Coseismic stress changes and crustal deformation on the Reykjanes Peninsula due to triggered earthquakes on 17 June 2000

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[1] A large ($M_w = 6.5$) earthquake struck the South Iceland Seismic Zone (SISZ) on 17 June 2000. The 17 June main shock triggered increased seismicity over a large area and significant slip on at least three distinct faults on the Reykjanes Peninsula, up to 87 km to the west of the event. A second large ($M_w = 6.4$) earthquake in the SISZ occurred on 21 June 2000, about 17 km west of the 17 June main shock. This event does not appear to have triggered as much activity on the Reykjanes Peninsula as the 17 June main shock, although the epicenter was closer. Crustal deformation signals due to the June 2000 earthquakes on the Reykjanes Peninsula were observed with campaign and continuous GPS and synthetic aperture radar interferometry, with the largest coseismic deformation signal near lake Kleifarvatn. We model the faults using three uniform slip rectangular dislocations in an elastic half-space. Best fit uniform slip models consistent with seismic and geodetic data indicate that all three faults trend N-S and the motion on them was primarily right-lateral strike slip. Our study suggests that the event near Kleifarvatn had a significantly larger moment than seismic estimates, indicating a component of aseismic slip on the fault lasting no more than several hours. Static Coulomb failure stress change calculations indicate that the event at Kleifarvatn increased the Coulomb stress at the hypocenter of the Núpschildarháls event by 0.1–0.2 MPa as well as loading the Hvalhúnukur fault.

INDEX TERMS: 1242 Geodesy and Gravity: Seismic deformations (7205); 7209 Seismology: Earthquake dynamics and mechanics; 7230 Seismology: Seismicity and seismotectonics; 7215 Seismology: Earthquake parameters; 8168 Tectonophysics: Stresses—general

KEYWORDS: aseismic slip, triggered earthquakes, Reykjanes Peninsula


1. Introduction

[2] The Reykjanes Peninsula (RP) oblique rift zone is the onshore continuation of the Reykjanes Ridge. The Reykjanes Peninsula rift zone branches into the Western Volcanic Zone (WWZ) and the South Iceland Seismic Zone (SISZ) at the Hengill (He) triple junction (Figure 1). The SISZ is a left-lateral E-W transform zone, where the motion is accommodated by right-lateral motion on many short N-S strike-slip faults [Einarsson and Eiríksson, 1982; Einarsson et al., 1981]. Most of the plate spreading across central Iceland is currently occurring across the Eastern Volcanic Zone (EVZ), at the eastern side of the SISZ, with less than 15% of the motion taken up by the WWZ [Sigmundsson et al., 1995]. The NUVEL-1A plate motion model predicts spreading of Eurasia relative to North America at a rate of $18.9 \pm 0.5$ mm/yr in a direction of N102.7 ± 1.1°E across the Reykjanes Peninsula [DeMets et al., 1990, 1994]. All the volcanic systems on the Reykjanes Peninsula have produced lavas in historical times, the last known eruption took place in 1240 A.D. The dominant tectonic features are NE trending eruptive fissures, tension fractures and normal faults, as well as N-S surface faults with similar features as mapped in the SISZ [e.g., Einarsson and Sæmundsson, 1987; Erlendsson and Einarsson, 1996]. There are several geothermal areas on the Reykjanes Peninsula, and in one of them, the Svartsengi high temperature field, energy is harnessed for house heating purposes, with electricity generated as a by-product. The heat mining at Svartsengi produces...
subsidence that is detectable by interferometric synthetic radar (InSAR) and levelling [Vadon and Sigmundsson, 1997].

[3] GPS measurements from 1986 to 1992 confirm left-lateral shear strain accumulation of 0.2 μstrain/yr across the peninsula [Sturkell et al., 1994]. The deformation measured with GPS from 1993 to 1998, caused by plate spreading, has been explained with a simple screw dislocation model that is slipping at a rate of ~16.5 mm/yr below ~6.5 km depth [Hreinsdóttir et al., 2001].

[4] The seismicity on the Reykjanes Peninsula is concentrated in a narrow region trending about N80°E, with most hypocenters centered at 1–5 km depth. The seismicity is episodic, with periods of high activity in 1929–1935, and 1967–1973. The earthquakes on the western part of the peninsula are mostly normal faulting events that occur in swarms. Further east strike-slip faulting is more common, but the rate of seismicity in the area from Kleifarvatn to Ólfus (the southern part of the Hengill triple junction) following the 1967–1973 episode was low [Einarsson, 1991]. Increased seismicity at the Hengill triple junction from 1994 to 1998 has been associated with uplift due to magmatic activity in the crust [Sigmundsson et al., 1997; Feigl et al., 2000]. These studies have shown that the observed surface deformation can be explained by a single point source of inflation at about 7 km depth with maximum uplift rate of about 2 cm/yr during the period. The seismic episode at Hengill culminated with two $M_w \sim 5$ earthquakes, on 4 June and 13 November 1998 [Árnadóttir et al., 1999; Feigl et al., 2000].

[5] The largest known earthquake on the Reykjanes Peninsula, of surface wave magnitude $M_S \sim 6.2$, occurred on 23 July 1929 (International Seismological Centre (ISC) (2004), online bulletin, available at http://www.isc.ac.uk/Bull). This event and a $M_S \sim 6.0$ event on 5 December 1968 (International Seismological Centre (ISC) (2004), online bulletin, available at http://www.isc.ac.uk/Bull), probably occurred on the Hvalhúkur fault [Erlendsson and Einarsson, 1996]. The Hvalhúkur fault is the longest mapped N-S strike-slip fault on the Reykjanes Peninsula and lies close to the capital of Reykjavík (labeled in Figure 1), posing a major seismic hazard in the area.


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**Figure 1.** Map showing the main tectonic features of the study area. The Reykjanes Peninsula (RP), the Hengill triple junction (He), the Western Volcanic Zone (WVZ), the South Iceland Seismic Zone (SISZ), and the Eastern Volcanic Zone (EVZ) are indicated. The light shaded areas are individual fissure swarms with associated central volcanoes. Mapped surface faults of Holocene age are shown with black lines [Einarsson and Sæmundsson, 1987]. The epicenter locations of earthquakes on June 2000 are shown with black stars. The 17 June main shock is labeled J17. The Reykjanes Peninsula events are on the Hvalhúkur fault (H), near lake Kleifarvatn (K), and Núpsvöllur (N). Location of the 21 June 2000 main shock is shown with a white star labeled J21. Available focal mechanisms for the main events in June 2000 are shown with a lower hemisphere projection (USGS and K. Vogfjörd personal communication, 2003). The location of Reykjavík, the capital of Iceland, is noted. The geothermal area at Svartsengi (Sv) is noted by a hexagon. Inset shows a simplified map of the plate boundary across Iceland, with the location of main part of figure indicated by the black rectangle. The black arrows show the direction of the NUVEL-1A plate motion of North America and Eurasian plates.
was located at 63.975°N, 20.370°W and 6.3 km depth, with a moment magnitude of $M_w = 6.5$ (Figure 1). Seismicity increased over a large area in SW Iceland following the June 17 main shock. Three moderate size ($M_w \approx 5$) earthquakes occurred on the Reykjanes Peninsula, progressively shallower and further west, within 5 min of the main shock [Clifton et al., 2003]. The first event occurred near the northern end of the Hvalhnu´kur fault, at 1541:07 UT, at 63.951°N, 21.689°W and about 9 km depth (labeled H in Figure 1). The second event occurred near Kleifarvatn at 1541:11 UT on 17 June 2000. The hypocenter location is 63.937°N, 21.940°W and about 4 km depth (labeled K in Figure 1). It is difficult to determine the location, moment, and fault plane solution for this event as the closest seismic stations were saturated and the seismic signal is also hidden in the coda of the 17 June main shock and the first triggered event [see Clifton et al., 2003, Figure 2]. The third event located near Núpshlı´darha´ls occurred at 1545:27 UT on 17 June 2000. The hypocenter location is 63.902°N, 22.124°W and about 3 km depth (labeled N in Figure 1). The timing of the events indicates that the event on the Hvalhnu´kur fault and Kleifarvatn were triggered by surface waves from the 17 June main shock and the first triggered event [see Clifton et al., 2003, Figure 2]. The third event located near Núpshlı´darha´ls occurred at 1545:27 UT on 17 June 2000. The hypocenter location is 63.902°N, 22.124°W and about 3 km depth (labeled N in Figure 1). The focal mechanisms for the Hvalhnu´kur and Núpshlı´darha´ls events are shown in Figure 1 (K. Vogfjörd, personal communication, 2003).

2. Geodetic Data

2.1. GPS Data

The Reykjanes Peninsula GPS network was surveyed in 1986, 1989, 1992, 1993, 1998, and 2000 [Foulger et al., 1987; Sigmundsson et al., 1995; Starkell et al., 1994; Hreinsdóttir et al., 2001; this study]. In the present paper we use data collected in 1998 and 2000 (Figure 2).

Surface faulting observed for the three events indicate rupture on N–S trending faults, with severe disruption found at the eastern shore of lake Kleifarvatn [Clifton et al., 2003]. The level of lake Kleifarvatn dropped by about 4 m during the first 18 months following the June 2000 earthquakes, as the water drained into fresh surface cracks in the lake bottom [Clifton et al., 2003].

The June 2000 earthquake sequence was recorded by the SIL digital seismic network [Bödvarsson et al., 1999; R. Stefánsson et al., The two large earthquakes in the South Iceland Seismic Zone on 17 and 21 June 2000, available at http://hraun.vedur.is/ja/skyrslur/June17and21_2000, last modified 26 July 2000]. Crustal deformation signals were recorded by the continuous GPS (CGPS) network in Iceland [Arnadóttir et al., 2000; Geirsson, 2003], campaign GPS measurements [Arnadóttir et al., 2001] and the present study, and interferometric analysis of synthetic aperture radar (InSAR) [Pedersen et al., 2001, 2003; Pagli et al., 2003].

In this paper we model the surface deformation measured by GPS and InSAR by three dislocation sources on the Reykjanes Peninsula representing the three events on 17 June 2000 (labeled H, K, and N in Figure 1). Furthermore, we calculate the changes in the static Coulomb failure stress from these dislocation models.

Figure 2. GPS stations on the Reykjanes Peninsula observed in 1998 (open squares) and 2000 (open triangles). Continuous GPS (CGPS) stations are indicated by black squares (VOGS and REYK are labeled), and the 17 June 2000 epicentral locations by black stars. The white star marks the epicentral location of $M_w = 5.1$ event that occurred in Ólfus on 13 November 1998.
The July 2000 GPS survey was a continuation of a survey of the SISZ following the June 2000 earthquake sequence [Arnadóttir et al., 2001]. It was a collaboration between the Icelandic Meteorological Office (IMO) and SIUI. The western most part of the network was not resurveyed. The 1998 and 2000 surveys were performed using dual-frequency GPS receivers, collecting data every 15 s, occupying each station for about 24 hours.

We analyze the GPS data collected in 2000 using the Bernese V4.2 software [Hugentobler et al., 2001] and precise orbital information from the Center for Orbit Determination in Europe (CODE). Data from the IGS station in Reykjavík (REYK) and several CGPS stations in SW Iceland were included in the analysis. The network of CGPS stations (ISGPS) operated by the IMO was initiated in Iceland in March 1999 [Arnadóttir et al., 2000; Geirsson, 2003]. The first CGPS station that was installed in the ISGPS network is VOGS, which is located in the study area. We use coseismic offsets observed at VOGS in the modeling. A detailed description of the CGPS network and data analysis is given by Geirsson [2003]. Figure 3 shows the horizontal and vertical location of VOGS relative to REYK as a function of time, from of 1 May to 31 August 2000, are shown. Each dot represents a 24 hour solution using CODE final orbits. Motion in east, north, and up directions is defined to be positive and is given in units of millimeters. The error bars indicate 1σ confidence limits. The vertical lines mark the times of the 17 June and 21 June main shocks. The east and north components of motion show a clear jump on 17 June of about 19 mm and about −10 mm, respectively.

Figure 3. Time series from the continuous GPS station VOGS, located on the SE coast of the Reykjanes Peninsula (see Figure 2). The horizontal and vertical motions of VOGS relative to REYK as a function of time, from of 1 May to 31 August 2000, are shown. Each dot represents a 24 hour solution using CODE final orbits. Motion in east, north, and up directions is defined to be positive and is given in units of millimeters. The error bars indicate 1σ confidence limits. The vertical lines mark the times of the 17 June and 21 June main shocks. The east and north components of motion show a clear jump on 17 June of about 19 mm and about −10 mm, respectively.
smaller signal was observed on 21 June 2000, when a second large \( (M_w = 6.4) \) earthquake occurred in the SISZ, although the epicenter was closer to VOGS. The predicted displacements at VOGS, assuming uniform slip models for the SISZ main shocks [Árnadóttir et al., 2000], are 4 mm east and 3 mm north (17 June) and 5 mm east and north (21 June). The observed coseismic offset on 17 June at VOGS is therefore much larger than expected from modeling of the 17 June main shock, and we assume that it was primarily caused by motion on faults on the Reykjanes Peninsula. The total coseismic displacement of REYK due to the 17 June and 21 June main shocks has been estimated as 2.7 ± 0.5 mm west and 8.6 ± 0.5 mm north, using the GAMIT/GLOBK software (W. Jiang, personal communication, 2004). Motion at REYK due to the 17 June main shock, estimated from a uniform slip model, is 1 mm west and 2 mm north, and for the 21 June main shock 3 mm west and 3 mm north; thus a total of 4 mm west and 5 mm north for the two main shocks [Árnadóttir et al., 2001]. The coseismic motion of REYK due to the 17 June earthquakes on the Reykjanes Peninsula is therefore no more than 1 mm toward east and 4 mm toward north, which is at the level of uncertainty of the GPS campaign data. We therefore assume that the 17 June earthquakes on the Reykjanes Peninsula did not cause significant coseismic displacement at REYK.

[13] The station displacements from 1998 to 2000 (black arrows with 95% confidence ellipses) relative to the IGS station REYK (indicated with a black square) are shown in Figure 4. The largest horizontal station displacement we observe is about 0.1 m at a station on the southern shore of lake Kleifarvatn.

[14] In this study we are concerned with the coseismic signals due to the 17 June 2000 earthquakes on the Reykjanes Peninsula. Although the 2000 GPS campaign measurements did not cover the whole Reykjanes Peninsula, they span the area affected by the three largest earthquakes that occurred on 17 June on the peninsula. The GPS measurements in this study also include deformation caused by an earthquake swarm (maximum magnitudes \( M_w = 5.1 \)) in Ölfus on 13 November 1998 (shown with white star in Figure 4) [Árnadóttir et al., 1999]. We use GPS data collected in August and November 1998 to correct for a small deformation signal (maximum about 10 mm at the stations closest to the epicenter) due to this swarm [Hreinsdóttir, 1999]. We also subtract a small component of motion due to the two SISZ main shocks in June 2000, based on a model by Árnadóttir et al. [2001] (shown with white arrows in Figure 4). The station displacements caused by almost 3 years of plate motion across the Reykjanes Peninsula are estimated from 1993–1998 station velocities, assuming they represent interseismic motion during the period 1998–2000 [Hreinsdóttir et al., 2001]. We calculate the June 2000 coseismic deformation on the Reykjanes Peninsula by subtracting interseismic motion, and coseismic displacements for the 17 June main shock and 13 November 1998 earthquake, from the 1998–2000 station displacements (Figure 4). Data from stations shown with black triangles in Figure 4 are not included in the inversion, as the stations are in the far field and contain local signals not related to the 17 June 2000 earthquakes on the Reykjanes Peninsula. Because of the corrections we make to obtain the coseismic GPS displacements, we assign an uncertainty of 5 mm in the horizontal and 10 mm in the vertical...
components of the GPS displacements in the inversion. The resulting coseismic horizontal displacements are shown with black arrows in Figure 5b.

2.2. InSAR Data

[15] Thirteen interferograms from the Reykjanes Peninsula, from ERS-1 and 2, track 367, frame 2313, spanning the period 1993–2000 were analyzed by Pagli [2002] and Pagli et al. [2003]. We use the analysis of Pagli et al. [2003] in this study. The interferograms are calculated using the standard two-pass method of the PRISME/DIAPARSON software [Centre National d’Etudes Spatiales, 1997]. Topographic contributions are removed with a digital elevation model (DEM). The DEM used has a pixel size of roughly 90 by 90 m, and a vertical precision of about 30 m. Orbital modeling and filtering were done following the method described by Feigl et al. [2000]. Further details of the image processing of the InSAR analysis is given by Pagli et al.
The parameters are estimated for all three sources: Kleifarvatn, Núpsvölfur, and Hvalhnukur events.

3. Surface Deformation and Dislocation Modeling

[2003]. They found four interferograms that cover the coseismic period, with one superior quality image pair from ERS-2, with an altitude of ambiguity \( h_a \) [Massonnet and Feigl, 1998] of 2654 m, spanning 2 October 1999 to 16 September 2000 (Figure 5a). The other three coseismic images span longer time periods, have lower \( h_a \) (about 100 m) and large incoherent areas [see Pagli et al., 2003, Table 1 and Figure 2]. We use the best coseismic interferogram in our analysis, but expand the area studied by Table 1 and Figure 2]. We use the best coseismic interferogram, but expand the area studied by Pagli et al. [2003] toward east, to include the deformation signal due to motion on the Hvalhnukur fault (Figure 5a). The largest coseismic signal is near lake Kleifarvatn, with smaller signals near Núpsvölfur and Hvalhnukur. An interferogram spanning the period from 19 June 2000 to 24 July 2000, shows no deformation in the area, indicating that the deformation occurred before 19 June 2000 [Pagli et al., 2003]. The combined effect of interseismic motion during the period spanned by the interferogram, and coseismic deformation due to the 17 June and 21 June 2000 main shocks, is small in the InSAR data, and we do not attempt to correct for these. Before modeling, the InSAR data are unwrapped using a Markov random field regularization and a simulated annealing optimization algorithm [Gudmundsson et al., 2002]. The number of data was reduced by a two-dimensional quantization algorithm known as quadtree partitioning as described by Jónsson et al. [2002]. The unit vector in the direction from ground to the satellite is assumed to be constant (0.408, −0.112, 0.906) for the east, north, and vertical components, respectively. The InSAR data are therefore most sensitive to vertical deformation. The resultant data used for the modeling are shown in Figure 5b where the displacements in the direction to the SAR satellite (range change) are represented by the color scale. Here, red is range decrease, and blue is range increase, indicating ground motion toward and away from the satellite, respectively. The maximum range increase is about 0.1 m east of lake Kleifarvatn. The horizontal coseismic GPS displacements are shown with black arrows in Figure 5b.

3. Surface Deformation and Dislocation Modeling

[16] We perform a joint inversion of the GPS and InSAR data, estimating the best fit model parameters using a simulated annealing nonlinear search algorithm, followed by a derivative-based method as described by Jónsson et al. [2002]. We model the three earthquakes on the Reykjanes Peninsula as rectangular dislocations with uniform slip in an elastic half-space [Okada, 1985], estimating nine parameters for each dislocation (see Table 1) and three constants to correct for an arbitrary shift (due to uncertainty in assigning zero displacement) and tilt in the unwrapped InSAR data. The optimal model parameters are found by minimizing the weighted-residual sum of squares, \( \mathbf{r}^T \Sigma^{-1} \mathbf{r} \). Here \( \mathbf{r} \) is the difference between the observed and modeled data, and \( \Sigma \) is the data covariance matrix. We assign a standard deviation of 10 mm for each quadtree point, 5 mm and 10 mm for the horizontal and vertical components of the GPS data respectively, and a diagonal covariance matrix for the observations in the inversion. The InSAR data set contains 245 data points, and there are 69 GPS data from 23 stations. The total number of data is \( n = 314 \). The relative weight of the data sets in the inversion is determined by the number of data and their assigned uncertainties. Here the relative weight is 69%, 13%, 13% and 5% for the InSAR, the two horizontal, and the vertical GPS data, respectively.

[17] The epicenters of the three earthquakes are about 10 km apart, and they occurred within a few seconds of each other. The coseismic interferogram shows clear signals for the Kleifarvatn and Núpsvölfur events, and a more subtle signal for the Hvalhnukur event. The GPS network is not dense enough to allow us to clearly separate the three events in space and time. In addition, the events are rather small, causing small deformation signals. In particular the signal for the Hvalhnukur event is very small since the hypocenter was located at 9 km depth. We therefore fix the model parameters for the Hvalhnukur event by the earthquake hypocenter location and focal mechanism (K. Vogfjörd, personal communication, 2003) and locations of surface faulting (A. Clifton, personal communication, 2003).

[18] The best fit model parameters are listed in Table 1 (model K, N and H). The model for the event at Kleifarvatn is a 6 km long fault, dipping 77°E, striking N4°E, extending from the surface down to about 5 km depth. The model has 0.7 m of right-lateral strike-slip and 0.1 m of normal motion. The model for the event at Núpsvölfur is a 3 km long fault, dipping about 70°E, striking N4°W, extending from the surface down to about 4 km depth. The model has 0.25 m of right-lateral strike-slip motion and a small component (0.05 m) of thrust. The preferred fault model has a mean square error (MSE = \( \mathbf{r}^T \Sigma^{-1} \mathbf{r} / (n - m) \)) of 0.85 (with the number of model parameters \( m = 22 \)) and misfit equal to \( \sqrt{\text{MSE}} = 0.92 \). This model explains about 95% of the ground deformation signal. If we only estimate model parameters for the Kleifarvatn event (\( m = 9 + 3 = 12 \)), we find similar fault parameters as for our preferred model for the Kleifarvatn event, with larger slip and moment (model K1 in Table 1). However, model K1 only fits about 89% of the signal, with a MSE of 1.7 and misfit equal to 1.3. If we estimate parameters for both the Kleifarvatn and Núpsvölfur-Hvalhnukur faults (\( m = 21 \)) the misfit is reduced to 0.93 (models K and N in Table 1). A best fit model with the Kleifarvatn

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<th>Table 1. Best Fit Dislocation Parameters for 17 June 2000 Events on the Reykjanes Peninsulaa</th>
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a The parameters are the fault length along strike, depth to the upper edge, width along dip, dip toward east, strike east of north, location of the upper edge of the southern end of the fault, right-lateral strike slip (ss), normal dip slip (ds), and geodetic moment (\( M_{00} \)). In model K1 we estimate parameters only for the Kleifarvatn event. In models K, N, and H the parameters are estimated for all three sources: Kleifarvatn, Núpsvölfur-Hvalhnukur, and Hvalhnukur events.

b Parameter is fixed before the inversion.
fault and an E-W oriented fault for Núpshlidarháls, similar to the preferred model by Pagli et al. [2003], has a misfit equal to 1.0 and does not fit the northward motion observed at the GPS station closest to the epicenter. We feel it is important to include the Hvalhnu´kur fault in our modeling for the Coulomb failure stress calculation, discussed in section 4. As explained above, we fix the model parameters for the Hvalhnu´kur fault and estimate the fault slip (model H in Table 1). Introducing slip on the Hvalhnu´kur fault model lowers the misfit for the two fault models (models K and N in Table 1) from 0.93 to 0.92.

In general, the model parameters are less well constrained by the GPS data than the InSAR data because of sparser spatial coverage. The best fit dislocation sources found when using only GPS data, are narrower and have higher slip than our preferred models. The source parameters for the Kleifarvatn event are better constrained by the InSAR and GPS data than the parameters for the Núpshlidarháls event. In particular, the width, strike, and dip of the Núpshlidarháls model are poorly constrained by the geodetic data.

Figure 6 shows the observed coseismic horizontal displacements for the GPS data (black arrows) and calculated displacements (white arrows) for our preferred models (gray squares). It also shows well-located earthquakes (horizontal error ≤2.5 km, vertical error ≤5.0 km) recorded by the SIL seismic system from 17 June to 31 December 2000. Hypocenter location error may explain why our fault model for the Kleifarvatn event is south of the epicenter determined from seismic data. Surface faulting was most pronounced along the SE shore of lake Kleifarvatn [Clifton et al., 2003], suggesting that large slip occurred in that area, which agrees well with our model. The seismicity is fairly diffuse and does not reveal a clear correlation between the locations of large events, the aftershocks, and our fault models.

Figure 7a shows the predicted motion for the InSAR data for our preferred model. Figure 7b shows the residual (r) range change and horizontal GPS displacements (black arrows). The uniform easterly motion of the GPS residual vectors in the eastern part of the area are probably due to residual plate motion. The residual InSAR signal is small (within ±5 mm for most of the area), and no clear systematic pattern is evident in the vicinity of the modeled faults. This suggest that the residual signal is for the most part due to processes not related to the earthquakes.

The geodetic moment is calculated from the relation $M_0 = \mu Au$, where $\mu$ is the shear modulus (we use $\mu = 3 \times 10^{10} \text{ N/m}^2$), $A$ is the area, and $u$ is the total slip on the fault (see Table 1). The geodetic moments for our models are well constrained and agree reasonably well with moment estimates based only on InSAR data for the Kleifarvatn ($M_0 = 6.2 \times 10^{17} \text{ N m}$) and Núpshlidarháls ($M_0 = 1.0 \times 10^{17} \text{ N m}$).
events [Pagli et al., 2003], even though the event near the Hvalhnu´kur fault (termed Bla ´fjo¨ll in their paper) is not included in Pagli et al.’s modeling. Our fault locations and geometries are also similar to the results of Pagli et al. [2003]. The largest difference is that our model for the fault near Núpslidharháls has a N–S orientation and has primarily right-lateral strike-slip motion, rather than an E–W orientation with a significant component of reverse slip, as the model preferred by Pagli et al. [2003]. The main reason why we prefer a N–S fault, rather than an E–W, is that the latter model does not fit the displacements at the GPS station that is closest to the hypocenter of the Núpslidharháls event. The inherent uncertainty in the data and nonuniqueness of the inversion does not allow us to rule out an E–W model, but a N–S fault model is in better agreement with our GPS data, the surface ruptures observed after the June 2000 earthquakes [Clifton et al., 2003], previously mapped surface faults, and the general tectonics in the area; hence we prefer the N–S model.

4. Static Coulomb Failure Stress

[23] Slip on a fault will modify the stress field in the surrounding rock. A fault that is close to failure may slip when the applied stress increment, defined as the change in Coulomb failure stress (ΔCFS) [e.g., Harris, 1998; Beeler et al., 2000; Cocco and Rice, 2002],

\[
\Delta \text{CFS} = \Delta \tau_s + \mu \left( \Delta \sigma_n - \frac{B}{3} \Delta \sigma_k \right),
\]

exceeds a threshold value. Here \(\Delta \tau_s\) is the change in shear stress resolved in the slip direction of a fault that may fail in a subsequent earthquake; \(\mu\) is the coefficient of friction, \(\Delta \sigma_n\) is the change in normal stress due to the first earthquake, perpendicular to the subsequent earthquake fault plane (positive for extension); \(\sigma_{kk}\) is the volumetric stress; and \(B\) is the Skempton coefficient of the rock-fluid mixture [e.g., Harris, 1998, and references therein]. A positive ΔCFS implies an increase in Coulomb failure stress, indicating that the first earthquake brought the second fault closer to failure.

[24] The Kleifarvatn event occurred about 4 s after the Hvalhnu´kur event, whereas the Núpslidharháls event occurred about 4 min after the Kleifarvatn event. We want to examine the role of static Coulomb failure stress changes in triggering of these events. To do this, we calculate the change in static Coulomb failure stress at 3 km depth, from equation (1), for the 17 June 2000 earthquakes, using a nonuniform slip fault model for the main shock [Pedersen et al., 2003], and the uniform slip models for the Hvalhnu´kur and Kleifarvatn events on the Reykjanes Peninsula found in the present study. We assume an elastic half-space, and the receiver faults are N–S, right-lateral strike-slip faults. We calculate the Coulomb failure stress change on vertical and 70° east dipping faults (Figures 8a and 8b, respectively). The effect of the dip is largest in the area close to the fault that slips but decreases rapidly with distance. We use a Poisson’s ratio of \(\nu = 0.28\), a shear modulus of 30 GPa, a coefficient of friction \(\mu = 0.75\), and a Skempton’s coefficient \(B = 0.47\). The 17 June main shock in the SISZ caused a negligible (less than 0.01 MPa) increase in the static Coulomb failure stress on the Reykjanes Peninsula [Árnadóttir et al., 2003]. We find that the 17 June main shock and Hvalhnu´kur event increased the CFS at Kleifarvatn hypocenter by about 0.02 MPa. The Kleifarvatn event had a much bigger effect, increasing the CFS at Núpslidharháls and on the Hvalhnu´kur fault by 0.1–0.2 MPa on both faults, irrespective of the dip (Figures 8a and 8b). Thus the static stress calculations indicate a very small increase in the static Coulomb failure stress due to the 17 June main shock and first event (on the Hvalhnu´kur fault) on 17 June on the Reykjanes Peninsula. This and the timing of the events supports the view that the first two events (Hvalhnu´kur and Kleifarvatn) were dynamically triggered by surface waves from the 17 June main shock (K. Vogfjord and M. Belardinelli, personal communication, 2004).

5. Discussion and Conclusions

[25] Repeated GPS surveys and interferometric radar observations were used to estimate coseismic displacements on the Reykjanes Peninsula due to three earthquakes on 17 June 2000. Simple uniform slip dislocation models for the events provide a good fit to the surface displacements, and indicate that the earthquakes ruptured N–S oriented...
faults, with primarily right-lateral strike-slip motion. We find the largest deformation signal at lake Kleifarvatn, indicating that this event had the largest moment of the three. Using the formula $M_w = \left( \frac{2}{3} \times \log M_0 \right) - 6.03$, our moment magnitude estimate for the events is $M_w = 5.5, 5.9, \text{ and } 5.3$ for the Hvalhnu´kur, Kleifarvatn, and Núpshlı´darha´ls events, respectively. The amplitudes of the seismic signals from the three events, recorded at distant stations are comparable, suggesting that they had similar seismic moments. The estimated moment magnitudes for the Hvalhnu´kur and Núpshlı´darha´ls events are $M_w \approx 5$ (K. Vogfjörd, personal communication, 2004). If we assume that the seismic moment for the Kleifarvatn event is $M_w = 5$, then our moment estimate for the Kleifarvatn event is an order of magnitude larger than the seismic estimates. Our estimates should be considered maximum moment magnitude values as the geodetic data span much longer time intervals than the seismic data. A model for a $M_w = 5$ earthquake, with a fault area of the same dimensions as our preferred model for the Kleifarvatn event, would have a slip of about 35 mm as compared to 710 mm for our preferred model. The Coulomb failure stress increase caused by such an earthquake is only about 0.02 MPa at the Núpshlı´darha´ls hypocenter. One could therefore argue that most of the aseismic slip must have occurred within the 4 min between the events in order for the static Coulomb failure stress change due to the Kleifarvatn event to be large enough to contribute significantly to the failure of the Núpshlı´darha´ls fault.

Aftershocks from 17 June to 31 December 2000 (crosses in Figure 6) do not outline clear failure planes for the three events. In contrast a clear main shock–aftershock sequence started with an $M_w = 5.1$ event on 23 August 2003, about 2 km east of the Núpshlı´darha´ls event. The aftershock sequence outlined a clear N–S trending 5 km long rupture, with a NE trending splay fault toward Kleifarvatn. The Núpshlı´darha´ls event increased the CFS at the 23 August 2003 main shock hypocenter by 0.7 MPa. The lack of aftershocks for the 17 June 2000 event near Kleifarvatn indicates that the seismic part of the Kleifarvatn event was smaller than the 23 August earthquake, and a large part of the moment captured by the geodetic data was released by aseismic slip.

Coulomb failure stress calculations indicate a very small increase in the static Coulomb failure stress both due to the 17 June main shock and the event on the Hvalhnu´kur fault at the hypocenter location of the second triggered event. This supports the view that the first two events (Hvalhnu´kur and Kleifarvatn) were dynamically triggered. Our calculations also indicate that the Kleifarvatn event caused significant stress increase at the Núpshlı´darha´ls hypocenter provided that the aseismic slip took place during the four minutes between the events. The Kleifarvatn event also

Figure 8. Static coseismic Coulomb failure stress changes (ΔCFS, in MPa) due to the 17 June main shock and first two triggered earthquakes (H and K) on the Reykjanes Peninsula, calculated at 3 km depth for N–S faults with right-lateral strike-slip motion. (a) Vertical receiver faults and (b) receiver faults dipping 70° east. The contours surround areas where the CFS increased by more than 0.01 MPa (0.1 bar). The white star shows the location of the third event (N) on the Reykjanes Peninsula. Black crosses mark the same earthquake locations as shown in Figure 6. The coastline and the locations of the K and H fault models are shown with white lines.
appears to have loaded the Hvalhnumur fault. This suggests
that both dynamic and static stress changes played a role in
triggering of the 17 June events on the Reykjanes Peninsula.

[25] Our results support the notion that the present
seismicity of the eastern part of the Reykjanes Peninsula
oblique rift is primarily caused by slip on faults arranged
transversely to the plate boundary [Hreinsdottir et al., 2001;
Erlendsson and Einarsson, 1996]. Thus the general left-
lateral motion along the plate boundary is taken up by
rotation of fault blocks and right-lateral motion on the
faults. This “bookshelf type” tectonic model is similar to
the one proposed for the South Iceland Seismic Zone farther
east. We document here a discrepancy between observed
deforation during the Kleifarvatn event and the estimated
seismic magnitude of the earthquake. This event thus falls
into the category of “slow earthquakes.” Slow earthquakes
have been identiﬁed in different tectonic environments, at
convergent plate boundaries in Japan, Cascadia subduction
zone and Mexico [e.g., Sacks and Linde, 1981; Heki et al.,
1997; Dragert et al., 2001; Lowry et al., 2001], at conser-
vative plate boundaries in California [Linde et al., 1996],
and an unstable volcano flank [Cervelli et al., 2002]. They
are characterized by slow slip and occur on a time scale of
minutes to months. They are accompanied by a highly
variable amount of seismic waves. The aseismic slip fol-
lowing the 1994 earthquake at the Japan trench reported by
Heki et al. [1997] was of comparable magnitude as the
earthquake itself. The largest earthquakes that could be
linked with the slip event on the south ﬂank of Kilauea in
2000 [Cervelli et al., 2002] was of magnitude 2.7. The slip
event, on the other hand, had the moment of a magnitude
5.7 event. Some of the reported slow events were not
associated with any seismically detectable signal. In
the case of Kleifarvatn the seismically detected signal corre-
sponds to an order of magnitude smaller event than the one
detected by geodesy. This is similar to the discrepancy
observed on the San Andreas Fault in central California in
1992 [Linde et al., 1996]. The time source function of the
slow Kleifarvatn event cannot be directly constrained by our
data beyond stating that it was signiﬁcantly slower than a
normal seismic event and that it lasted no more than a few
hours based on data from the CGPS station at VOGS.

[29] The main conclusions of our study may be summa-
rized as follows:

[10] 1. The three seismic events (Mw ≥ 5) that were
triggered along the Reykjanes Peninsula oblique rift on
17 June 2000 were associated with right-lateral strike-slip
faulting on N-S faults, arranged transversely to the
plate boundary, lending support to the model of bookshelf
faulting for this part of the plate boundary at present.

that the ﬁrst two events (Hvalhnumur and Kleifarvatn) were
most likely dynamically triggered by the Mw = 6.5 earth-
quake that occurred in the South Iceland Seismic Zone at
a distance of about 64 km and 78 km, respectively. Failure of
the westernmost fault, near Nupsghildharrs, on the other
hand, may have been promoted by static stress changes
caused by Kleifarvatn event.

[12] 3. The Hvalhnumur and Nupsghildharrs events show
good agreement between seismic moment and geodetically
determined moment. For the Kleifarvatn event, on the other
hand, the measured deformation is up to an order of
magnitude larger than can be explained by the seismic
moment. This event may therefore be termed a “slow
earthquake” with a source time of the order of minutes to
hours. The event falls within the range of previously
documented slow earthquakes, both with regard to timescale
of slip and the discrepancy between seismic and geodetic
moment. Previous studies have suggested several plausible
explanations for the aseismic behavior of the Kleifarvatn
event, including the draining of the lake Kleifarvatn [Clifton
et al., 2003] and the possibility of lower frictional values
due to hydrothermal alteration in the local geothermal area
[Pagli et al., 2003]. We feel that neither of these expla-
nations is entirely satisfactory as the lake level dropped over
several months, and the Nupsghildharrs event is also close
to an active geothermal area. We do not, however, have an
alternative hypothesis at this time.

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