GEOPHYSICAL RESEARCH LETTERS, VOL. 34, L14314, doi:10.1029/2007GL030215, 2007

Rayleigh-wave group-velocity of the Icelandic crust from correlation of ambient seismic noise

Ó. Gudmundsson,[1] A. Khan,[2] and P. Voss[3]

Received 13 April 2007; revised 11 June 2007; accepted 22 June 2007; published 26 July 2007.

[1] By cross-correlating 2 years of ambient seismic noise from 31 broadband stations distributed across Iceland, we have been able to measure Rayleigh wave group velocity in three discrete frequency bands centered on 4, 7 and 17 s, sensitive to approximately the top 6, 10 and 25 km of the crust, for 289, 420 and 139 station-to-station paths, respectively. These are inverted to yield tomographic group velocity maps of the Icelandic crust. The results correlate with the major tectonic features on Iceland, and agree well with earlier seismic studies based on traditional methods (active source and earthquake). Specifically, we obtain low velocities in the central parts of Iceland, coinciding with the rift zones and young crust. In contrast, the parts of Iceland covered with older Tertiary volcanics reveal relatively higher velocities. At the shorter periods (4 and 7 s) this correlation reflects porosity, degree of fracturing and alteration. At the longest period (17 s), i.e. at greater depth, this correlation may reflect the thermal state of the crust, with warm and seismically slow material near the rifts and cooler and seismically faster material beneath the Tertiary formations to the east and west.
Citation: Gudmundsson, Ó., A. Khan, and P. Voss (2007), Rayleigh-wave group-velocity of the Icelandic crust from correlation of ambient seismic noise, *Geophys. Res. Lett.*, *34*, L14314, doi:10.1029/2007GL030215.

1. Introduction

[2] Traditional seismic imaging techniques have been augmented in recent years with the theoretical and experimental demonstration that the Green's function for an elastic medium can be retrieved by cross-correlating ambient seismic noise recorded at two seismographs [e.g., *Lobkis and Weaver*, 2001; *Roux et al.*, 2005; *Larose et al.*, 2005; *Shapiro and Campillo*, 2004; *Sabra et al.*, 2005]. In practice, the method amounts to extracting Green's functions from a fully diffusive wavefield, composed of waves with random phases and amplitudes, and propagating in all possible directions [e.g., *Weaver and Lobkis*, 2001]. Numerous studies have used such cross-correlation of long-term seismic noise records from several pairs of stations from which emerge coherent broadband dispersive Rayleigh wavetrains, demonstrating that a statistical treatment of seismic noise provides an additional means of obtaining information on the subsurface elastic structure of the Earth [e.g., *Shapiro et al.*, 2005], the Moon [*Larose et al.*, 2005] and the Sun [*Duvall et al.*, 1993].

[3] This passive seismic imaging technique provides an intriguing means of obtaining information on an area relatively densely populated with seismometers such as Iceland, whose crustal and upper mantle structure has been investigated previously by traditional seismic techniques, but for which the velocity structure, especially shear velocity, of the upper 20 km of the crust is relatively poorly known.

[4] Iceland is situated on the Mid-Atlantic Ridge where it interacts with the Icelandic hotspot, believed to be centered on the NW Vatnajökull icecap (see Figure 1) and of particular importance to our understanding of ridge-plume interaction. Much effort has gone into investigating its geological, geochemical and geophysical characteristics. Recent geophysical studies include work on mantle structure using surface waves [*Li and Detrick*, 2006; *Pilidou et al.*, 2004; *Allen et al.*, 2002a, 2002b] and body waves [*Kumar et al.*, 2005; *Foulger et al.*, 2000; *Wolfe et al.*, 1997], crustal thickness from receiver functions and surface waves [*Du and Foulger*, 2001; *Allen et al.*, 2002a; *Darbyshire et al.*, 2000; see also *Schlindwein*, 2006], and crustal structure from refraction profiling and local seismicity [e.g., *Yang and Shen*, 2005; *Darbyshire et al.*, 1998; *Staples et al.*, 1997; *Bjarnason et al.*, 1993]. Iceland is clearly underlain by a cylindrical low-velocity anomaly at depth in the upper mantle [e.g., *Wolfe et al.*, 1997] and a low velocity anomaly at the top of the upper mantle that is stretched along the rifts to the north and southwest [*Pilidou et al.*, 2004; *Li and Detrick*, 2006]. The crust is up to 40 km thick [e.g., *Darbyshire et al.*, 2000] and the lower crust is anomalously dense [e.g., *Gudmundsson*, 2003]. Less is known about the middle and upper crust except for localized information along profiles [e.g., *Darbyshire et al.*, 2000] and beneath seismographs [e.g., *Du and Foulger*, 2001]. Better distributed but cruder information is available from older refraction profiles (see summary by *Flóvenz and Gunnarsson* [1991]) indicating low velocities in the upper crust around the neovolcanic zones compared to the Tertiary formations in the west and east of the country attributed to porosity and increasing degree of alteration with age [*Flóvenz*, 1980].

[5] It is the purpose here to apply the passive seismic imaging technique to ambient noise from an array of 31 broadband stations, distributed over most of Iceland (see Figure 1), to produce maps of group-velocity at periods primarily sensitive to crustal shear-velocity structure. This study presents an independent and complementary way of

[1]Institute of Earth Sciences and Nordic Volcanological Centre, University of Iceland, Reykjavik, Iceland.
[2]Niels Bohr Institute, University of Copenhagen, Copenhagen, Denmark.
[3]Geological Survey of Denmark and Greenland, Copenhagen, Denmark.

0094-8276/07/2007GL030215$05.00

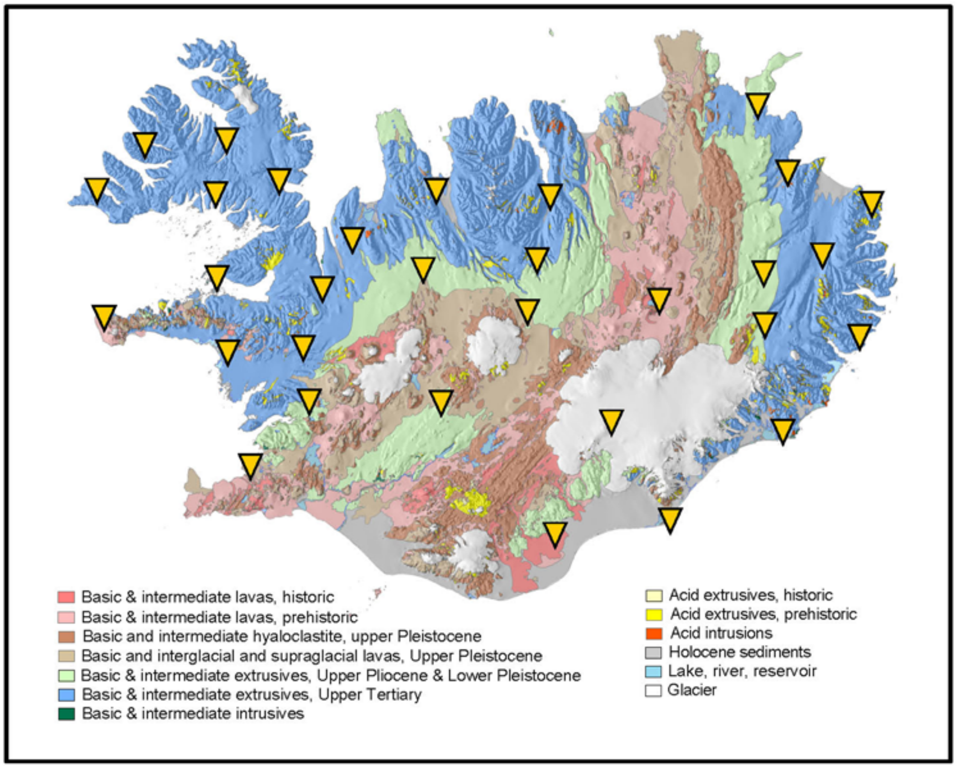

Figure 1. Locations of Hotspot stations and geological map of Iceland. Adapted from a map by Iceland Land Survey, with permission.

acquiring information, as we are able to sound to a depth of 25 km.

2. Method of Analysis

[6] We processed continuous data available from the Hotspot experiment [*Allen et al.*, 1999] and BORG station, which extended from July 1996 to August 1998. As large earthquakes would dominate any cross-correlation, we removed these prior to correlation. The threshold for removal was set at twice the standard deviation of the noise. Cross-correlated traces were computed for a month at a time from corrected traces $\varphi_i(t)$ and $\varphi_j(t)$ for a pair of stations i, j as

$$C_{ij}(\tau) = \int_0^T \varphi_i(t+\tau)\varphi_j(t)dt/T$$

The monthly correlations were then stacked keeping only those where a clear Rayleigh wave was present. After stacking we filtered the correlograms in three frequency bands centered on 4, 7 and 17 s and computed their envelope. Examples are shown in Figure 2, where a well-defined wave packet is apparent for both directions of the correlation, i.e. for positive lag times ($\tau > 0$), arising from Rayleigh waves that propagate from j to i and sum up coherently, as well as for negative lag times ($\tau < 0$), due to Rayleigh waves traveling in the opposite direction. Other directions of wave travel produce incoherent additions to the correlograms and are visible as the remaining fluctuations in the correlations [*Snieder*, 2004; *Roux and Kuperman 2004*]. Finally, we determined group times at the three periods by picking the envelope travel time, i.e. the center of the pulse for the correlated traces with a signal-to-noise ratio above a certain threshold. This resulted in a total of 848 group-time measurements. Using synthetic tests, we estimated the uncertainty of arrival times by determining how the uncertainty varied with central period (T_0) and signal-to-ratio (A) and found it to be proportional to T_0 and inversely proportional to A.

[7] Microseismic noise arises from coupling between the atmosphere, ocean and seafloor [*Rhie and Romanowicz*, 2004]. Wind-generated gravity waves on the ocean are converted locally to seismic energy that propagates away from the source region primarily as Rayleigh waves. Noise in the range 0.05–0.1 Hz is due to direct interaction of waves with the seafloor, while noise in the range 0.1–0.25 Hz arises from non-linear wave interactions in shallow water doubling the wave frequency [*Webb*, 1998]. *Wilcock et al.* [1999] found that the level of microseismic noise fluctuates with the seasons in the North Atlantic. In the summer, large regions of the ocean are often calm, while in the winter, there are many big storms near Iceland, resulting in higher seismic noise peaks. We find that correlograms often produce a clearer dispersive wave packet in winter than in summer (presumably due to a higher noise level and a better distributed source). We also find a clear seasonal variation in frequency content of the noise. The secondary noise peak has a higher amplitude and consistently extends to lower frequencies (0.12–0.16 Hz) in winter than in summer (see Figure 3).

[8] In order to produce velocity maps from the measured interstation Rayleigh wave group times, we employed the stochastic inversion scheme of *Franklin* [1970], assuming that the propagation paths are straight, connecting the two

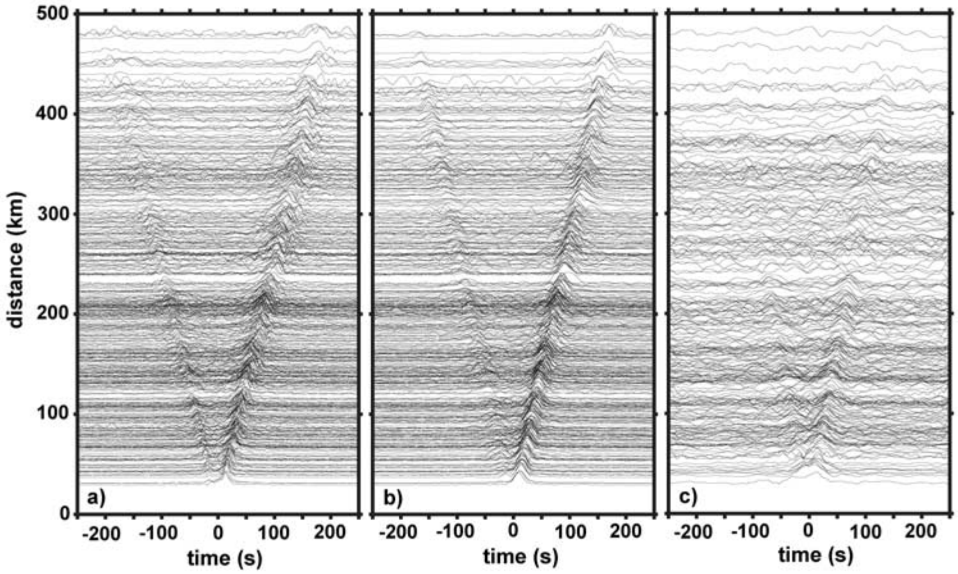

Figure 2. Envelopes of stacked correlograms as a function of distance in the three frequency bands used; (a) at 4 s, (b) at 7 s and (c) at 17 s.

stations. We invert for perturbations in group slowness from a reference slowness taken as the inverse average group velocity in each frequency band (2.4, 2.7 and 3.2 km/s at 4, 7 and 17 s, respectively). The model is parametrized into a spherical grid with a cell size of 0.2° by 0.4° (latititude, longitude; approx. 20 km by 20 km), resulting in 600 model parameters in all. Data uncertainties are assumed to be independent and equal for all the data. The model covariance is assumed to take a Gaussian form with a correlation length (half width) equal to the maximum width of the Fresnel zone $\sqrt{(3\lambda/2 \cdot \tan(\Delta/2)}$, where λ is the wavelength at some reference velocity and Δ is the mean distance between pairs of stations [*Spetzler and Snieder*, 2004]. Correlation lengths used are 48, 75 and 104 km, respectively, at the frequencies of interest (4, 7 and 17 s). The amplitude ratio of the data and model covariance is taken as a free tuning parameter for the inversion and plays the role of a damping parameter that controls the tradeoff between residual data and model variance, and is determined by computing the L-curve for various values of damping, thus resulting in varying data and model variance [*Menke*, 1984]. Values used are 20, 60 and 50 km at 4, 7 and 17 s, respectively. Prior to inversion the data set was inspected for possible outliers. The variance reduction achieved by the inversion was 73%, 55% and 27% with 64, 30 and 16 degrees of freedom in the models at 4, 7 and 17 s, respectively. The residual data have a standard deviation of approximately 2 s at all periods and exceed uncertainty estimates significantly at 4 and 7 s periods indicating some inadequacy of theory.

3. Results and Discussion

[9] In Figure 4 we display tomographic resolution and velocity maps at the three frequencies. The number of rays traversing each grid cell (Figures 4d–4f) is an indicator of resolution. From these maps we can see that the best data coverage is at 7 s, slightly worse at 4 s and somewhat worse at 17 s, reflecting the numbers of observations in the respective frequency bands (420 at 7 s, 289 at 4 s and 139 at 17 s). These periods correspond to sensitivity to depths of around 6, 10 and 25 km. Poorly sampled regions are limited to the southwestern part and northern tip of Iceland. The resolution length is about 50, 75 and 100 km where sampling is densest, to a large degree controlled by the correlation length of the imposed a priori model covariance.

[10] The tomographic maps (Figures 4a–4c) show velocity variations computed from the tomographic group-slowness model. They correlate well with surface tectonics and the geologic map in Figure 1. Velocities in the central part of Iceland, coinciding with the location of the rift zones (see Figure 1), are significanty lower than the surroundings, and vary by as much as 40% at 4 s, 30% at 7 s and 15% at 17 s (peak-to-peak). Velocities in the eastern and western parts of Iceland are higher. This pattern is quite continuous

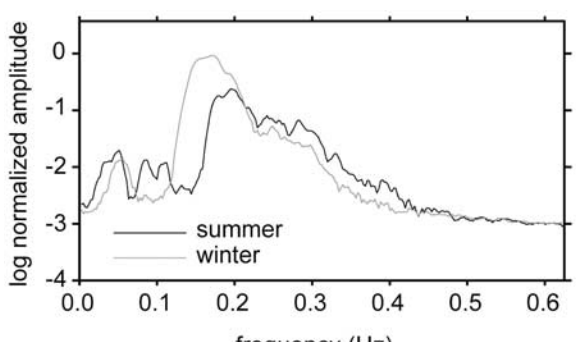

Figure 3. The average amplitude spectra of correlograms from station pair HOT01 and HOT16 for winter months (gray curve) and summer months (black curve). These spectra are typical for all station pairs. The spectra of correlograms are proportional to the spectra of noise at the two stations used.

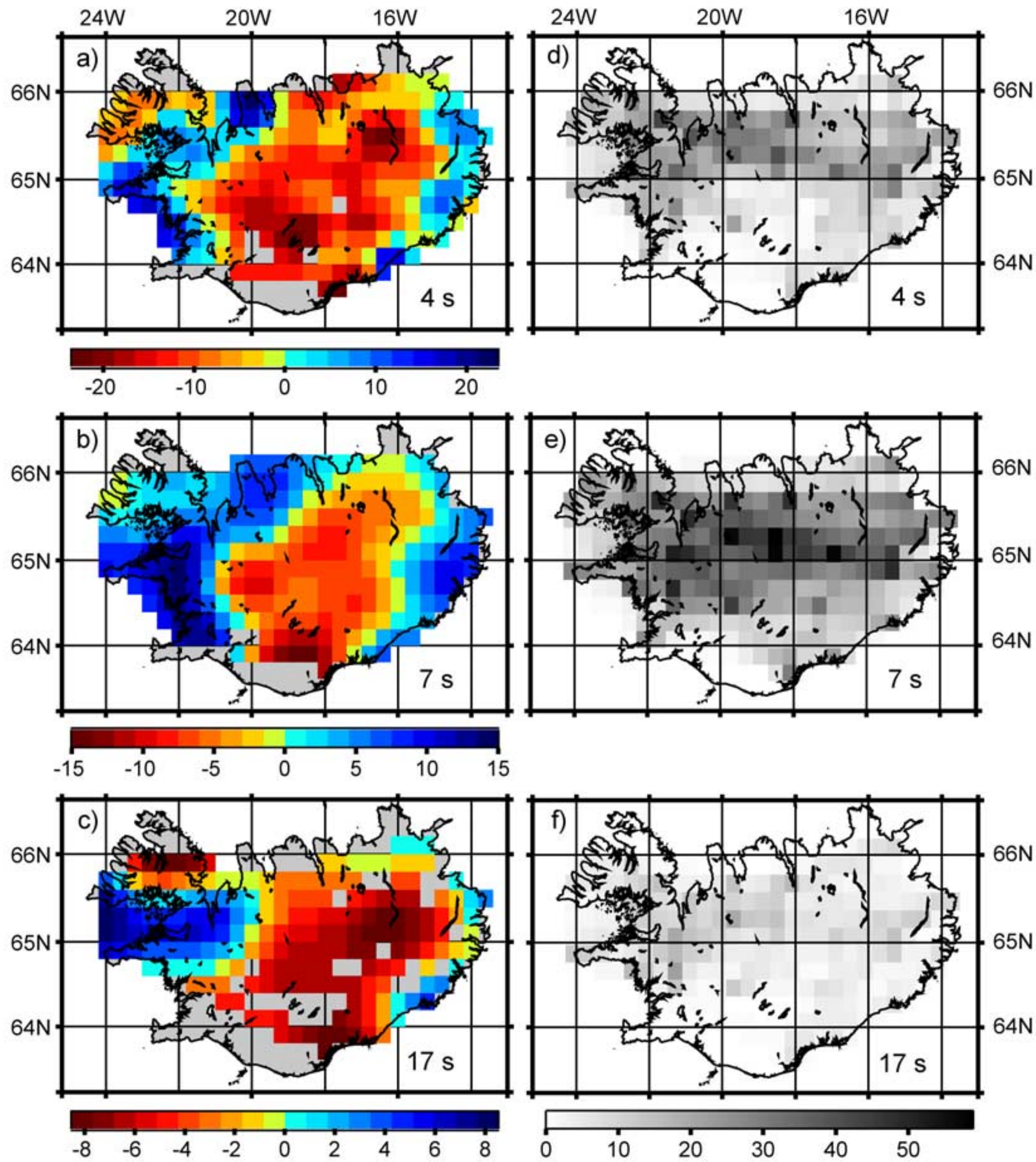

Figure 4. (a)–(c) Group velocity maps (percentile variation from reference given in text) and (d)–(f) maps of hitcounts for the three frequency bands used.

with period, with the strongest deviations at the model edges where resolution is poorest. The overlapping volcanic zones in Iceland are separated by ~100 km and are not clearly distinguished in the result as this separation is on the order of the resolution length. The results do, however, suggest a decreasing signature of the western volcanic zone with increasing period (i.e increasing depth) lending support to the common belief that this is a dying rift. Low velocities in NW Iceland coincide with a Bouguer gravity low and may at 17 s period indicate thickened crust in that part of the country.

[11] The uppermost 3 to 10 km of the Iceland crust (upper crust) are characterized by a strong velocity gradient based on *P*-wave refraction results [*Flóvenz and Gunnarsson*, 1991]. This is explained by a reduction of porosity and elevation of alteration state with depth [*Flóvenz*, 1980]. This is verified by a clear correlation across the 1500 m depth range in the upper crust that is exposed by varying degrees of erosion of the extrusive basalt pile. Therefore we conclude that the variation of group velocity at the shorter periods, 4 and 7 s, corresponding to sensitivity to 6 and 10 km depth, is primarily caused by these effects. They can accommodate a ±20% variation as is modelled at 4 s period (Figure 4a) simply by variations of upper crustal thickness because *P*-wave velocity variations in the upper crust exceed that level [*Flóvenz and Gunnarsson*, 1991]. At 17 s period the explanation may lie in decreasing temperature with age or distance from central Iceland as suggested by, for example, *Li and Detrick* [2006], who obtained shear-wave velocities in the lower crust by inversion of Rayleigh waves in the period range 20–125 s showing reduced velocities beneath the rift zones, surrounded by relatively higher velocities to the east and west. Our work at 4, 7 and 17 s can be regarded as an upward extension of their work

and the general agreement of our results at 17 s and their results at 20 s is encouraging, although we have not deconvolved the effects of shallow shear velocity on the group velocity. Velocity heterogeneity at shallow depths (12 km) has also been mapped by *Yang and Shen* [2005] in their study of *P*-wave structure using local earthquakes. There is no clear correlation between their results (presented at 12 km depth) and our results at 7 s. Velocity variations at 17 s may be affected by variations of crustal thickness.

[12] Low velocities in the middle and lower crust can be accounted for by invoking elevated temperatures and possible localized presence of partial melt. *Darbyshire et al.* [2000], suggest presence of accumulated melt in the cust as an explanation of a zone of low shear-wave velocity in the depth range 10–14 km near Krafla, northern Iceland. *Maclennan et al.* [2001] find that modeled *P*-wave velocities at depths from 4 to 20 km based on chemical analyses of crustal samples from Krafla and Þeistareykir, northeastern Iceland, agree with a previous seismic survey in the Krafla region, and suggest that cumulates are a significant component of the Icelandic lower crust.

[13] Crustal velocity is relatively low down to a depth of 20–25 km in central Iceland, where the crust appears to be thick (up to ∼40 km). If the crust is that thick, the average crustal density must be high where the crust is thickest [*Gudmundsson*, 2003]. This suggests a concentration of high crustal density in the lowermost crust (25–40 km depth) rendering the lower crust barely distinguishable from the mantle.

[14] **Acknowledgments.** We acknowledge the IRIS data centre for help with downloading data, the scientists that conducted the HOTSPOT experiment for producing a unique data set and the Iceland Land Survey for permission to use the geologic map in Figure 1. A. Khan was supported by a grant from the Danish Agency for Science, Technology and Innovation. Ó. Gudmundsson was partially funded by the Danish Research Foundation.

References

Allen, R. M., et al. (1999), The thin hot plume beneath Iceland, *Geophys. J. Int.*, *137*, 51–63.

Allen, R. M., et al. (2002a), Plume-driven plumbing and crustal formation in Iceland, *J. Geophys. Res.*, *107*(B8), 2163, doi:10.1029/2001JB000584.

Allen, R. M., et al. (2002b), Imaging the mantle beneath Iceland using integrated seismological techniques, *J. Geophys. Res.*, *107*(B12), 2325, doi:10.1029/2001JB000595.

Bjarnason, I. T., W. Menke, Ó. Flovenz, and D. Caress (1993), Tomographic image of the Mid-Atlantic plate boundary in southwestern Iceland, *J. Geophys. Res.*, *98*, 6607–6622.

Darbyshire, F. A., I. T. Bjarnason, R. S. White, and Ó. Flovenz (1998), Crustal structure above the Iceland mantle plume imaged by the ICE-MELT refraction profile, *Geophys. J. Int.*, *135*, 1131–1149.

Darbyshire, F. A., K. F. Priestley, R. S. White, R. Stefánsson, G. B. Gudmundsson, and S. Jakobsdóttir (2000), Crustal structure of central and northern Iceland from analysis of teleseismic receiver functions, *Geophys. J. Int.*, *143*, 163–184.

Du, Z., and G. R. Foulger (2001), Variation in the crustal structure across central Iceland, *Geophys. J. Int.*, *145*, 1–24.

Duvall, T. L., S. M. Jefferies, J. W. Harvey, and M. A. Pomerantz (1993), Time-distance helioseismology, *Nature*, *362*, 430–432.

Flóvenz, Ó. G. (1980), Seismic structure of the Icelandic crust above layer 3 and the relation between body wave velocity and the alteration of the basaltic crust, *J. Geophys.*, *47*, 211–220.

Flóvenz, Ó., and K. Gunnarsson (1991), Seismic crustal structure in Iceland and surrounding area, *Tectonophysics*, *189*, 1–17.

Foulger, G. R., et al. (2000), The seismic anomaly beneath Iceland extends down to the mantle transition zone and no deeper, *Geophys. J. Int.*, *142*, F1–F5.

Franklin, J. N. (1970), Well-posed stochastic extensions of ill-posed linear problems, *J. Math. Anal. Appl.*, *31*, 682–716.

Gudmundsson, Ó. (2003), The dense root of the Iceland crust, *Earth Planet. Sci. Lett.*, *206*, 427–440.

Kumar, P., et al. (2005), The lithosphere-asthenosphere boundary in the North-West Atlantic region, *Earth Planet. Sci. Lett.*, *236*, 249–257.

Li, A., and R. S. Detrick (2006), Seismic structure of Iceland from Rayleigh wave inversions and geodynamic implications, *Earth Planet. Sci. Lett.*, *241*, 901–912.

Larose, E., A. Khan, Y. Nakamura, and M. Campillo (2005), Lunar subsurface investigated from correlation of seismic noise, *Geophys. Res. Lett.*, *32*, L16201, doi:10.1029/2005GL023518.

Lobkis, O. I., and R. L. Weaver (2001), On the emergence of the Green's function in the correlations of a diffuse field, *J. Acoust. Soc. Am.*, *110*, 3011–3017.

Maclennan, J., D. McKenzie, K. Grönvold, and L. Slater (2001), Crustal accretion under northern Iceland, *Earth Planet. Sci. Lett.*, *191*, 295–310.

Menke, W. (1984), *Geophysical Data Analysis: Discrete Inverse Theory*, Academic, Orlando, Fla.

Pilidou, S., K. F. Priestley, Ó. Gudmundsson, and E. Debayle (2004), Upper mantle *S*-wave-speed heterogeneity and anisotropy beneath the North Atlantic from regional surface-wave tomography: The Iceland and Azores plumes, *Geophys. J. Int.*, *159*, 1057–1076.

Rhie, J., and B. Romanowicz (2004), Excitation of Earth's continuous free oscillations by atmosphere-ocean-seafloor coupling, *Nature*, *431*, 552–556.

Roux, P., and W. A. Kuperman (2004), Extracting coherent wave fronts from acoustic ambient noise in the ocean, *J. Acoust. Soc. Am.*, *116*, 1995–2003.

Roux, P., K. G. Sabra, W. A. Kuperman, and A. Roux (2005), Ambient noise cross correlation in free space: Theoretical approach, *J. Acoust. Soc. Am.*, *117*, 79–84.

Sabra, K. G., P. Gerstoft, P. Roux, W. A. Kuperman, and M. C. Fehler (2005), Surface wave tomography from microseisms in Southern California, *Geophys. Res. Lett.*, *32*, L14311, doi:10.1029/2005GL023155.

Schlindwein, V. (2006), On the use of teleseismic receiver functions for studying the crustal structure of Iceland, *Geophys. J. Int.*, *164*, 551–568.

Shapiro, N. M., and M. Campillo (2004), Emergence of broadband Rayleigh waves from correlations of the ambient seismic noise, *Geophys. Res. Lett.*, *31*, L07614, doi:10.1029/2004GL019491.

Shapiro, N. M., M. Campillo, L. Stehly, and M. H. Ritzwoller (2005), High-resolution surface-wave tomography from ambient seismic noise, *Science*, *307*, 1615–1618.

Snieder, R. (2004), Extracting the Green's function from the correlation of coda waves: A derivation based on stationary phase, *Phys. Rev. E.*, *69*, 046610, doi:10.1103/PhysRevE.69.046610.

Spetzler, J., and R. Snieder (2004), Tutorial: The Fresnel volume and transmitted waves, *Geophysics*, *69*, 653–663.

Staples, R. K., R. S. White, B. Brandsdottir, P. Einarsson, H. Shimamura, and H. Shiobara (1997), Faroe-Iceland ridge experiment: 1. Crustal structure of northeastern Iceland, *J. Geophys. Res.*, *102*, 7849–7866.

Weaver, R. L., and O. I. Lobkis (2001), Ultrasonics without a source: Thermal fluctuation correlations at MHz frequencies, *Phys. Rev. Lett.*, *87*, 134301, doi:10.1103/PhysRevLett.87.134301.

Webb, S. C. (1998), Broadband seismology and noise under the ocean, *Rev. Geophys.*, *36*, 105–142.

Wilcock, W. S. D., S. C. Webb, and I. T. Bjarnason (1999), The effect of local wind on seismic noise near 1 Hz at the MELT site and in Iceland, *Bull. Seismol. Soc. Am.*, *89*, 1543–1557.

Wolfe, C. J., I. T. Bjarnason, J. C. VanDecar, and S. C. Solomon (1997), Seismic structure of the Iceland mantle plume, *Nature*, *385*, 245–247.

Yang, T., and Y. Shen (2005), *P*-wave velocity structure of the crust and uppermost mantle beneath Iceland from local earthquake tomography, *Earth Planet. Sci. Lett.*, *235*, 597–609.

Ó. Gudmundsson, Institute of Earth Sciences and Nordic Volcanological Centre, University of Iceland, Sturlugotu 7, 101 Reykjavik, Iceland. (olafurg@hi.is)

A. Khan, Niels Bohr Institute, University of Copenhagen, Juliane Maries Vej 30, DK-2100 Copenhagen, Denmark.

P. Voss, Geological Survey of Denmark and Greenland, Oester Voldgade 10, DK-1350 Copenhagen, Denmark.