

Deep downward fluid percolation driven by localized crust dilatation in Iceland

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Received 21 May 2008; revised 28 June 2008; accepted 3 July 2008; published 3 September 2008.

[1] From a detailed seismic survey on a transform-like plate setting in Iceland, we propose a spatial relationship between highly deformed crustal columns and areas of low V_p/V_s ratio, which are best explained by high fluid pressures at the base of the seismogenic crust. Combined with geological observations, our seismic data suggest transient and local increases in crust permeability at specific highly dilating areas triggered by dynamical stresses from high-magnitude regional earthquakes. Fluids are then trapped at supercritical conditions in the deep brittle crust which, in turn, increases the capacity of the rocks to deform through cracking and faulting. **Citation:** Geoffroy, L., and C. Dorbath (2008), Deep downward fluid percolation driven by localized crust dilatation in Iceland, *Geophys. Res. Lett.*, *35*, L17302, doi:10.1029/2008GL034514.

1. Introduction

[2] Hydrous fluids in the Earth's crust play a fundamental role in most, if not all, geological processes. Fluids notably reduce the normal stress acting on fractures and thus exert a first-order control on seismogenic displacements along faults in the brittle crust [e.g., *Byerlee*, 1990]. Moreover, the convective circulation of water increases the cooling of the upper layers of the lithosphere, especially at oceanic ridges. While it remains unclear how fluids can penetrate deep into the oceanic or continental crust, it is nevertheless considered that meteoric water can reach as deep as the ductile crust in regions of active deformation [e.g., *McCaig*, 1988].

[3] Iceland is an onshore exposure of the mid-Atlantic spreading ridge. With its active hydrothermalism, Iceland is an ideal laboratory to study the interplay between fluids and tectonics at oceanic ridges. The oceanic crust is abnormally thick, up to 40 km beneath the central area and decreasing to 15–20 km beneath the Reykjanes Peninsula [*Foulger et al.*, 2003]. The active rift system [e.g., *Einarsson*, 1991] (Figure 1 inset) is composed of two overlapping spreading axes connected southward through an EW-trending left-lateral shear zone, the Southern Iceland Seismic Zone (SISZ), which is historically affected by high-magnitude earthquakes along NS-trending dextral faults (e.g. 17th June 2000: $M_w = 6.5$). The Reykjanes Peninsula (RP) represents the northward prolongation of the Reykjanes oceanic ridge (Figure 1 inset). The area is made up of en-echelon volcano-tectonic segments (Figure 1). The central part of each volcano-

tectonic segment is associated with \sim N040E-trending sets of sub-parallel normal-faults, mode-I fissures and basaltic dikes inherited from the previous stages of magmatic activity during the Holocene [e.g., *Gudmundsson*, 1986]. The center of volcano-tectonic segments is a zone associated with marked hydrothermal activity [*Arnórsson*, 1995] (Figure 1).

2. The Experiment

[4] We installed a network of 18 seismic stations over the Grindavik and Krisuvik volcano-tectonic segments (Figure 1) during a 5-month period from 10th April 2005 to 30th August 2005. Over the same period, our data were supplemented by the arrival-times obtained at five nearby stations belonging to the permanent SIL network [*Stefánsson et al.*, 1993] (Figure 1). The installed network consisted of autonomous three-component continuous recording instruments of short-period type, operating with GPS time-synchronization. We recorded and located \sim 900 seismic events (magnitudes: 0 to 3.7), using the 1D local P-wave velocity model derived by *Weir et al.* [2001] from seismic refraction experiments. The travel-times of 724 high-quality events were then used for tomographic inversion using the double-difference tomography method of *Zhang and Thurber* [2003], providing a joint solution for velocity structure and event localization. The initial velocity model was interpolated from *Weir et al.*'s [2001] model, with a V_p/V_s ratio of 1.78 and an inversion grid spacing of 2 km in the three dimensions, down to the maximal depth of the hypocenters. The final uncertainty on relocation is estimated to be less than 100 m horizontally and 200 m vertically, which makes our experiment the most precise ever carried out in this area.

3. Main Results

3.1. Local Plate Boundary

[5] The epicenters in our survey are distributed along two short (\sim 7 km long) N060E-trending segments, which are connected to a longer (\sim 12 km) EW-trending segment (Figure 2a). In the Krýsuvík-Kleifarvatn sector, the epicenters are arranged in clusters, roughly aligned along the \sim N060E trend, and overlapping the hot-spring areas (Figures 1 and 2a). In the “Grindavik area” farther west, the epicenters do not form clusters, but instead are aligned along a single N060E fracture zone. Combining our results with those obtained from a similarly high-precision survey at the extremity of the RP [*Klein et al.*, 1977], we propose that the active plate boundary is more complex than the simple \sim N070E-trending dislocation model usually assumed [e.g., *Einarsson*, 1991]. Since the NNR-NUVEL-1A model [*Argus and Gordon*, 1991] yields a local plate velocity vector trending \sim N101.4E \pm 1°, we can infer that

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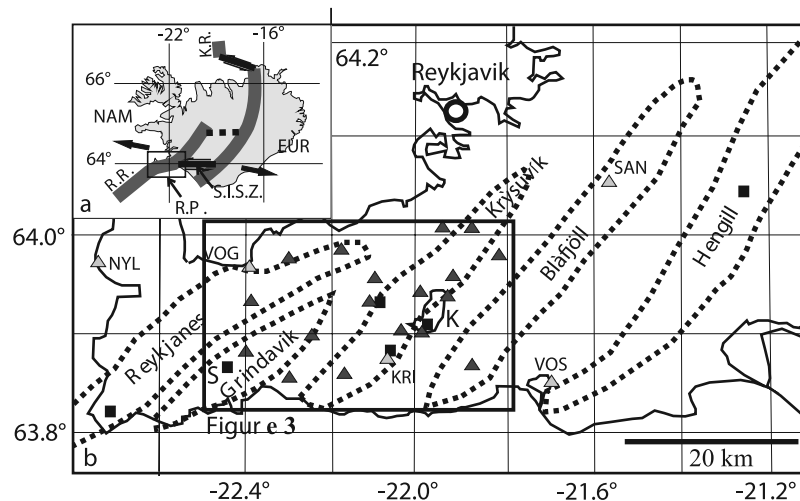


Figure 1. The Reykjanes Peninsula. Dotted lines: volcano-tectonic segments. Black square: main geothermal field. Triangles: seismic stations (dark grey: network set up in this study; light grey with letter codes: SIL stations); K: Kleifarvatn Lake; S: Svartsengi geothermal field. (inset) Active rift system in Iceland. Large black arrows: motion vectors for North American (NAM) and Eurasia (EUR) plates [Argus and Gordon, 1991]; R.R. and K.R.: Reykjanes and Kolbeinsey spreading axes; S.I.S.Z.: South Iceland Seismic Zone; R.P.: Reykjanes Peninsula.

the RP active plate boundary more likely resembles a transtensional segmented sinistral transform. In such a context, the \sim EW-trending segments are connected towards the center of the main volcano-tectonic segments, i.e. in areas with the greatest dilatation and active hydrothermalism (Figure 2b). Although the RP upper-crust is modeled as globally locked at \sim 7 km over a continuously deforming ductile lower crust and mantle [Arnadóttir *et al.*, 2006; Keiding *et al.*, 2007], the highest shear-strain rates inferred from GPS surveys (up to \sim $-0.7 \cdot 10^{-6}$ /yr) occurred close to the proposed E–W trending segments over the past 6 years [Keiding *et al.*, 2007] (Figure 2b). The \sim N060E segments are the loci of high-intensity but relatively unstable seismic deformation [Hjaltadóttir and Vogfjord, 2006], possibly resulting from the particular strain field between the N090E segments. We computed fifty-five double-couple focal mechanisms based on the inversion of P-wave polarities (Figure 2a). On average, the nodal planes trend N010E or N105E, and are associated with extensional oblique-slip displacements that range from dip-slip (predominantly normal) to chiefly extensional oblique strike-slip.

3.2. Seismic Tomography

[6] To interpret the recorded seismicity, we related P and S waves velocities to the hypocenter distribution (Figure 3 and auxiliary¹). As elsewhere in Iceland [Foulger *et al.*, 2003], the RP upper crust is highly heterogeneous. For $V_p < 4 \text{ km.s}^{-1}$ and $V_s < 2.5 \text{ km.s}^{-1}$ ($z < 2 \text{ km}$, on average), the RP upper crust is significantly aseismic, as commonly observed both in Iceland [Foulger *et al.*, 2003] and along the adjacent Reykjanes Ridge [Mochizuki *et al.*, 2000]. Two main areas of anomalously low V_p/V_s are identified in the lower part of the brittle crust (Figure 3), forming zones that are superimposed onto the \sim N060 high-dilatation crustal areas defined above (Figure 2). The first conspicuous

anomaly is broadly located beneath the Krýsuvík-Kleifarvatn geothermal areas, i.e. at the center of the Krýsuvík segment (Figures 2b and 3). This first anomaly increases in depth from \sim 3 km at the edge of the lake area to \sim 7 km in the center of the studied area. A second anomaly is located east of the exploited geothermal field of Svartsengi (Figure 1, Grindavík segment), at depths of between 4 km and 6 km. The Krýsuvík-Kleifarvatn anomaly at 3–6 km depth is marked by a strong decrease in P-wave velocity (-15% : 5.8 km.s^{-1} , instead of the mean value of 6.85 km.s^{-1} at 6 km depth; see auxiliary material) and a relatively less marked decrease in S velocities (-10% : 3.5 km.s^{-1} compared with an average of 3.87 km.s^{-1} at 6 km depth; see auxiliary material). The V_p/V_s ratio can be as low as 1.55/1.6, compared with the mean value of 1.78. The Grindavík anomaly is even more marked, with a V_p/V_s ratio falling to \sim 1.52 (i.e. -14%), which is due to the combined decrease of V_p and increase in V_s in this area (Figure 3 and auxiliary material). These low V_p/V_s anomalies at the base of the brittle crust compare closely with results obtained farther east beneath the Hengill volcano [e.g., Tryggvason *et al.*, 2002].

4. Discussion

[7] The clear relation between seismicity and low V_p/V_s zones, as well as the clustering of seismicity beneath the geothermal areas (especially in the Krýsuvík area, Figures 2 and 3), suggests that earthquakes are linked to high fluid pressures at depth. Indeed, it is unlikely that these anomalies are linked to crustal magma chambers, since, in that case, we would expect an increase of V_p/V_s [Iyer, 1992]. Because P-wave velocities are more strongly reduced (see auxiliary material), the anomalies more likely represent lateral variations of the bulk rock compressibility. This could result from two effects (that may be combined): i) the presence of reservoirs containing a compressible fluid [Julian *et al.*, 1996], ii), a decreasing influence of thin cracks (compared to thick cracks) on fracture porosity, due

¹Auxiliary materials are available in the HTML. doi:10.1029/2008GL034514.

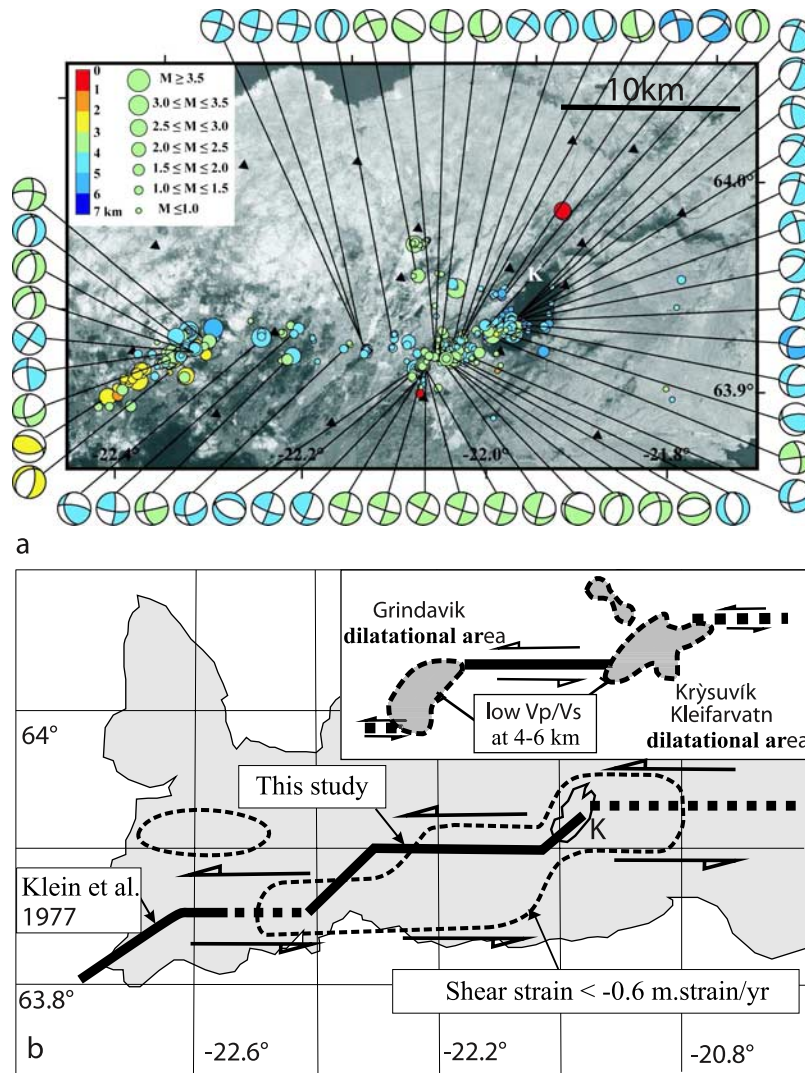


Figure 2. (a) Map view of relocated epicenters from 10th April to 30th August 2005, with fault mechanisms for 55 events shown on inset photo. Symbol size represents magnitude, colors indicate focal depth. (b) Tectonic interpretation. Plate separation in the RP is dominated by left-lateral transform-like shearing along N090E-trending segments arranged *en echelon* (bold dashed lines: inferred trace). Data from this study (see Figure 2a) and from Klein et al. [1977]. Short dashed lines delimit areas of maximum left-lateral shear inferred from GPS data during the period 2000–2006 [Keiding et al., 2007]. Inset: grey areas are ground surface projections of zones with anomalously low Vp/Vs ratio at 4–6 km depth (see Figure 3).

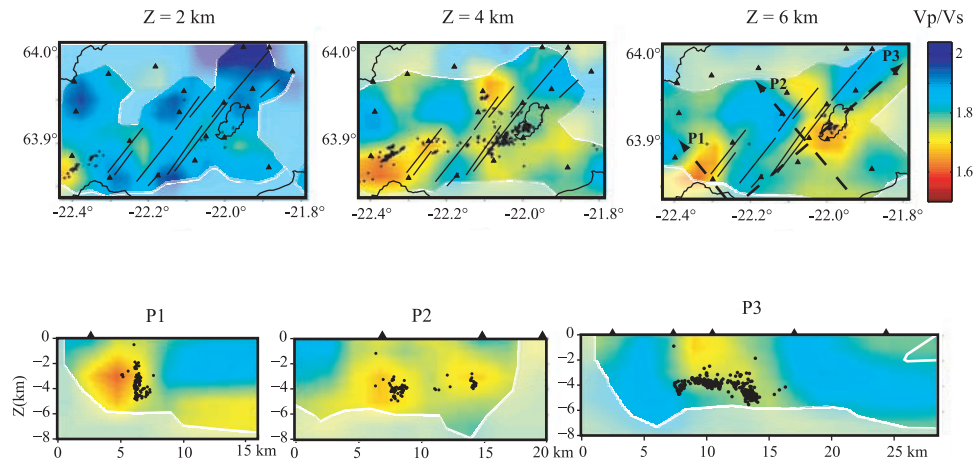


Figure 3. (top) Map view of Vp/Vs models at different depths (see auxiliary material). The color code bars are centered on the initial velocity within the layer, so the attributed velocities are different for each layer. Unresolved zones are indicated as opaque. (bottom) Vertical cross-sections through the 3D Vp/Vs model and corresponding events, along profiles P1, P2 and P3 (defined in the top right plot). Scale 1:1.

to sealing of the thinner cracks by hydrothermal minerals [Barclay *et al.*, 2001]. As in other studies [e.g., Zencher *et al.*, 2006], we propose that significant amounts of geothermal fluids exist in Iceland as deep as the base of the brittle crust. Given the high temperatures recorded in deep drill holes along the active rift system (e.g. $>300^{\circ}\text{C}$ at a depth of 2500 m at the Reykjanes volcano-tectonic center), these fluids are probably in a supercritical state [Friedleifsson and Elders, 2005] with high pressures, intermediate between hydrostatic and lithostatic [e.g., Arnórsson, 1995]. Supercritical fluids are highly corrosive compressible fluids thought to play a significant role in hydrofracturing [Hashida and Takahashi, 2004]. These fluid properties could account for the low P velocities associated with seismicity in the Grindavik and Krýsuvík-Kleifarvatn areas. The localized high fluid content at the base of the RP upper crust could reflect transient increases in crust permeability related to the dilatational strain field acting at the extremities of the postulated *en-échélon* EW-trending shear zones. During 2000, there was a spectacular example of a local but considerable increase in upper-crust permeability in the area around Kleifarvatn Lake, i.e. immediately above the Krýsuvík Vp/Vs anomaly. About 12% by volume of the lake drained down through the upper crust at an average rate of $2\text{ m}^3/\text{s}$ during the 18 months following 17th June 2000 [Clifton *et al.*, 2003]. The water drained through NE to NNE-trending open fissures visible at the surface, controlled by a concealed NS-trending dextral fault bounding the lake. This event was triggered by a major seismic crisis in the SISZ during June 2000 [Clifton *et al.*, 2003]. By combining these observations with our data, we propose a two-step conceptual model linking downward fluid penetration and deformation within the RP crust (Figure 4). The absence of earthquakes down to $\sim 2\text{--}3\text{ km}$ and the data from extensive geothermal surveys in Iceland [e.g., Arnórsson, 1995; Friedleifsson and Elders, 2005] suggest the presence of a dual fluid-reservoir in the brittle crust (see also converging arguments by Jónsson *et al.* [2003]). In the uppermost part of the crust ($z < \sim 2\text{--}3\text{ km}$), fluids are

composed of water or brine at boiling point conditions [Arnórsson, 1995; Friedleifsson and Elders, 2005] in a hydrostatic state. This implies that the rock porosity is interconnected due to the extremely high density of sub-vertical fractures in this uppermost part of the crust. This high-permeability upper reservoir is associated with thermal or thermohaline convection through the rock fracture porosity (Figure 4a). The $\sim 2\text{--}3\text{ km}$ transition zone between the two reservoirs probably represents a permeability barrier [Jónsson *et al.*, 2003] (Figure 4a). At greater depths, the recorded seismicity suggests that fluids are over-pressurized due to trapping in a lower-permeability system of unconnected cracks (notably due to hydrothermal sealing of the smaller cracks). At $z > 3\text{ km}$, such a high-enthalpy supercritical fluid could dramatically increase the potential for rock fracturing, particularly through stress-corrosion micro-cracking at the grain scale [Hashida and Takahashi, 2004] (Figure 4a). The higher fluid contents along major faults in the ‘dilatational areas’ (Figure 2), and the eventual localized separation of vapor bubbles from the supercritical fluid, could explain the high seismic loading along these faults and their sensitivity to dynamic stresses, possibly through rectified diffusion processes [Sturtevant *et al.*, 1996] (Figure 4a).

[8] We speculate that transient but dramatic increases in upper crustal permeability could lead to bulk fluid penetration down to the “2–3 km-barrier”, leading to recharge of the lower upper-crustal reservoir (Figure 4b). Such phenomena would accompany large fault-slip events in the ‘dilatational areas’, as illustrated at Kleifarvatn. Over a period of several months, there would be a recharge of the deep brittle crust by higher-density colder fluids from the upper reservoir, associated with a temporary lowering of the hydrostatic/non-hydrostatic pressure boundary (Figure 4b). The mixing of supercritical fluid with downward penetrating colder water, combined with a coeval decrease in pore-fluid pressure, could lead to phase separation and the formation of a highly dynamic superheated

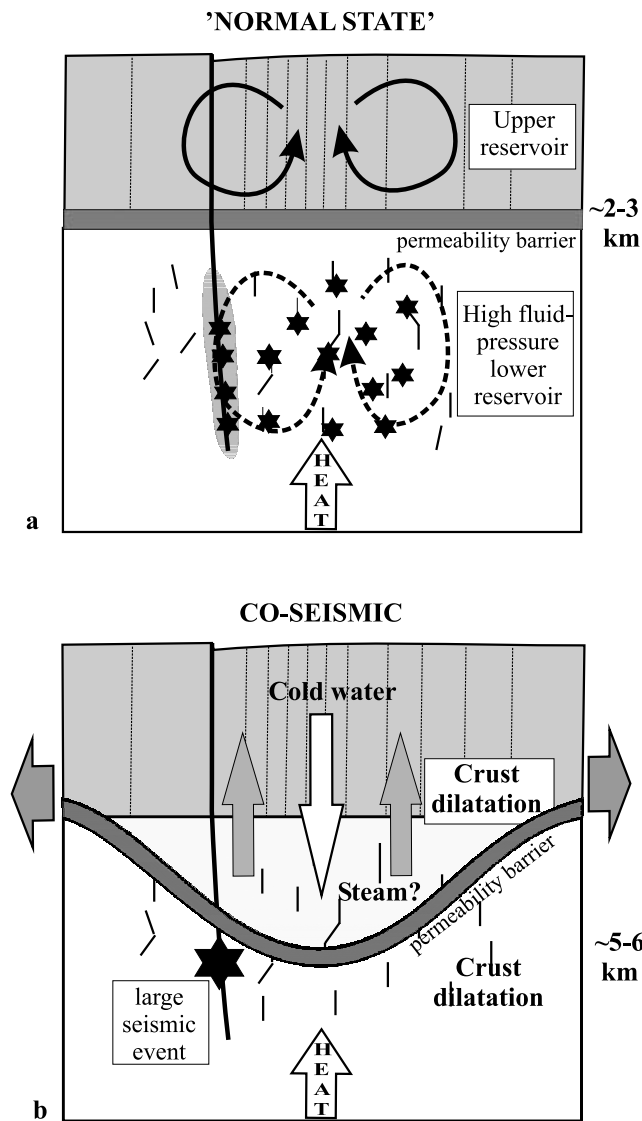


Figure 4. Two-stage conceptual model of fluid behavior at the center of volcano-tectonic segments (NW–SE vertical transects). (a) ‘Normal stage’. Free fluid convection (curved solid lines with arrows) in the hydrostatic upper reservoir, probable “forced convection” (curved dashed lines with arrows) in the seismic (stars indicate events) overpressured reservoir. Major crustal faults are seismically loaded due to high fluid pressure at their roots. (b) ‘Coseismic stage’. Fault displacement (large star: earthquake triggered by a distant event) associated with crustal dilatation at its periphery, increasing bulk crust permeability and allowing mixing between the two reservoirs. This leads to a recharge of water towards the lower reservoir and heat supply to the upper reservoir.

vapor phase [Friedleifsson and Elders, 2005] that would convey heat to the upper reservoir (Figure 4b).

[9] Therefore, we suggest that far-field stresses associated with plate separation could indirectly control the deep percolation of fluids through the seismic cycle in specifically dilated areas of the Icelandic crust. This sheds light on how oceanic water could efficiently cool the upper oceanic crust at slow-spreading ridges.

[10] **Acknowledgments.** We are grateful to Michel Frogneux and Henri Haessler for their involvement in the seismic campaign. We thank Sophie Cravoisier, who was in charge of the maintenance of equipment during the field work and data processing. The Icelandic Meteorological Office is warmly thanked for its logistic support and allowing access to SIL data. An anonymous reviewer and Gillian Foulger are thanked for their thorough reviews. Michael Carpenter post-edited the English style. Study supported by the “région Pays de La Loire”.

References

- Argus, D. F., and R. G. Gordon (1991), No-Net-Rotation model of current plate velocities incorporating plate motion model NUVEL-1, *Geophys. Res. Lett.*, *18*, 2039–2042.
- Árnadóttir, T., W. Jiang, K. L. Feigl, H. Geirsson, and E. Sturkell (2006), Kinematic models of plate boundary deformation in southwest Iceland derived from GPS observations, *J. Geophys. Res.*, *111*, B07402, doi:10.1029/2005JB003907.
- Arnórssson, S. (1995), Geothermal systems in Iceland: Structure and conceptual models—I. High-temperature areas, *Geothermics*, *24*, 561–602.
- Barclay, A. H., D. R. Toomey, and S. C. Solomon (2001), Microearthquake characteristics and crustal V_p/V_s structure at the Mid-Atlantic Ridge, 35°N, *J. Geophys. Res.*, *106*, 2017–2034.
- Byerlee, J. (1990), Friction, overpressure and fault normal compression, *Geophys. Res. Lett.*, *17*, 2109–2112.
- Clifton, A., C. Pagli, J. F. Jónsdóttir, K. Eythorsdóttir, and K. Vogfjöld (2003), Surface effects of triggered fault slip on Reykjanes Peninsula, SW Iceland, *Tectonophysics*, *369*, 145–154.
- Einarsson, P. (1991), Earthquake and present-day tectonism in Iceland, *Tectonophysics*, *189*, 261–279.
- Foulger, G. R., Z. Du, and B. R. Julian (2003), Icelandic-type crust, *Geophys. J. Int.*, *155*, 567–590.
- Friedleifsson, G. O., and W. A. Elders (2005), The Iceland Deep Drilling Project: A search for deep unconventional geothermal resources, *Geothermics*, *34*, 269–285.
- Gudmundsson, A. (1986), Mechanical aspects of postglacial volcanism and tectonics of the Reykjanes Peninsula, southwest Iceland, *J. Geophys. Res.*, *91*, 12,711–12,721.
- Hashida, T., and T. Takahashi (2004), Microcrack formation and fracture characteristics in granite under supercritical water conditions, in *Coupled Thermo-Hydro-Mechanical-Chemical Processes in Geo-Systems: Fundamentals, Modelling, Experiments and Applications*, Elsevier *Geo Eng. Book Ser.*, vol. 2, edited by O. Stephansson et al., pp. 661–665, Elsevier, Amsterdam.
- Hjaltadóttir, S., and K. S. Vogfjörd (2006), Kortlagning sprungna I Fagradalsfjalli a Reykjaneskaga með smaskjalftum, *Rep. 06001*, 35 pp., Greinargerð Verdurstofa Isl., Reykjavík.
- Iyer, H. M. (1992), Seismological detection and delineation of magma chambers: Present status with emphasis on the western USA, in *Volcanic Seismology*, *IAVCEI Proc. Volcanol.*, vol. 3, edited by P. Gasparini, R. Scarpa, and K. Aki, pp. 299–338, Springer, Berlin.
- Jónsson, S., P. Segall, R. Pedersen, and G. Björnsson (2003), Post-earthquake ground movements correlated to pore-pressure transients, *Nature*, *424*, 179–183.
- Julian, B. R., A. Ross, G. R. Foulger, and J. R. Evans (1996), Three-dimensional seismic image of a geothermal reservoir: The Geysers, California, *Geophys. Res. Lett.*, *23*, 685–688.
- Keiding, M., T. Arnadóttir, E. Sturkell, H. Geirsson, and B. Lund (2007), Strain accumulation along an oblique plate boundary: The Reykjanes Peninsula, southwest Iceland, *Geophys. J. Int.*, *172*, 861–872, doi:10.1111/j.1365-246X.2007.03655.x.
- Klein, F. W., P. Einarsson, and M. Wyss (1977), The Reykjanes Peninsula, Iceland, earthquake swarm of September 1972 and its tectonic significance, *J. Geophys. Res.*, *82*, 865–888.
- McCaig, A. M. (1988), Deep fluid circulation in fault zones, *Geology*, *16*, 867–870.
- Mochizuki, M., B. Brandsdóttir, H. Shiobara, G. Gudmundsson, and R. Stefánsson (2000), Detailed distribution of microearthquakes along the northern Reykjanes Ridge, off SW-Iceland, *Geophys. Res. Lett.*, *27*, 1945–1948.
- Stefánsson, R., R. Böðvarsson, R. Slunga, P. Einarsson, S. Jakobsdóttir, H. Bungum, S. Gregersen, J. Havskov, J. Hjelme, and H. Korhonen (1993), Earthquake prediction research in the South Iceland seismic zone and the SIL project, *Bull. Seismol. Soc. Am.*, *83*, 696–716.
- Sturtevant, B., H. Kanamori, and E. E. Brodsky (1996), Seismic triggering by rectified diffusion in geothermal systems, *J. Geophys. Res.*, *101*, 25,269–25,282.
- Tryggvason, A., S. T. Rögnvaldsson, and O. G. Flóvenz (2002), Three-dimensional imaging of the P- and S-wave velocity structure and earthquake locations beneath southwest Iceland, *Geophys. J. Int.*, *151*, 848–866.

- Weir, N. R. W., R. S. White, B. Brandsdottir, P. Einarsson, H. Shimamura, H. Shiobara, and the RISE fieldwork team (2001), Crustal structure of the northern Reykjanes Ridge and Reykjanes Peninsula, southwest Iceland, *J. Geophys. Res.*, *106*, 6347–6368.
- Zencher, F., M. Bonafede, and R. Stefansson (2006), Near lithostatic pore pressure at seismogenic depths: A thermo-poro-elastic model, *Geophys. J. Int.*, *166*, 1318–1334.
- Zhang, H., and C. Thurber (2003), Double-difference tomography: The method and its application to the Hayward Fault, California, *Bull. Seismol. Soc. Am.*, *93*, 1875–1889.
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