porosity between flow margins and flow centers. Due to rapid cooling, gases have been trapped on flow bases and tops producing high porosity vesicular margins which make flows readily identifiable. Since previous investigations of oceanic basalt physical properties have shown strong porosity dependent relationships [e.g., *Hyndman*, 1979], care was taken to sample at different levels within the flows so that both high and low porosity samples were represented at all levels in the core. Individual dike units exhibited a much greater degree of homogeneity. This has been substantiated by our laboratory measurements which show no significant correlation between physical properties and position within the dike units.

Samples were cut to right circular cylinders, 2.5 cm in diameter and 3–6 cm in length, air-dried, measured, and weighed to allow determination of dry bulk densities. Grain densities were calculated from the porosities and bulk densities. Wet bulk densities were obtained after saturation of the samples, and effective porosities were calculated from the dry and wet bulk densities. Compressional wave (V_p) and shear wave (V_s) velocities were measured in saturated samples using the pulse transmission technique of *Birch* [1960] and 2 MHz transducers. The samples were jacketed in copper foil with 100 mesh screen between the jacket and the rock. The screen insured that pore pressure in the sample would stay well below the applied confining pressure. Velocities in samples with porosities less than 5% were run to confining pressures of 6.0 kbar, while those with porosities greater than 5% were run to 2.0 kbar in order to avoid crushing the samples. The accuracy of the velocity measuring technique is estimated at 1% for shear wave velocities and 0.5% for compressional wave velocities [*Christensen and Shaw*, 1970].

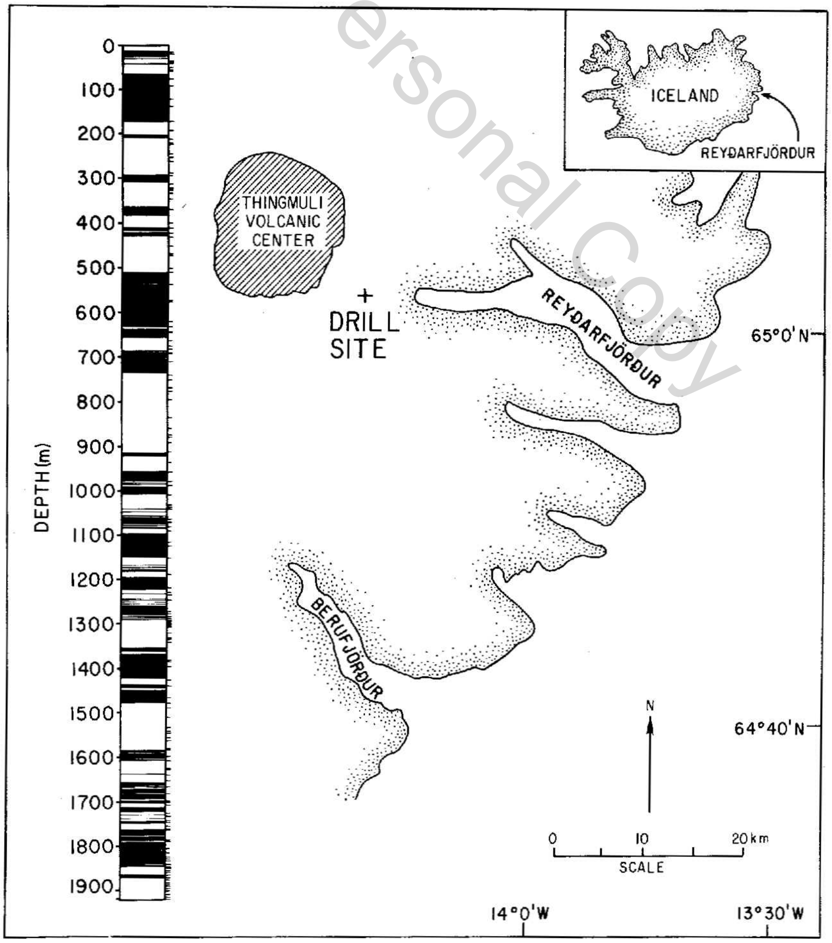

Fig. 1. Location of the drill site and stratigraphic column of the drill core. Locations of physical properties samples in the Iceland drill core are shown as tickmarks. Dikes are shown as solid regions.

TABLE 1. Compressional (V_p) and Shear (V_y) Wave Velocities, Poisson's Ratios (σ), Bulk Densities (ρ_b), Grain Densities (ρ_g), and Porosities (ϕ)

				0.4 kbar			1.0 kbar			2.0 kbar			6.0 kbar		
	ρ_b, g cm^{-3}	ρ_g, g cm^{-3}	ϕ, %	V_p, km s^{-1}	V_s, km s^{-1}	σ	V_p, km s^{-1}	V_s, km s^{-1}	σ	V_p, km s^{-1}	V_s, km s^{-1}	σ	V_p, km s^{-1}	V_s, km s^{-1}	σ
							All Flows (90 Samples)								
Max.	3.07	3.09	23.1	6.53	3.71	0.36	6.66	3.80	0.34	6.74	3.87	0.33			
Min.	2.36	2.69	0.1	3.46	1.70	0.19	3.58	1.84	0.19	3.67	1.94	0.21			
Mean	2.85	2.92	4.0	5.67	3.09	0.29	5.74	3.15	0.29	5.84	3.19	0.29			
sd	±0.16	±0.09	±5.1	±0.65	±0.37	±0.02	±0.63	±0.36	±0.02	±0.58	±0.32	±0.02			
							Flow Tops (10 Samples)								
Max.	2.74	2.96	23.1	5.12	2.92	0.36	5.17	2.99	0.34	5.26	3.00	0.31			
Min.	2.37	2.73	10.6	3.46	1.70	0.19	3.58	1.84	0.19	3.67	1.94	0.21			
Mean	2.53	2.84	16.7	4.22	2.26	0.30	4.37	2.39	0.29	4.56	2.52	0.28			
sd	±0.14	0.08	±4.8	±0.55	±0.42	±0.05	±0.52	±0.38	±0.04	±0.51	±0.34	±0.03			
							Flow Centers (49 Samples)								
Max.	3.07	3.09	1.9	6.53	3.50	0.34	6.66	3.58	0.33	6.74	3.67	0.33	6.36	3.73	0.33
Min.	2.67	2.69	0.1	5.56	3.01	0.26	5.62	3.03	0.26	5.69	3.05	0.26	5.80	3.07	0.26
Mean	2.93	2.95	1.1	6.05	3.28	0.30	6.11	3.33	0.29	6.18	3.36	0.29	6.32	3.42	0.29
sd	±0.07	±0.07	±0.4	±0.22	±0.12	±0.01	±0.23	±0.13	±0.01	±0.25	±0.14	±0.01	±0.25	±0.15	±0.01
							All Dikes (45 Samples)								
Max.	3.04	3.05	3.3	6.66	3.52	0.32	6.71	3.56	0.32	6.75	3.58	0.32	6.86	3.63	0.33
Min.	2.83	2.85	0.3	5.54	2.93	0.25	5.63	2.98	0.25	5.70	3.02	0.26	5.90	3.07	0.27
Mean	2.95	2.98	1.3	6.02	3.24	0.30	6.10	3.29	0.29	6.17	3.33	0.29	6.32	3.38	0.30
sd	±0.07	±0.06	±0.8	±0.26	±0.15	±0.01	±0.26	±0.15	±0.01	±0.25	±0.14	±0.01	±0.24	±0.13	±0.01

Petrographic and visual examinations of the cores show that the higher porosities originate primarily from vesicles. Since the vesicles do not close significantly with applied pressure, velocities show much less pressure dependence than velocities in rocks with abundant microcracks. The average change of velocity for Iceland cores over the pressure interval from 0.5 to 1.0 kbar is less than 0.1 km/s. Thus, we have selected 1 kbar velocities for our correlations of velocity with density and porosity.

The actual pore structures of the basalts are likely to be quite complicated. Alteration, especially along the flow unit margins, has resulted in the deposition of secondary minerals along vesicle walls, which has likely sealed some vesicles from crack porosity. Because of this, the effective porosities reported in Table 1 are likely to be smaller than the actual rock porosities.

DATA

Table 1 presents maximum, minimum, and average values of compressional (V_p) and shear wave velocities (V_s), Poisson's ratios (σ), effective porosities (ϕ), wet bulk densities (ρ_b), and grain densities (ρ_g) for 135 samples. In order to apply these experimental results to the seismic structure of the Icelandic crust, it is necessary to understand the factors which are responsible for the variability in properties measured over the 2 km depth interval and summarized in Table 1. The results are of general importance and bear directly not only upon the nature of the Icelandic crust, but also upon the physical properties of the ocean crust and continental regions capped by plateau basalts.

The greater ranges of values measured for the flow samples compared to the dikes reflect the nonuniform nature of the flows, which are highly vesicular near their margins and massive in their centers. The margins of the flow units are characterized by high porosities, low bulk densities, highly variable Poisson's ratios, and low velocities. In Table 1 we have presented as flow top material all flow samples with porosity greater than 10% and as flow centers those samples of porosity less than 2%. These are arbitrary limits, chosen to accentuate the variability within the flow units. The physical properties of the dikes, on the other hand, show much less variability both within an individual unit and throughout the column. In fact, the average velocities, densities, and Poisson's ratios of the dike rocks are quite similar to the properties of samples from the flow interiors. This is completely concordant with our observations that the flow interiors and dike units both have low porosity and are similar in mineralogy.

In Figure 2, porosity is plotted against wet bulk density.

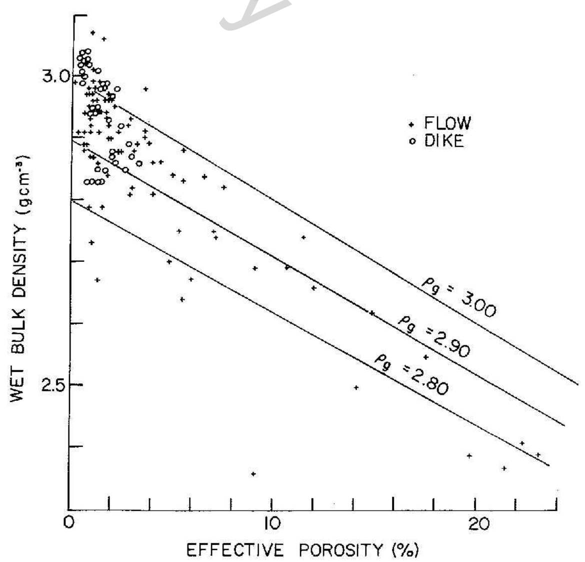

Fig. 2. Wet bulk density versus effective porosity. Lines are calculated values of bulk density-porosity relationships for assumed grain densities and water saturated samples.

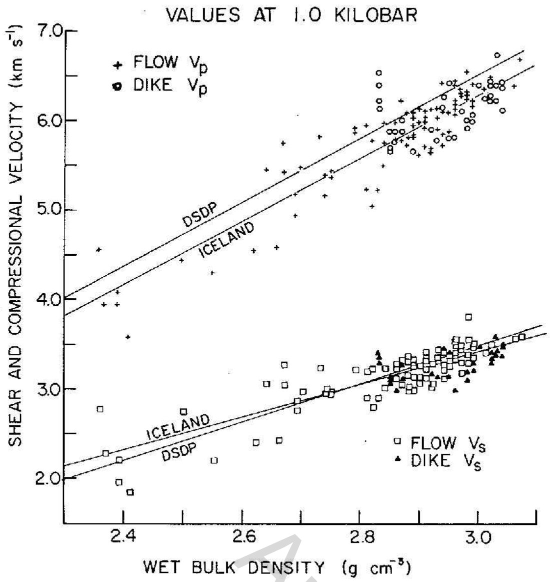

Fig. 3. Shear and compressional velocity versus wet bulk density. Linear fits are those of this study (Iceland) and *Christensen and Salisbury* [1975] (DSDP). See Table 2 for regression equations.

Superimposed on this plot are curves showing the relationship between porosity and bulk density for grain densities 2.8, 2.9, and 3.0 g cm^{-3}. The porosity samples, collected from flow margins, show relatively low grain densities, which is consistent with petrographic observations of relatively high alteration in these samples. This alteration is presumably related to high permeability associated with high porosity. Secondary mineralization in the vesicular zones also has a significant influence on Poisson's ratio. For example, the minimum measured value of Poisson's ratio (0.19) is in a vesicular sample that contains abundant secondary quartz. Compared with most rock-forming minerals, quartz has an extremely low Poisson's ratio [e.g., *Birch*, 1961 (see Table 3)], and thus, rocks containing abundant quartz are characterized by low Poisson's ratios.

VELOCITY-DENSITY-POROSITY RELATIONS

Previous studies of physical properties of basalts obtained from deep sea drilling have shown a consistent relationship between both shear and compressional velocity and wet bulk density [e.g., *Christensen and Salisbury*, 1975]. Figure 3 presents the velocity-density relationships for the Iceland samples. Also presented are the *Christensen and Salisbury* [1975] linear solutions for shear and compressional velocity variation with density for 77 samples of DSDP basalts. Table 2 presents linear regression parameters and correlation coefficients (r) for Iceland and DSDP basalts. An important conclusion from this table is that the bulk of the suite of

TABLE 2. Regression Line Parameters (Values at 1.0 kbar)

	ρ	r
	Iceland	
$V_p = -4.18 + 3.48\rho$	$1.53 + 0.230V_p$	0.90
$V_s = -1.97 + 1.79\rho$	$1.63 + 0.394V_s$	0.84
	DSDP	
$V_p = -4.10 + 3.52\rho$	$1.27 + 0.265V_p$	0.97
$V_s = -2.79 + 2.08\rho$	$1.49 + 0.428V_s$	0.94

Iceland samples exhibits significantly slower compressional velocities for a given density than DSDP basalts. *Birch* [1961] established the relationship between compressional wave velocity, density, and mean atomic weight, defined as the average atomic weight of a rock's chemical constituents. For a given density, velocity decreases with increasing mean atomic weight, while for a constant mean atomic weight, velocity increases with density. An increase in iron content, usually at the expense of magnesium in a rock, raises both the mean atomic weight and the density. The net effect can decrease velocity. Thus for a given density, iron rich samples will have slower seismic velocities. Chemical analyses of the Iceland drill core samples (G. Pritchard, personal communication, 1980) have low Mg/Fe ratios compared to oceanic basalts collected away from Iceland on the Mid-Atlantic Ridge [e.g., *Blanchard et al.*, 1976] Mg/(Mg + Fe) for 100 Iceland samples has a mean value of 0.384 ± 0.045, while the oceanic basalts range from a minimum of 0.51 to a maximum of 0.83. The effect of iron, producing different velocity-density relations for the two rock suites, is significant. At equivalent confining pressures, rocks from the Iceland drill core which have compressional wave velocities similar to oceanic tholeiites differ in density on the average of 0.06 g/cm^3. In the 2.80–3.00 g/cm^3 density range, where most of the data are concentrated (Figure 3), Poisson's ratio, calculated from the least squares solutions of Table 2, appears to be slightly higher for DSDP basalts (0.30) than the Iceland samples (0.29). This is probably the result of the presence of quartz in the Iceland samples, a mineral not often seen in abundance in true oceanic basalts.

The relationship between porosity and velocity (Figure 4) is a reflection of the density-velocity curve. Since porosity is the controlling factor in the bulk density, it is not surprising that porosity and velocity show a systematic relationship. Furthermore, much of the porosity is in the form of vesicles, cavities which are unlikely to close at elevated confining pressures [*Hyndman*, 1979]. Thus measurements of porosity carried out at 1.0 bar can be correlated with velocities

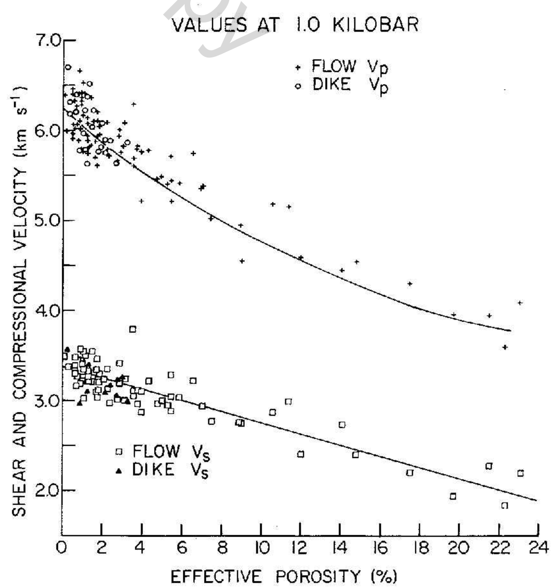

Fig. 4. Shear and compressional velocity versus effective porosity. Curve on compressional data is *Wyllie* [1958] time average, shear data fit with least squares. See text for equations.

measured at elevated pressures without seriously affecting the relationship. The curve superimposed on the compressional data is a calculated velocity-porosity relationship using *Wyllie*'s [1958] time average equation:

$$\frac{1}{V_m} = \frac{\phi}{V_f} + \frac{1-\phi}{V_r}$$

where V_m is measured velocity, V_f is the velocity of the pore fluid (1.5 km/s in this case), V_r is the rock velocity, and ϕ is the porosity expressed as a fraction. The fit is quite good using a rock velocity of 6.25 km/s, which is probably slightly slow for the densest basalts and slightly high for the most porous, low grain density samples. Shear velocity versus porosity can be adequately described with a straight line, the equation for which is $V_s = 3.38 - 6.15\phi$, where ϕ again is expressed as a fraction.

VARIATIONS WITH DEPTH

Density-Depth

Measured wet bulk densities plotted against depth are shown in Figure 5. It should be re-emphasized that bulk densities within individual flow units are highly variable, even at substantial subsurface depths. While hydrostatic pressure may be important in closing crack porosity under in situ conditions, the results of the Iceland drilling show that vesicular porosity can persist to substantial depths. The fact that the scatter is somewhat less at depth can be attributed to a higher degree of pore filling in vesicular zones.

Another significant feature of the density-depth plot is the general increase in density with depth. The flows have a density minimum between 800 and 1300 m depth which may be the result of a zone of high permeability and attendant alteration. At a depth of 600 m, approximately where the flow density values begin to fall off, the top of a major hot

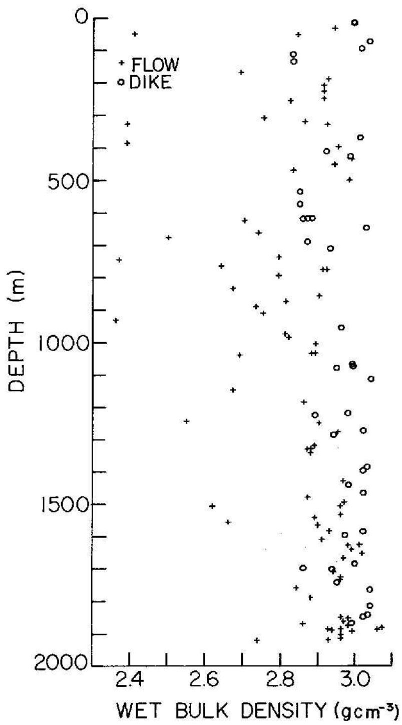

Fig. 5. Variation of wet bulk density with depth.

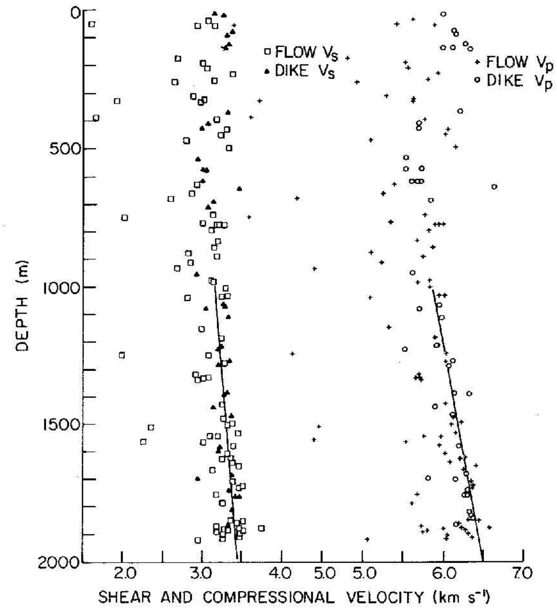

Fig. 6. Variation of shear and compressional velocity with depth. Straight line between 1000 and 2000 m represents layer 2 velocity gradient of *Flovénz* [1980].

water aquifer was penetrated. In addition, geochemical studies [*Mehegan et al.*, this issue] have shown that flow units within this depth interval are relatively evolved geochemically. The increase in dike bulk density is not as apparent as that in the flows, but less than 25% of the dike samples collected above 1000 m had density in excess of 3.0 g cm^{-3}, while 50% of those from below 1000 m exceeded this density.

It should also be noted that almost all of the samples with bulk densities greater than 3.00 g cm^{-3} were collected below 1 km, most of these from below 1500 m. The two highest density samples collected (3.07 g cm^{-3}) were from flow interiors very near the bottom of the hole.

In Figure 6, velocities measured at in situ effective pressures, calculated as lithostatic minus hydrostatic pressure, are plotted versus depth. An important feature of this figure is the increase in the maximum flow velocities with depth below the 900 m density minimum. None of the flow samples above 1500 m have a compressional velocity above 6.25 km/s, while there are 12 which exceed this velocity at depths greater than 1500 m. Superimposed on the V_p plot is a line representing the velocity gradient of *Flovénz* [1980] for layer 2 of the Icelandic crust. Using a V_p/V_s ratio of 1.85 derived from the average values in Table 1, a V_s gradient is also plotted. In order to isolate the effect of alteration on the rock matrix velocity, a velocity-depth regression line was calculated for all samples with porosity less than 1%. The result was a gradient of 0.45 s^{-1} (N = 34, r = 0.73). For the flows only, the gradient is 0.40 s^{-1} (N = 16, r = 0.70), and for dikes, 0.46 s^{-1} (N = 18, r = 0.74). While these gradients are somewhat less than that of Flovénz, they demonstrate that much of the velocity increase in layer 2 of the Icelandic crust can be accounted for without calling on large changes in crustal porosity. Indeed, to the extent that the velocity-porosity relationship of Figure 2 can be extrapolated to megascopic scale, porosity needs to only decrease by 1% per kilometer to make up the difference between the 0.65^{-1} gradient and that found in the rock samples of the drillhole.

TABLE 3. Density and Seismic Properties of Some Mineral Aggregates

Mineral	ρ_b, g cm^{-3}	V_p, km s^{-1}	V_s, km s^{-1}	V_p/V_s	σ
Quartz* Clarendon Springs, Vt.	2.63	6.04	3.94	1.54	0.13
Calcite Darrington, Wash.	2.70	6.81	3.59	1.90	0.31
Chlorite Ishpeming, Mich.	3.16	6.01	2.95	2.04	0.34
Epidote Rockbridge, Va.	3.33	6.96	3.82	1.82	0.28

* *Christensen* [1965, 1966].

The Petrologic Nature of Icelandic Layer 3

Although the drillhole at the Reydarfjördur site did not reach Icelandic layer 3, the seismic properties of the rocks from the drillhole provide important information concerning the composition of the lower Icelandic crust. Of major significance, the compressional wave velocities of the dike rocks are not of sufficient magnitude to account for the transition from layer 2 to layer 3. Our average velocity for 45 dike samples at 1.0 kbar confining pressure is 6.10 km/s (Table 1), well below the average layer 3 velocity of 6.50 km/s given by *Pálmason* [1971] and the 6.7 km/s velocity for layer 3 in the vicinity of the drill site estimated by *MacKenzie et al.* [this issue] and *Thomson et al.* [this issue]. Thus an intrusive sheet complex consisting of dikes similar to those exposed in southeastern Iceland will possess a maximum compressional wave velocity of approximately 6.1 km/s at confining pressures equivalent to those of Icelandic layer 3. In reality the actual velocity of the dike complex would be less, when the effects of large scale fracturing and elevated temperature, both of which lower velocity, are taken into account.

Evidence for the composition of Icelandic layer 3 and the nature of the layer 2–layer 3 transition is given by the mineralogy and seismic velocities of samples from the drillhole, and our observation that bulk density increases significantly in the lower 1 km of the section (Figure 5). An examination of thin sections cut from the samples in which the velocities were measured shows that abundant secondary minerals occur in significant proportions near the base of the section, which include chlorite/smectite, epidote, quartz, and calcite. These minerals occur as part of the basalt groundmass, as vein fillings and in vesicles. Densities and velocities of these minerals are given in Table 3. The data represent average values of three cores cut in mutually perpendicular directions from single samples of the mineral aggregates. The densities are bulk rock densities determined from the weights and dimensions of the cores, and the velocities and Poisson's ratios are for 1 kbar confining pressure. Since epidote and chlorite properties will vary with composition, densities and velocities for these minerals in Table 3 are only approximate. Nevertheless, there are significant differences in the densities, velocities, and Poisson's ratios for these minerals. Epidote and chlorite have relatively high densities, whereas calcite and epidote have high compressional wave velocities, quartz and epidote have high shear wave velocities, and both calcite and chlorite have Poisson's ratios above 0.30.

It is clear from Table 3 that increasing alteration with depth of Icelandic flows and dikes, which is accompanied by an increase in epidote content, could produce the observed compressional wave refraction velocities for Icelandic layer 3. An additional clue as to the petrologic nature of crustal layer 3 comes from determinations of Poisson's ratio calculated from compressional and shear velocities [*Christensen and Salisbury*, 1975]. Thus the Icelandic crustal data of *Pálmason* [1971] and *Flovénz* [1980] are particularly useful because shear velocities as well as compressional velocities are reported. *Pálmason*'s [1971] and *Flóvenz*' [1980] values of Poisson's ratio for Icelandic layer 3 are 0.27 and 0.28, respectively, which is also consistent with epidotization of the Iceland basaltic rocks in layer 3. Our highest velocity rocks, occurring near the base of the drillhole, contain significant amounts of epidote (up to 20%), and the observed increase in bulk density in the lower kilometer of the drillhole is at least, in part, related to increasing amounts of epidote with depth. The actual percentage of epidote required to produce layer 3 velocities is estimated at approximately 25% assuming an average velocity of 6.25 km/s at 1 kbar for an epidote-free rock. An increasing percentage of amphibole within the upper portion of layer 3 could also account for the observed velocities and Poisson's ratios [*Christensen*, 1974].

Velocities in chlorite and quartz are, on the other hand, sufficiently low that they cannot be present in large amounts in Icelandic layer 3. Similar arguments can be made on the basis of the anomalous Poisson's ratios for these minerals (low for quartz and high for chlorite). Calcite, if present in appreciable amounts, will increase the compressional wave velocity of Icelandic basalt and concomitantly increase Poisson's ratio to a value significantly higher than recorded by Pálmason and Flovénz.

In summary, the inference that the lower Icelandic crust is composed of a sheeted dike complex similar in mineralogy to dikes occurring in southeastern Iceland is not supported. The experimentally determined velocities for Icelandic rocks are in agreement with previous interpretations by *Pálmason* [1971] and *Flovénz* [1980] that the layer 2–layer 3 seismic boundary is metamorphic. The precise nature of the metamorphic transition is uncertain. It may mark the greenschist-amphibolite facies boundary [*Pálmason*, 1971], or it may simply be a consequence of an increase in epidote as suggested by our measurements.

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