GEOPHYSICAL RESEARCH LETTERS, VOL. 22, NO. 10, PAGES 1309-1312, MAY 15, 1995

Three-dimensional v_P and v_P/v_S structure of the Hengill Triple Junction and geothermal area, Iceland, and the repeatability of tomographic inversion

G. R. Foulger and A. D. Miller

Dept. Geological Sciences, University of Durham, U.K.

B. R. Julian and J. R. Evans

Branch of Seismology, U.S. Geological Survey, Menlo Park, California

Abstract. We investigate the crustal structure of the Hengill triple junction in southwestern Iceland, applying tomographic methods to local earthquake data recorded in two field experiments with different network geometries and instrumentation. Data from the two experiments enable us to derive three-dimensional models of the compressional-wave speed v_p and the wave-speed ratio v_p/v_s. Well resolved high-v_p bodies correlate with sites of gabbroic intrusions. A small reduction in v_p/v_s associated with the high-temperature part of the geothermal area is probably due to mineral alteration or supercritical fluids. The RMS difference between the two v_p models, about 0.26 km s^{-1}, indicates the approximate repeatability that may be expected of good tomographic inversions.

Introduction

The Hengill ridge-ridge-transform triple junction area (Figure 1) contains widespread geothermal resources that are associated with continuous, small magnitude earthquake activity. This area is ideal for studying three-dimensional structure using local earthquake tomography, as earthquake activity has a broad, predictable spatial distribution, and illuminates those volumes where the strongest structural heterogeneity is expected. Two such studies of the area have been conducted, using data collected in 1981 [*Toomey and Foulger*, 1989; *Foulger and Toomey*, 1989] and in 1991. These studies were independent, and provide a rare opportunity to study the repeatability of seismic tomography.

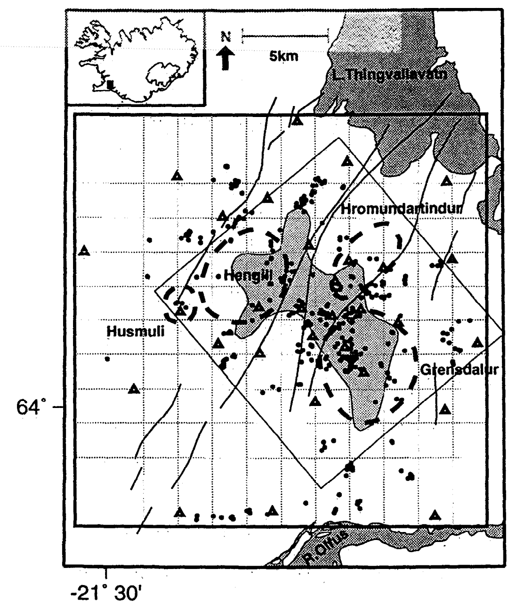

Figure 1. Map of the Hengill area, showing the main tectonic features. Dashed lines: eruptive sites of Húsmuli, Mt. Hengill, Mt. Hromundartindur, and Grensdalur. Shaded area: hot springs and fumaroles associated with the high-temperature geothermal area. Large dots: 1991 seismic stations. Small dots: 1991 earthquake epicenters. Small box: area studied using 1981 data [*Toomey and Foulger*, 1989]; large box: area studied using 1991 data and combined data. Values were computed at intersections of light grey lines. Station locations, nodal configuration, and events used for 1981 experiment are given by *Toomey and Foulger* [1989].

The 1981 Data and Inversion

In 1981, a temporary network of 23 analog seismic stations recorded 2000 locatable earthquakes [*Foulger*, 1988a;b]. *Toomey and Foulger* [1989] and *Foulger and Toomey* [1989] studied the three-dimensional variation of compressional-wave speed v_p by inverting arrival times for hypocentral parameters and crustal structure, using the SIMUL3 computer program of *Thurber* [1981; 1983]. A crustal block $14 \times 15 \times 6$ km was parameterized using nodes spaced at intervals of 2 and 3 km horizontally and 1 km vertically. The initial starting model was obtained from preliminary test inversions. The damping was set to 2 s^2 km^{-1}, after experimenting with different values (Table 1, inversion 1).

Wave speeds in the final model differ by +20%/−47% from those in the one-dimensional starting model. The major structural features are high-v_p bodies at depths

Paper number 94GL03387
0094-8534/95/95GL-03387$03.00

Table 1. Details of the simultaneous inversions discussed in the text

Inversion Number		EQs	Shots	Stat.	Arrivals P	Arrivals S	Model Dimensions (km)	Nodes	P RMS Residual (s)	P&S RMS Residual (s)
1	1981 data [*Toomey & Foulger*, 1989]	158	2	20	2409	0	$14 \times 15 \times 6$	$8 \times 8 \times 7 = 448$	0.043	
2	1981 data (with outliers removed)	158	2	20	2394	0	$14 \times 15 \times 6$	$8 \times 8 \times 7 = 448$	0.022	
3	1991 data (graded inversion)	228	1	33	4748	3678	$24 \times 24 \times 6$	$12 \times 11 \times 7 = 924$	0.020	0.038
4	1991 data (one-step inversion)	228	1	33	4748	3678	$24 \times 24 \times 6$	$12 \times 11 \times 7 = 924$	0.020	0.038
5	1981 & 1991 data (graded inversion)	386	3	55	7253	3678	$24 \times 24 \times 6$	$12 \times 11 \times 7 = 924$	0.023	0.036
6	1981 & 1991 data (one-step inversion)	386	3	55	7253	3678	$24 \times 24 \times 6$	$12 \times 11 \times 7 = 924$	0.023	0.037

of 0 to 3 km beneath the Grensdalur volcano, 2 to 4 km beneath the Hromundartindur system, and 0 to 4 km beneath Húsmuli. Coherent bodies with low velocities are absent except for a small (≈ 5 km^3) body with a v_p contrast of -7% beneath the northern edge of Mt. Hengill. The high v_p bodies were interpreted as solidified gabbroic intrusions and the low v_p body as possibly a small volume of partial melt. These results agree well with geology, tectonic structure, and gravity.

An improved 1981 data set, with 15 outliers removed, was inverted using SIMULPS12 [*Eberhart-Phillips*, 1993; *Evans et al.*, 1994], a modified version of SIMUL3 that uses pseudobending ray tracing [*Um and Thurber*, 1987] and can invert both *P* and *S-P* times to obtain v_p and v_p/v_s [*Thurber*, 1993]. The results are similar, showing that the modifications to SIMUL3 have a small effect for this data set (Table 1, inversion 2).

The 1991 Field Experiment and Data

The Hengill area was revisited in 1991, using 30 stand-alone seismic stations with Mark Products model L22D 2-Hz, three-component sensors and REFTEK model 72A data loggers. Data were recorded continuously at a sampling rate of 100 Hz for two months, during which time about 4,000 earthquakes were recorded.

Arrival times for 390 earthquakes were measured with an estimated accuracy of 0.01 s for *P* waves and 0.02 s for *S*. *S*-wave times were picked only from horizontal components that showed clear, impulsive arrivals.

Inversion of the 1991 Data

Inversions were performed using SIMULPS12 (Table 1, inversions 3 and 4), and data from 228 well-distributed earthquakes and one explosion.

The crustal block analyzed is $24 \times 24 \times 6$ km in dimensions, and has almost three times the volume of the region studied using the 1981 data. Figure 1 shows its location and the distributions of stations and earthquakes used. A laterally homogeneous initial v_p model was obtained using the program VELEST [*Kissling et al.*, 1994]. A starting v_p/v_s ratio of 1.77 was calculated using a modified Wadati diagram [*Chaterjee et al.*, 1985].

A series of "graded inversions" was performed with the number of nodes being progressively increased to allow the wave-speed of poorly sampled volumes at the periphery of the grid to be adjusted to the best values in the early inversions [*Eberhart-Phillips*, 1993]. *P* and *S* times were used in all inversions, but the v_p/v_s model was held fixed until the final inversion, a joint inversion for v_p and v_p/v_s with a horizontal node spacing of 2 to 4 km (Table 1, inversion 3). Damping values for each inversion were obtained by analyzing "trade-off curves" of data variance reduction against model variance [*Eberhart-Phillips*, 1986]. The node spacing is the factor most affecting these curves, with the optimum v_p damping parameter decreasing from 20 s^2 km^{-1} (12 km spacing) to 5 s^2 km^{-1} (2 km spacing). However, using 5 s^2 km^{-1} in the final inversion results in large v_p oscillations in the surface layer, so 20 s^2 km^{-1} was used for all inversions. v_p/v_s damping was 2 s.

To test the stability of the procedure, we conducted a single-step inversion for v_p and v_p/v_s, starting with the same initial conditions (Table 1, inversion 4). Damping parameters were 5 s^2 km^{-1} for v_p and 2 s for v_p/v_s. The results are similar to those of inversion 3, with RMS differences of 0.12 km/s for v_p and 0.01 for v_p/v_s. Inversion 4 yielded smaller v_p variations than inversion 3, especially in the surface layer, but the final RMS data residuals are the same (Table 1).

Our data show evidence of seismic anisotropy. The *S*-wave travel times before inversion vary systematically with azimuth by about 4%, with the slowest waves travelling sub-parallel to the ridge (N25°E). To investigate the effect of anisotropy, we applied an empirical sinusoidal correction to the *S*-wave arrival times and repeated the inversion for v_p/v_s. The resulting model differs only slightly from that obtained using uncorrected data, indicating that anisotropy does not affect our results to first order, probably because the azimuthal ray distribution is sufficiently uniform.

The resolution was assessed using the spread function

$$\text{spread} = \left[\|R_j\|^{-2} \Sigma_k D_{jk}^2 R_{jk}^2\right]^{1/2}, \qquad (1)$$

where $\mathbf{R}$ is the resolution matrix, $\|R_j\|$ is the Euclidean (L2) norm of its jth row, and D_{jk} is the distance between the jth and kth nodes. The spread indicates how widely the wave speed is averaged to yield the nodal values. Examination of individual rows of the resolution matrix and the ray distribution suggests that nodes with spread ≤ 4 km are well resolved and involve only local averaging.

Inversion of Combined Data

We inverted a combination of the 1981 and 1991 data, using the same grid, starting velocities, and damping parameters as with the 1991 data. The less accurate 1981 data were given half weight. Figure 2 shows the models resulting from the one-step inversion of the combined data set (Table 1, inversion 6). The graded inversion gives a marginally better data fit (Table 1, inversion 5), but has large v_p variations (up to +22%) in the surface layer.

Results

The v_p and v_p/v_s models obtained are insensitive to starting model, event set, grid configuration and inversion strategy. v_p varies laterally by −10% to +12% from its average value within each layer. The major v_p structures are high wave-speed (up to +12%) bodies near the extinct Grensdalur volcano at depths of 1 to 4 km, under the southern part of the Hromundartindur volcanic system at 2 to 4 km, and beneath the Húsmuli basalt shield at 0 to 3 km. At lower v_p contrasts ($\sim$ +2%) these bodies form a single zone oriented parallel to the spreading direction and traversing all three volcanic systems. No major low v_p bodies were imaged.

The ratio v_p/v_s varies by ±4% throughout the area. The most coherent anomaly involves low v_p/v_s at 0 to 2 km depth, and correlates closely with areas of hot springs and fumaroles (Figure 2). The mean v_p/v_s is 1.77, and shows no variation from the surface to 6 km.

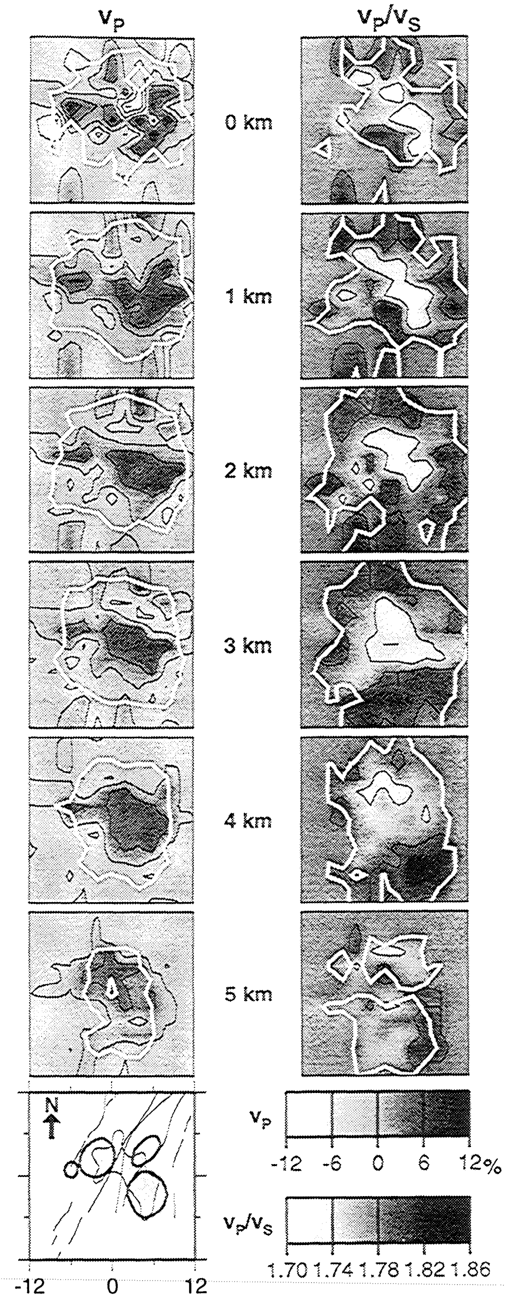

Figure 2. Horizontal cross sections of v_p and v_p/v_s structure obtained from 1981 and 1991 data (Table 1, inversion 6). For v_p, percentage difference from mean value in each layer is shown. Areas inside white lines are well resolved (*spread* < 4 km). Schematic tectonic features shown in Figure 1 are indicated at bottom left.

Interpretation

We interpret high v_p bodies within the volcanic complex as gabbro intrusions that were formerly the sources or conduits for surface volcanism.

The main factors affecting the ratio v_p/v_s are saturation [*Nur*, 1987], porosity, crack geometry and lithology. v_p/v_s commonly decreases with depth in situ [e.g., *Walck*, 1988; *Thurber and Atre*, 1993] because the closing of cracks affects v_s more than v_p. We find a mean v_p/v_s of 1.77 in all layers, a high ratio which suggests little tendency for cracks to close with depth above 6 km in the Hengill area.

The v_p/v_s variation of ±4% is much smaller than that found in other areas using local earthquake tomography [*Walck*, 1988; *Thurber and Atre*, 1993] and Wadati diagrams [*Chatterjee et al.*, 1985]. This suggests that there are only small variations in saturation and porosity across the region, and that the lithologies present have similar v_p/v_s ratios. No high v_p/v_s anomaly is observed, confirming that large volumes of partial melt are absent in the upper 6 km. The clear correlation of low v_p/v_s with the high-temperature geothermal area shows that it is a real feature related to the shallow part of the reservoir, possibly caused by alteration of rock to hydrated clay minerals, or changes in saturation or pore fluid temperature.

Comparing Results from the Separate and Combined Data Sets

In the Hengill area, a unique situation exists where two independent tomographic inversions have been conducted, along with a final combined inversion. The 1981 and 1991 data comprise separate events measured on different seismic networks, the crustal block was parameterized differently, and somewhat different inversion methods were used. The earlier inversion was for v_p only, so v_p/v_s cannot be compared between the two solutions.

The overall pattern of anomalies found in the two inversions is similar. High-v_p bodies are located beneath the Grensdalur and Hromundartindur systems and Húsmuli. The low-v_p body beneath the northern part of Mt. Hengill was not detected in the second inversion, so its existence is questionable.

The amplitudes of the v_p anomalies found using the 1991 data are smaller than those for the 1981 data, probably due to the larger errors in the 1981 data. Anomaly patterns agree better than absolute velocities. The RMS difference between v_p in the models is 0.26 km s^{-1}. It is slightly smaller (0.19 km s^{-1}) if variations from the mean in each layer are compared. This is about three times larger than the RMS difference between models obtained using the 1991 data set and different inversion strategies. Comparing the differences between the models with the calculated uncertainties of the 1981 model suggests that the statistical uncertainty underestimates the repeatability of v_p by a factor of about 5, a higher value than the factor of 2 estimated from quasi-empirical testing for the relationship between model uncertainty and accuracy by *Thurber* [1981].

Acknowledgments. This work was supported by a USGS G. K. Gilbert Fellowship, NERC Grant GR9/134, NERC Geophysical Equipment Loan 328/0990, and a loan of digital seismic equipment from IRIS/PASSCAL. A.D.M. is supported by a NERC Ph.D. studentship. H.M. Iyer, A.M. Pitt and R. Nowack gave helpful reviews. Data analysis was assisted by the GMT mapping software [*Wessel and Smith*, 1991].

References

Chaterjee, S. N., A. M. Pitt, and H. M. Iyer, v_p/v_s ratios in the Yellowstone National Park Region, Wyoming, *J. Volcanol. Geoth. Res., 26*, 213-230, 1985.

Eberhart-Phillips, D., Three-dimensional velocity structure in the northern California coast ranges from inversion of local earthquake arrival times, *Bull. Seismol. Soc. Am., 76*, 1025-1052, 1986.

Eberhart-Phillips, D., Local tomography: earthquake source regions, in *Seismic Tomography: Theory and Practice*, eds. H. M. Iyer and K. Hirahara, Chapman and Hall, 613-643, 1993.

Evans, J. R., D. Eberhart-Phillips, and C. H. Thurber, User's manual for SIMULPS12 for imaging v_p and v_p/v_s, a derivative of the Thurber tomographic inversion SIMUL3 for local earthquakes and explosions, *U.S. Geol. Surv. Open File Rep., 94-431*, 142 pp., 1994.

Foulger, G. R., The Hengill triple junction, SW Iceland: 1. Tectonic structure and the spatial and temporal distribution of local earthquakes, *J. Geophys. Res., 93*, 13,493-13,506, 1988a.

Foulger, G. R., The Hengill triple junction, SW Iceland: 2. Anomalous earthquake focal mechanisms and implications for process within the geothermal reservoir and at accretionary plate boundaries, *J. Geophys. Res., 93*, 13,507-13,523, 1988b.

Foulger, G. R. and D. R. Toomey, Structure and evolution of the Hengill-Grensdalur central volcano complex, Iceland: Geology, geophysics and seismic tomography, *J. Geophys. Res., 94*, 17,511-17,522, 1989.

Kissling, E., W. L. Ellsworth, D. Eberhart-Phillips, and U. Kradolfer, Initial reference models in local earthquake tomography, *J. Geophys. Res., 99*, 19,635-19,646, 1994.

Nur, A., Seismic rock properties for reservoir descriptions and monitoring, in *Seismic Tomography*, G. Nolet (ed.), D. Reidel Pub. Co., Holland, 203-237, 1987.

Thurber, C. H., *Earth structure and earthquake locations in the Coyote Lake area, central California,* Ph.D. thesis, Massachusetts Institute of Technology, Cambridge, MA., 332 pp, 1981.

Thurber, C. H., Earthquake locations and three-dimensional crustal structure in the Coyote Lake area, central California, *J. Geophys. Res., 88*, 8226-8236, 1983.

Thurber, C. H., Local earthquake tomography: velocities and v_p/v_s-theory, in *Seismic Tomography: Theory and Practice*, eds. H. M. Iyer and K. Hirahara, Chapman and Hall, 563-583, 1993.

Toomey, D. R. and G. R. Foulger, Application of tomographic inversion to local earthquake data from the Hengill-Grensdalur central volcano complex, Iceland, *J. Geophys. Res., 94*, 17,497-17,510, 1989.

Um, J. and C. H. Thurber, A fast algorithm for two-point seismic ray tracing, *Bull. Seismol. Soc. Am., 77*, 792-786, 1987.

Wessel, P. and W. H. F. Smith, Free software helps map and display data, *EOS Trans. AGU, 72*, 441, 445-446, 1991.

J. R. Evans and B. R. Julian, U.S. Geological Survey, 345 Middlefield Road, Mail Stop 977, Menlo Park, CA 94025 (e-mail: evans or julian@andreas.wr.usgs.gov).

G. R. Foulger and A. D. Miller, Dept. Geological Sciences, University of Durham, U.K. (e-mail: g.r.foulger or a.d.miller@durham.ac.uk).

(received June 14, 1994; revised October 31, 1994; accepted November 2, 1994.)