

Crustal response to inflation imaged by mapping of subsurface faults, slip directions, and stress changes during the 1993–1998 unrest in Hengill, SW-Iceland

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ARTICLE INFO

Keywords:

Hengill volcano
High-resolution hypocentre relocation
Earthquake source mechanism
Fault mapping
Coulomb stress

ABSTRACT

Between 1993 and 1998, the Hengill volcanic region in southwest Iceland was subject to a significant volcano-tectonic event which caused a total of 8 cm uplift and induced >90 thousand earthquakes, most of magnitude smaller than 2 M_L and a small intrusion has been proposed as its cause. In this paper, we study fault activation around the uplift center to analyze whether the Coulomb stress change induced by such an intrusion would cause the observed crustal response. The locations of a subset of 12,000 micro-earthquakes are significantly improved by a double-difference relocation algorithm and the use of cross-correlation of waveforms to refine phase arrival times. The vast majority of seismic activity is scattered, but twenty-five (25) rectangular prisms, consisting of 11 to 363 earthquakes and 200 to 3000 m long, were identified and are interpreted as faults. Most of these faults are sub-vertical (within 5° from vertical) and only two faults dip at angles of 62° and 65°, respectively. Focal mechanism analysis of individual events was applied to estimate slip direction on these faults. Simulation of the Coulomb stress changes reveals that the stress changes would have been able to induce the observed fault movement on 22 of 25 faults. Discrepancies on the remaining faults are likely caused by uncertainties in the rake directions. Our results emphasize that the seismicity caused by a volume change in the crust is controlled by the interplay between the local stress field and the induced stresses.

1. Introduction

Intrusions in active volcanic areas are often accompanied by earthquakes in the surrounding crust whose distribution is controlled by the intrusion geometry and other regional and local stresses (e.g., Ágústsson et al., 2019; Sigmundsson et al., 2015). The seismicity increases permeability in the vicinity of the newly formed heat source, promoting conditions that favor geothermal energy production. Another promoting factor are active faults that act as pathways for geothermal fluids. Active faults can be mapped by studying the distribution of small earthquakes and are preferred targets when drilling for geothermal fluids.

The Hengill volcanic system is one of about 30 active volcanic systems in Iceland, with its last major eruption occurring about 2000 years ago (Sæmundsson, 1995). Just east of the Hengill volcanic system lies the Hrómundartindur volcanic system (Fig. 1). Collectively, along with

the less active Grændalur volcanic system, these systems are referred to as the Hengill volcanic region (HVR) in this study. Its highest peak reaches 803 m above sea level and consists almost entirely of basaltic rocks (Sæmundsson, 1995). Numerous fumaroles and hot springs are visual indicators of heat sources in the shallow crust. Since 1990 the heat source has been exploited to provide the nearby capital area of Reykjavik with hot water and electricity. The HVR has been studied extensively over the years using geological, geochemical, geodetic and geophysical techniques (e.g., Jousset et al., 2011; Juncu et al., 2017; Sæmundsson, 1995).

In Iceland, most earthquakes occur along the plate boundaries, that is, the rift zones and transform segments, in particular the South Iceland Seismic Zone (SISZ) and the Tjörnes Fracture Zone (Fig. 1; e.g., Jakobsdóttir, 2008). The Hengill volcanic system is located at a triple junction between the North American and Eurasian plates and the Hreppar microplate connecting two extensional and one transform plate

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<https://doi.org/10.1016/j.jvolgeores.2022.107666>

Received 10 May 2022; Received in revised form 26 August 2022; Accepted 30 August 2022

Available online 6 September 2022

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boundary (Fig. 1; Einarsson, 2008):

- (1) the Reykjanes Peninsula Rift (RPR), which is the oblique on-land continuation of the Atlantic mid-ocean spreading ridge. It manifests itself in NNE-striking, sub-parallel series of normal faults and volcanic ridges, intersected by NS-trending sub-parallel right-lateral strike-slip faults.
- (2) The Western Volcanic Zone (WVZ), which is a spreading segment from Hengill towards the NNE. The WVZ takes up a lesser part of the plate spreading compared to the Eastern Volcanic Zone (Geirsson et al., 2012; LaFemina et al., 2005; Sigmundsson et al., 1995).
- (3) The South Iceland Seismic Zone (SISZ), which is a transform zone connecting the RPR and the Eastern Volcanic Zone, consisting of a series of NS-trending parallel right-lateral strike-slip faults

covering an area that is about 15–20 km wide and roughly 70 km long. In the SISZ, earthquakes of maximum magnitudes of M_s 7.2 can occur (Stefánsson and Halldórsson, 1988).

The NS-trending strike-slip faults continue westward across the HVR and the RPR and become shorter further to the west, limiting the maximum earthquake magnitude (Einarsson et al., 2020). Tectonic processes driven by the plate motions are also affected by magmatic and geothermal processes, all of which induce stresses that cause seismicity that varies in size, distribution and other characteristics. In particular, three cases are of special interest to this study:

(a) Tectonic earthquakes, which in the HVR have been observed to reach magnitude M_w 5.5 (Pétursson and Vogfjörð, 2009; Panzera et al., 2016) and in the neighboring SISZ have been up to M_s 7.2 (Stefánsson and Halldórsson, 1988).

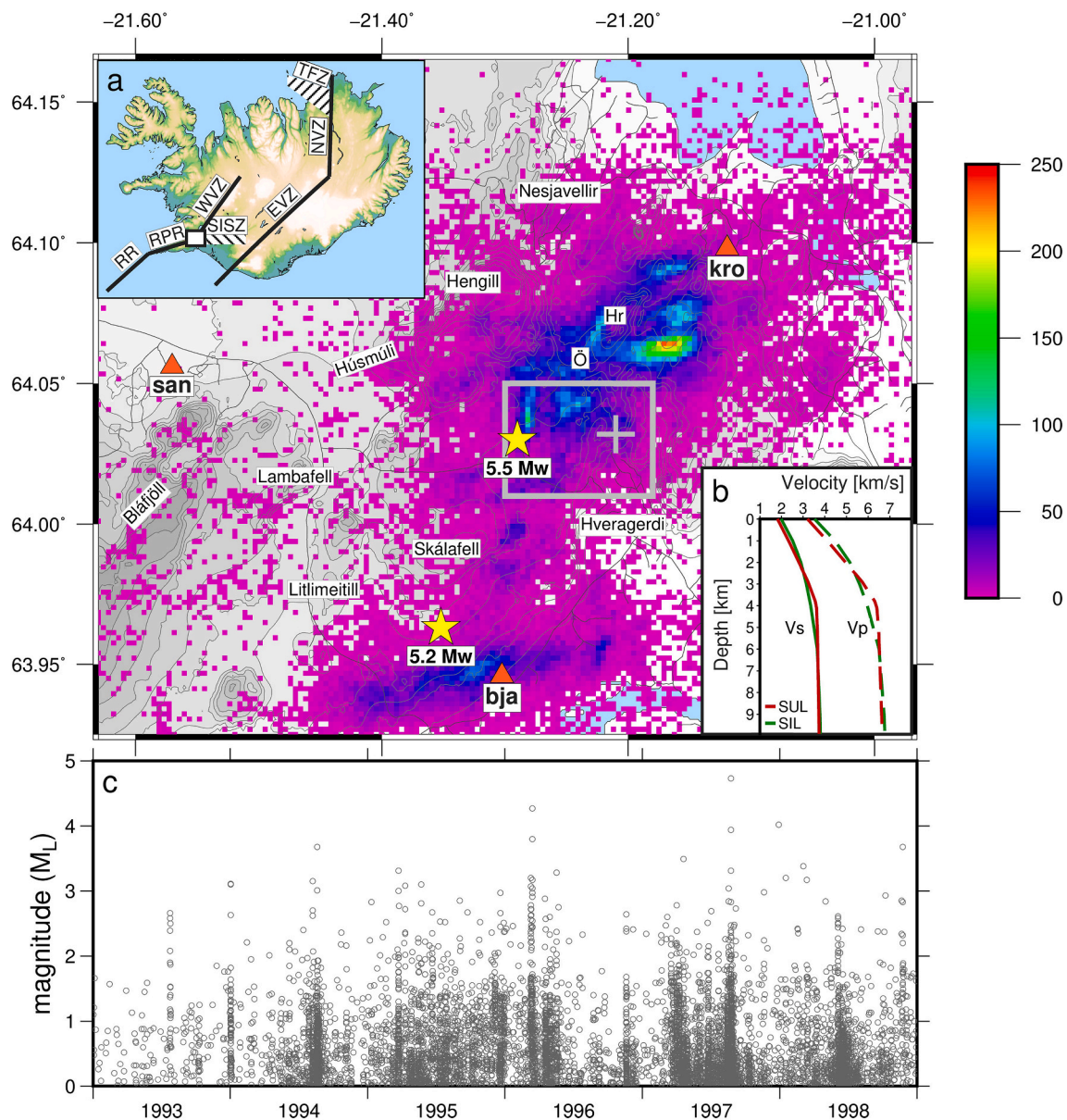


Fig. 1. Seismicity in the HVR from Jan 1st 1993 to Dec 31st 1998. (a) The colors on the map indicate the number of earthquakes located in each 200×200 m square. The study area is marked by the grey box and the grey cross shows the location of the uplift center as estimated from geodetic measurements (Feigl et al., 2000). Red triangles show the locations of SIL seismic stations. The letters Hr mark the location of the Hrómundatindur and Ö marks the location of Ölkelduháls. The white box in the top corner map shows the location of the HVR at the intersection of the Reykjanes Peninsula Rift (RPR), the Western Volcanic Zone (WVZ) and the South Iceland Seismic Zone (SISZ). The box in the bottom right corner shows the SIL and the SUL 1D (Vogfjörð et al., 2002) velocity models. (b) Magnitude time distribution of the catalogue data plotted in Fig. 1a. (For interpretation of the references to colour in this figure legend, the reader is referred to the web version of this article.)

(b) Earthquakes caused by propagating magma, which in the Icelandic crust are typically small in magnitudes ($M_L < 3.5$) (e.g., Greenfield et al., 2020). However, significantly larger events can be triggered, for example the 1976 M_S 6.4 Kópasker earthquake which was triggered by dike propagation from the Krafla central volcano (Passarelli et al., 2013) and the fifteen $4.5 \leq M_W \leq 5.64$ earthquakes triggered by the propagating dyke-intrusion preceding the 2021 Fagradalsfjall eruption on Reykjanes peninsula (Sigmundsson et al., 2022). The nature of focal mechanisms accompanying magmatic intrusion varies depending on the background stress field, as well as geometry and location of the intrusion, but the majority of tension axes conform with the direction of the regional tensional stress. The location of these earthquakes is typically confined to the volumes surrounding the dike or sill (Rubin and Gillard, 1998).

(c) Earthquakes caused by geothermal processes such as fluid propagation and cooling, which typically cause micro earthquakes with magnitudes of $M_L < 2$. Experience from geothermal fields in Iceland show that a wide range of fault orientations can be found in this environment and hypocentre locations are scattered. Fluid extraction and re-injection at geothermal fields further causes local changes in seismicity rates (e.g., Cardiff et al., 2018).

The HVR is one of the most seismically active areas in Iceland and systematic seismic monitoring of the region with the Icelandic regional network (SIL) started in 1991 (Böðvarsson et al., 1999). However, seismographs had been operated periodically in the HVR since the 1970s (Einarsson and Jakobsson, 2021). In addition, detailed seismological studies based on dense local networks have been carried out studying earthquake distribution and focal mechanisms (Foulger, 1988a, 1988b) and moment tensors (Miller et al., 1998). Li et al. (2019) relocated all events in the 1993–2012 SIL catalogue from the area using a collapsing method and were able to reproduce seismic structures seen in other temporary, more densely installed networks. According to Li et al. (2019), the depth of the brittle-ductile boundary varies between 5.5 and 8 km across the HVR and Weir et al. (2001) estimate the Moho at about 15 km depth. Foulger (1988a) suggests that the local seismicity can be roughly classified into two groups: (1) infrequent episodes associated with the rifting process caused by plate motion; and (2) continuous activity of smaller magnitude events in the geothermal system, possibly caused by thermal contraction and cracking (Foulger and Long, 1984; Foulger, 1988b). While an earlier study claims that about 50% of small earthquakes have non-double couple components (Foulger, 1988b), a later survey suggests, after studying moment tensors of 70 well-constrained earthquakes inside the high-temperature geothermal area, that up to 75% of events might have a significant non-double-couple component (Miller et al., 1998) showing opening of faults together with shear motion. The studies of Foulger (1988b) and Miller et al. (1998) were conducted in 1979/80 and 1991 when neither inflation or deflation nor large seismic sequences were taking place in the HVR, hence the long-term applicability of seismicity grouping may be questioned.

Between 1993 and 1998, the HVR was subjected to a significant inflation episode resulting in a maximum 8 cm of uplift. Based on GPS and leveling data, Sigmundsson et al. (1997) were the first to model the location of the uplift. Using additional InSAR measurements and Coulomb stress modelling, Feigl et al. (2000) confirmed and refined those findings, proposing an inflating point source (Mogi source) at latitude 64.032° and longitude $-21.213^\circ \pm 1$ km (Fig. 1) at about 7 ± 2 km depth. The cause of the uplift remains unresolved, a small magmatic intrusion has been suggested (Feigl et al., 2000) while Tryggvason et al. (2002) suggest the inflation may be of hydrothermal origin.

The uplift was accompanied by over 90 thousand earthquakes detected by the SIL network (Fig. 1). The seismicity is located all over the wider HVR and south of Mt. Skálafell. Feigl et al. (2000) modeled the Coulomb stress change from the intrusion and found that most of the earthquake hypocentres are located in parts of the crust where the Coulomb stress increases. While the vast majority of those earthquakes

were originally located in between 2 and 8 km depth (e.g., Foulger, 1988a, 1988b), more recent studies (Hjaltadóttir and Vogfjörð, 2009; Li et al., 2019; Vogfjörð et al., 2005) concludes that most of the events are shallower than 5 to 7 km. Faults active from 1996 to 1998 have been mapped by Vogfjörð et al. (2005) using relative relocation, showing a variety of fault directions with most faults striking between NNE and ENE, in accordance with the local stress field.

Comparing the Hengill uplift from 1993 to 1998 to other intrusion events in Iceland (e.g., Ducrocq et al., 2021; Metzger et al., 2011; Pedersen et al., 2007), it becomes evident that the uplift is accompanied by an unusual amount of seismicity, raising the question how a relatively small uplift could have such extensive consequences. The aim of this study is to analyze the seismic activity in the direct vicinity of the uplift in greater details to identify faults which are most likely to be directly triggered by the uplift in contrast to those at greater distances which are considered to occur as part of the stress build-up in the SISZ. We then model the Coulomb stress change caused by the uplift source to analyze whether it would have promoted the observed movement on these faults.

2. Data

We use catalogue data and waveforms from the SIL network from Jan 1st 1993 to Dec 31st 1998 in the Hengill area (Fig. 1). In 1993, seismic activity started to exceed the background seismicity rate and an average of 86 earthquakes were recorded each month until October 1993. From November 1993 until July 1994 activity decreased to <20 earthquakes/month (Fig. 1c). A large earthquake swarm occurred August 15 to August 25, 1994, with about 5600 earthquakes focused north and northwest of the uplift center. From September 1994 to January 1995 earthquake rates went down to a minimum before a series of swarms lasting for much of the year started in February 1995. Afterwards, seismicity rates remained elevated with several hundred up to 7700 earthquakes per month until the end of 1998 with peaks rates from April to August 1997 when a magnitude 5.0 M_W earthquake occurred on August 24th in the central part of the study area. The single one month with the highest number of recorded earthquakes was September 1998 ($> 18,500$). The seismicity also propagated to the south of Mount Skálafell (Fig. 1a) where the episode reached its climax in 1998 when two earthquakes of magnitude M_W 5.5 and M_W 5.2 were recorded on June 4 and November 13, respectively.

The seismicity was detected and recorded by the SIL national network, which in 1993 consisted of 18 seismic stations covering the main seismically active zones in Iceland (Böðvarsson et al., 1999). Of these, 12 seismometers were located in South Iceland. Another 2 stations were added to the regional network around the Hengill volcano in 1996 after which the depth range of hypocentres decreased (Fig. A2 in Supplementary material 1). At that time, the SIL network consisted of three-component Lennartz short period (1 s and 5 s) sensors and one Guralp 3T broadband sensor, recording 3-component waveform data sampled at 100 Hz.

The earthquakes used in this study were automatically detected and located by the SIL software, using the SIL velocity model (Stefánsson et al., 1993). The P- and S-wave arrival times were manually revised for all events in the catalogue, and P-wave polarities were picked. During the study period, only waveform data from detected earthquakes were stored due to limitations in data transmission and storage capacity. As a consequence, a correlation algorithm or comparable software cannot be used to search for more events in the data set.

In this study, we focus on the seismic activity in the direct vicinity of the uplift (grey box in Fig. 1) to learn about fault activation in those parts of the crust which are most affected by the induced stress changes. The study area reaches from latitude 64.00° to 64.05° and from longitude -21.18° to -21.33° and is centered not on, but north-west of the center of uplift and was selected to capture the distribution of the seismicity. In the study area, the SIL network located over 12,000 earthquakes,

ranging between 0.0 and 4.7 M_L (local magnitude) as estimated by the SIL analysis system (Fig. 1c). For earthquakes with magnitudes under approximately 3.5, the local magnitude determined by the SIL system is a reasonable estimate of actual magnitude while magnitudes of larger earthquakes tend to be underestimated (Greenfield et al., 2020; Panzera et al., 2016). Those events are better described using M_w magnitudes and therefore, both M_L and M_w magnitudes will be used in this study, when appropriate.

The distribution of earthquake hypocentres in the study area, based on manually revised locations, appears to be rather diffuse, and linear alignments or clear faults are not obvious to the naked eye (Supplementary material 1 Fig. A1). The seismic activity is not equally distributed around the center of uplift but is denser to the north and west compared to the east and south (Fig. 1; Clifton et al., 2002; Feigl et al., 2000). The seismic activity is mostly continuous with varying intensity, but we could identify about two swarm-like episodes per year.

After October 1996, when four additional seismic stations were installed, including station KRO at about 7 km distance from the uplift centre (Fig. 1), the seismic activity is mostly confined to a depth range of 4 to 8 km. The catalogue earthquake locations have only limited accuracy due to the sparsity of the seismic network and, in addition, unmodeled variations in the earth structure. Since the network configuration lacks a seismic station in the center of seismicity, the depth is the least constrained of the location coordinates. The accuracy of the routine locations of the SIL system is in the range of 500–700 m for the horizontal coordinates and about 1000 m in depth, depending on picking accuracy and wavelength (Slunga et al., 1995).

3. Methods

3.1. Relative relocation method

To improve the accuracy of earthquake locations, we relatively relocated the earthquakes making use of the Slunga double-difference relocation software (Slunga et al., 1995). The time difference between arrival times of seismic waves can be estimated with high precision using cross-correlation (CC) if their waveforms are similar (Slunga et al., 1995). This is the case if earthquakes occur on the same or neighboring faults and have the same or similar mechanisms. If the distance between a pair of relocated earthquakes is small in comparison to the distance to the recording seismic station, a small change in the absolute locations of this pair of as little as 200 m will cause an arrival time difference of 0.5 ms, big enough to be identified with cross-correlation. This method not only improves relative locations, but absolute locations as well (Slunga et al., 1995).

The software automatically arranges the input seismicity into groups and then the earthquakes are relocated pairwise for all possible pairs of events within each group. Input parameters for this process are the minimum and maximum number of earthquakes allowed in a group as well as the size of the group (radius in km) and the distance between groups (distance between midpoints of groups in km). We use groups containing 9 to 45 events. The maximum group size is dictated by the software while groups of smaller size are not of interest when mapping faults. Suitable parameters for the size of groups and the distance between groups highly depend on the characteristics of the seismicity which are the size of the faults or fissure swarms where earthquakes are expected to be similar. For each pair of events in a group, a cross-correlation of the waveforms is carried out. Only pairs with a CC coefficient of 0.8 and higher are used for further processing steps. Earthquakes that do not correlate with others are not considered further. This ensures that earthquakes that are spatially close but not similar are not relocated and do not appear in the output. Then, a least squares problem is solved considering the absolute P- and S-arrival times together with the relative P-, S- and (S–P)-arrival times for each group. Here, the CC coefficient is used to weight the arrival times of each pair of events. If an event is part of more than one group, the final location is an average of

the different relocated locations.

The SIL catalogue locations were calculated with the SIL velocity model which is routinely used by the IMO for earthquake location in Iceland and is based on a refraction seismic experiment conducted in 1990 (Bjarnason et al., 1993; Stefánsson et al., 1993). The seismic stations used in this study are mostly located in the SISZ where the SUL velocity model (Vogfjörð et al., 2002) (Fig. 1b) is a more accurate approximation of the crustal structure and is therefore used for the relative relocation. At depths shallower than 3 km, the seismic velocities in the SUL model are lower than the SIL model and at depths between 3 km and 6 km the SUL velocities are higher. From 6 to 9 km depth, where the deeper events in the study area occur, the models are similar and below 9 km the SIL model velocities are slightly higher (Fig. 1b). For further information on how the relative relocation changed the catalogue locations, see Fig. A3 in Supplementary material 1.

3.2. Focal mechanisms

Focal mechanisms describe the rupture plane and slip direction of an earthquake and are often determined based on first motion data (i.e., P-wave polarities). For small earthquakes, a dense network of seismic stations is required to reach a sufficient number of first motion data points to derive a well constrained focal mechanism. Since most of the earthquakes in this study are very small (Fig. 1c), the network installed in and around Hengill during the study period is too sparse to allow for a focal mechanism analysis based on first motions only. Therefore, we use the method from Rögnvaldsson and Slunga (1993) which in addition to first motions uses spectral amplitude (i.e. power) of the far-field P- and S-waveforms. Similar to the absolute location method described above, this method is applied by the IMO in routine analysis of small earthquakes. The expected spectral peaks of the P- and S-phases are calculated for all possible combinations of strike, dip and rake of the fault plane at 4° increments (about 40,000 possibilities) and compared to the observations at the seismic stations. All focal mechanisms which fulfill the first motion observations and where the spectral amplitude residuals are below a predefined value are stored as possible solutions but the solution with the smallest misfit is used to represent the “best fitting” mechanism in the routine analysis.

The described method from Rögnvaldsson and Slunga (1993) allows for double-couple (DC) solution only. For our data set, this is an appropriate assumption since the vast majority of the earthquakes are very small and the seismic network is not dense enough to enable full moment tensor inversion. Also, since the goal of this study is the mapping and analysis of faults, the focus is exclusively on DC events and the DC component of events. We do not rule out the existence of non-DC events though, and earlier studies have shown them to exist in the HVR (Foulger, 1988b; Jousset et al., 2011; Miller et al., 1998).

The spatial distribution of the earthquakes is used to further constrain the nodal planes of the relocated earthquakes. To achieve this, we use the QuakeLook software by Ragnar Slunga, which provides an additional constraint to select a “best fitting” mechanism for each event from the stored range of possible mechanisms. The QuakeLook software is a visualization and interpretation software which reads a subset of selected earthquake locations that have been visually identified and manually selected by the operator and calculates a best fitting plane through the locations (e.g., Lund and Bödvarsson, 2002). These subsets of earthquakes are interpreted as faults. From the dataset of possible focal mechanisms for each event, the mechanism with a nodal plane closest to the best fitting plane is selected. QuakeLook also enables distinguishing between the two nodal planes of the mechanisms to identify the fault plane of an assembly of earthquakes.

3.3. Calculation of average slip vectors on faults

A slip vector describes the direction and magnitude of movement along a fault. We calculate average slip vectors on each of the manually

identified faults from the focal mechanisms which best fit the common plane through the group of selected earthquakes. We calculate two slip vectors, the first by averaging the observed slip of all earthquakes and the second by weighting the slip of each earthquake linearly with its magnitude before averaging.

3.4. Coulomb stress change

Coulomb stress change modelling is used to evaluate whether the stress changes induced by the uplift could have induced the slip on the faults as they were observed. We use the Coulomb 3 software to model the stress change caused by the inflation source (Lin and Stein, 2004; Toda et al., 2005). The Coulomb 3 software then allows us to calculate the stress change on faults of given orientations (strike and dip) and slip directions (i.e., rake). In particular, we use the subsurface faults and rake as mapped by the joint interpretation of the relative earthquake locations and the focal mechanisms (see Section 4) as receiver faults for the stress calculations. Positive Coulomb stress changes facilitate movements on faults while negative Coulomb stress changes make movement less likely, assuming those faults are critically stressed (e.g., King et al., 1994). For the calculations of the Coulomb stress change, we assume a Poisson ratio of 0.25, a friction coefficient of 0.4 and a Young's modulus of 80 GPa.

4. Results

In this study, a subset of 12,000 earthquakes located close to the

modeled uplift center during 1993–1998 is selected for relative relocation and analysis. The seismic activity is both continuous and of swarm-like nature (Fig. 1). We are able to correlate and relatively relocate 10,406 earthquakes, some of which form clusters and alignments, but the majority of the activity is scattered (Fig. 2). We manually identify a total of 25 faults using the QuakeLook software, each delineated by 11 to 363 earthquakes (average 67, median 46), located at about 0.5 to 7.5 km depth (Fig. 2). For more details on the fault parameters, see Table A1 in Supplementary material 1. The length of the faults, as defined by the seismicity extent, is 122 to 3821 m with an average of 836 m and a median of 562 m. Twenty-three out of 25 faults dip within 5° from the vertical. NNE and SSW are the most frequent strike directions (see rose diagram in Fig. 2). In the interest of clarity and comprehensibility, the crust containing the faults is divided into five faults areas and the faults in each area are numbered consecutively. A map view as well as profile views of each area together with more details on the faults in each area can be found in Figs. A5 to A9 in Supplementary material 1.

The distribution of the average rake of earthquakes defining each of the faults shows that most faults are predominantly strike-slip with a smaller normal component (Figs. A5 to A9 in Supplementary material 1). Right-lateral strike-slip prevails. Only two faults dip sub-vertically at angles of 62° and 66° (faults 3.3 and 3.6). They are in the same location but at different depth ranges and strike in similar directions (215° and 223°).

Fig. 3 shows fault 1.3 in more detail. Similar plots for the other faults and more detailed information can be found in Supplementary material 1. Fault 1.3 consists of 128 earthquakes, is oriented in a northerly

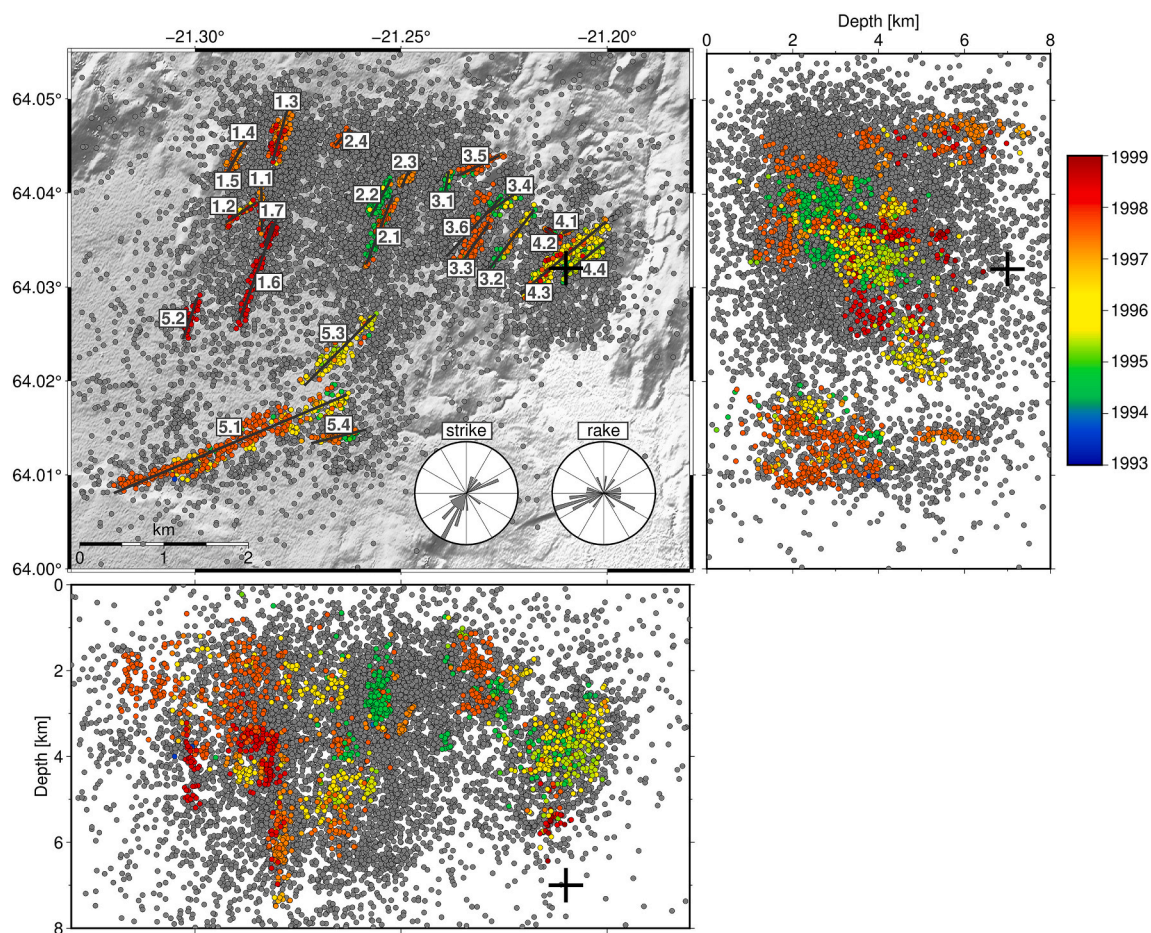


Fig. 2. Relatively relocated seismicity in the study area shown in map view as well as on N-S and E-W profiles; vertical compression is 1:2. While the majority of the relatively relocated seismicity is scattered (grey), a smaller number of events is located on one of 25 faults (colored dots, black lines). The black cross shows the modeled location of the uplift center (Feigl et al., 2000). The two rose diagrams display the distributions of strike and average rakes of earthquakes on the faults. (For interpretation of the references to colour in this figure legend, the reader is referred to the web version of this article.)

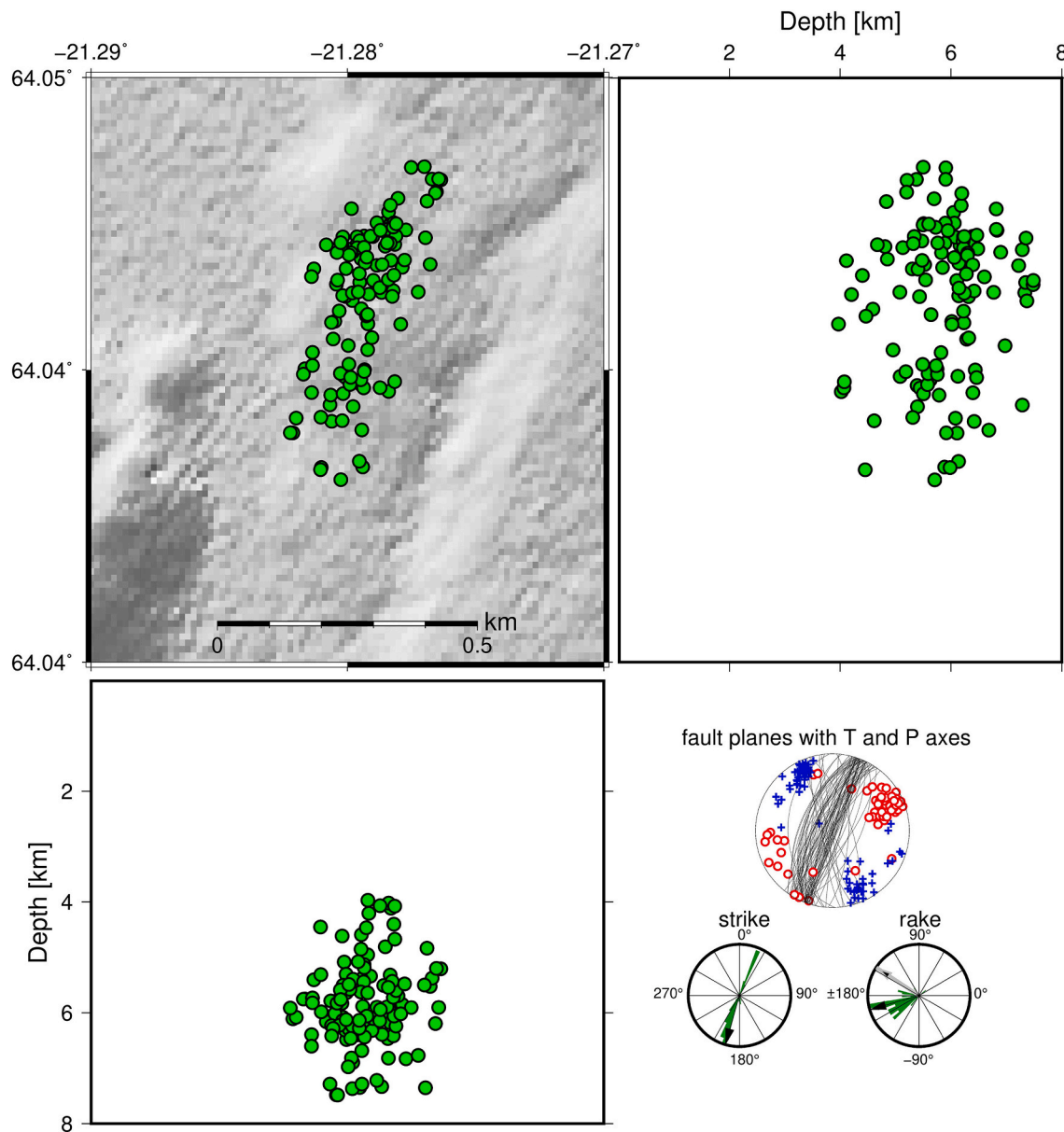


Fig. 3. Map view and E-W and N-S profiles of earthquakes located on fault 1.3, vertical compression is 1:10. The event distribution defines a common fault plane with strike 196° , and dip 89° . The focal sphere in the right lower corner shows the nodal planes (grey) of the individual earthquakes that are closest to the strike of the common fault plane. Red circles show the P-axes, and the blue crosses show T-axes. The two rose diagrams in the lower right corner respectively display the distribution of the strike and the rake angles of the individual earthquakes. The black arrows show the averages of their strike and rake angles, respectively; the grey arrow the average rake weighted by the magnitude of the earthquakes. (For interpretation of the references to colour in this figure legend, the reader is referred to the web version of this article.)

direction, 617 m long and ranges over a depth from 4 to 7.5 km. The P-axis of most earthquakes is slightly dipping and oriented in roughly NW or SE direction while the T-axes, also slightly dipping, are oriented in NE or SW directions. Most of the earthquakes show right-lateral strike-slip movement with a small normal component (rose diagram on Fig. 3). The larger earthquakes on this fault dominating the weighted average slip vector have a thrust component.

5. Discussion

5.1. Uncertainty of relocated hypocentres

To estimate the uncertainties of the relatively relocated hypocentres, we test different input parameters in the relocation process, that is, the size of groups of earthquakes and the distance between groups. In a first

step, we compare the results visually. The relative locations vary little between parameter sets while the absolute location of the earthquake sequences show lateral movement of up to 500 m between sets. To refine this first estimate, we selected 7 sets of physically meaningful parameters, that is, the group diameter is of the same order as the lengths of the faults, which result in similar earthquake locations for a subset of about 1700 earthquakes. After excluding outliers, we calculate the average and standard deviation of the differences in coordinates and depths which we use as a rough estimate of their accuracy. The standard deviation of the longitudes and latitudes suggest that the horizontal errors lie within ca. 260 m (Supplementary material 1 Fig. A4). With a standard deviation of 800 m, the depth is much less refined due to the lack of seismic stations near the center of the seismic activity. In addition, the choice of velocity model has a major effect on the depth at which earthquakes are located as has been shown for the Hverahlíð area in the

southern part of the study area by Vogfjörð and Hjaltadóttir (2007).

5.2. Comparison to other faults in region

We compare the faults mapped in this study to surface faulting documented by Clifton et al. (2002) and Steigerwald et al. (2018). Both of these studies extensively mapped surface faults in the HVR and a subset of these faults lies within our study area (Fig. 4). The comparison shows that most of the faults mapped at depth with the relocated earthquake hypocentres do not have surface expressions while there is a reassuring concordance between the strike of our faults and the mapped surface faults. On two occasions faults mapped at the surface and at depth partially overlap. Faults 1.4 and 1.5 do overlap but are located at very different depths (Supplementary material 1 Fig. A5) and they coincide with a surface fault that was mapped by both Clifton et al. (2002) and Steigerwald et al. (2018). Faults 1.6 and 1.7 are located in direct continuation of one of the longer surface faults located by Steigerwald et al. (2018).

5.3. Stress changes and triggering of seismicity

Pedersen et al. (2007) compared the stressing rate caused by the Hengill intrusion to the seismic moment release by earthquakes in the area from 1993 to 1998 to two other intrusion events in Iceland that were also accompanied by an increase in seismic activity, that is a rifting event in Krafla central volcano in July 1978 and multiple intrusion events in Eyjafjallajökull volcano between 1994 and 2000. They found that there is no simple correlation between intrusion rate and seismicity

and that each case must be considered individually taking the background stress field into account, which plays a major role controlling the amount of seismic energy released. Examples for intrusion events in Iceland where little to no change in seismicity was observed are the Theistareykir intrusion in 2007–2008 (Metzger et al., 2011) and the Hengill intrusion from 2017 to 2018 (Ducrocq et al., 2021), further emphasizing the conclusion of Pedersen et al. (2007) that intrusion induced stress changes are not predictive of changes in seismicity rates. A study of the spatial distribution of seismicity around intrusions by Rubin and Gillard (1998), using modelling calculations and a case study of the Kilauea dike intrusion in 1983, concludes that seismicity reflects ambient stresses rather than dike geometry. In their entirety, these observations support the notion that seismicity accompanying intrusion differs immensely and is dictated by the pre-stress state. For the case of the Hengill intrusion, this implies that the region was near-failure prior to the uplift and a relatively small intrusion could cause a major change in seismicity. To further investigate this relationship, in the following, we look at the stress changes that would have been caused by the intrusion.

Positive volume changes in the crust, caused by intrusion of material from depth, lead to a Coulomb stress change in the surrounding crust which can bring faults closer to or further from failure. The deformation caused by the 1993–1999 uplift is relatively small, suggesting that the uplift acts as a mere trigger while the stress field pre-loads the faults and controls their orientation. A study of the neighboring seismically active areas on the Reykjanes Peninsula, comparing stress derived from earthquake focal mechanisms and strain calculated from GPS data (Keiding et al., 2009), suggests that the stress is predominately driven by

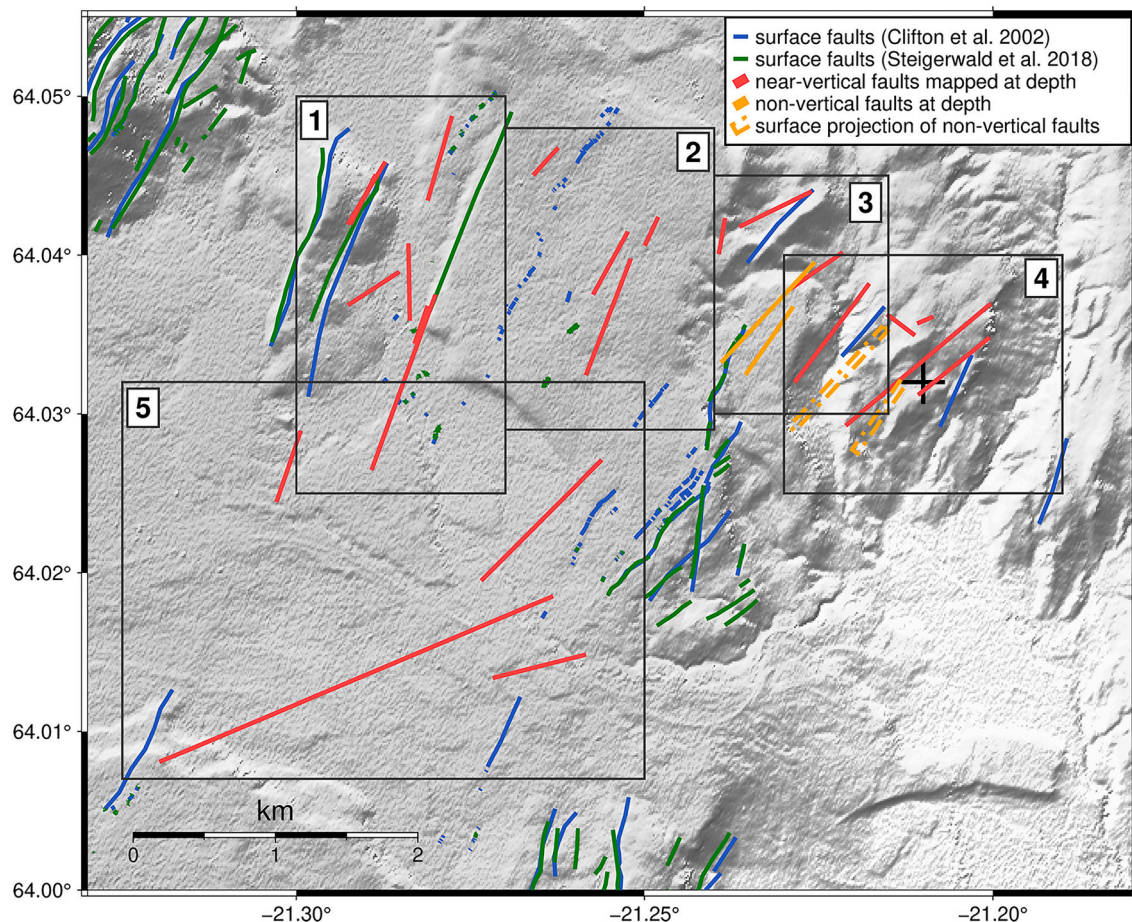


Fig. 4. Comparison of faults mapped in this study to surface faults. Red lines show the locations of the near-vertical faults, the orange lines and dashed boxes show the locations at depth of two non-verticals faults and their surface projection. Blue and green lines show the location of surface faults mapped by Clifton et al. (2002) and Steigerwald et al. (2018), respectively. (For interpretation of the references to colour in this figure legend, the reader is referred to the web version of this article.)

the plate motion, and that most of the time, geothermal utilization and small episodes of magma movement, such as the 1993–1999 uplift, act as a secondary triggering mechanism only. If, however, the geothermal or inflation-induced deformation becomes large enough, then it can cause faulting on its own.

We calculated the Coulomb stress change (CSC) caused by the inflation at the location and with volume change proposed by Feigl et al. (2000), using the strike, dip, and rake from the faults defined by the relative earthquake locations as receiver faults (Table A1). We found a marked increase on 18 of the 25 mapped faults while there is no change on 4 faults and on 3 faults, the CSC is negative (Fig. 5). When considering the magnitude-weighted average rake, results vary slightly, showing positive CSC on 17 faults, no change on 3 faults and negative CSC on 5 faults. It needs to be taken into consideration that there are inherent uncertainties in the key parameters (strike, dip, depth, location, and rake) of each fault that certainly affect the calculated CSC. The CSC calculations also do not take into account other transient sources of stress change, for example slip on the mapped faults or other earthquakes outside the study area.

To test whether another uplift location might better explain the observed fault movements, we test two other locations within the error margins provided by Feigl et al. (2000). The longitude of the modeled uplift centre has a lateral uncertainty of about 300 m while the latitude has an uncertainty of 1000 m. We re-calculated the CSC on the faults assuming an inflation source both 1 km further north and south. When

moving the inflation source 1 km further north and using the average rake, the CSC is positive on 17 faults, negative on 4 faults and no change is calculated for 4 faults, for weighted average rake, CSC is positive on 20 faults, neutral on 3 faults and negative on 2 faults. Moving the inflation source 1 km south yields a positive CSC on 17 faults, negative on 5 faults and no change on 3 faults when average rake is used and positive CSC is obtained on 11 faults, neutral on 3 faults and positive on 11 faults with weighted average rake. Based on these findings, moving the center of uplift 1 km further north compared to the modeled value from Feigl et al. (2000) appears more plausible (i.e., closer to the deformation suggested by Sigmundsson et al. (1997)), but the difference is minor.

While most of the faults are in agreement with the static stress triggering of the seismicity, some faults would not have been triggered by the modeled CSC. This is most likely due to uncertainties in the fault locations or the parameters describing the faults and the movements thereon, that is, strike, dip and rake. Of those, rake is the least well defined (Figs. A5 – A9 in Supplementary material 1) and the most likely explanation for the discrepancy between the observations and the modelling results.

The recent uplift and the induced seismicity in the Hengill region should potentially increase or maintain the permeability of the surrounding crust and therewith favor heat production in the area of the uplift and in the geothermal field in the Hveragerdi village. Both are areas with known surface manifestations of geothermal activity such as

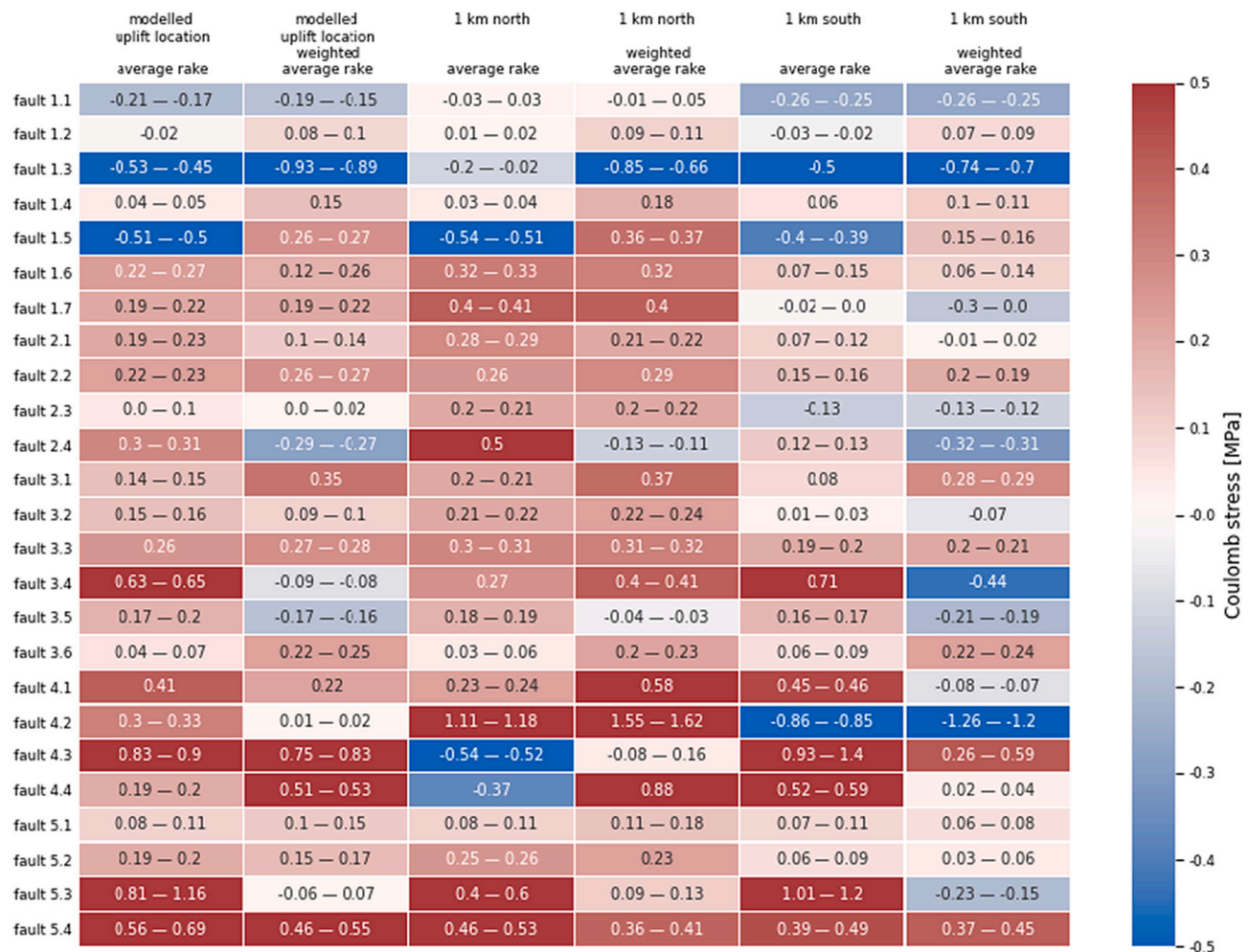


Fig. 5. Comparison of Coulomb stress change (in MPa) on the mapped faults caused by inflation sources in different locations. For the stress calculation both the average rake on the fault as well as the average rake weighted by magnitudes is used (see section 4). Positive changes in Coulomb stress are highlighted red while negative changes are highlighted blue. (For interpretation of the references to colour in this figure legend, the reader is referred to the web version of this article.)

hot springs, fumaroles and clay alteration. Assuming the uplift was indeed caused by an intrusion, the cooling magma would emit heat into the surrounding crust, eventually providing additional heat to fuel those manifestations, however the migration of heat to the surface through convection and conduction would take centuries and be too diffuse to be noted.

6. Conclusion

1. The relative relocation of seismicity considerably improves the earthquake locations. Tests show that the locations have small internal uncertainties while the absolute locations of the earthquakes are dependent on the parameter selection for the group sizes used in the algorithm.
2. The relocated hypocentres reveal 25 lineaments that we interpret as faults but most of the activity is scattered or located in wider, possibly weaker, areas of the crust. 23 out of the 25 faults are near vertical and the rupture mechanism is strike-slip dominated, suggesting that the stress regime in the study area is dominated by the SISZ transform zone rather than the rifting along the Reykjanes Ridge and the Western Volcanic Zone.
3. The Coulomb stress change, caused by a volume change in the modeled uplift location (including its error margins), does account for the movement on most faults. We conclude that uncertainties in the estimated rake angles assumed in the Coulomb failure analysis cause the false negative results on the few remaining faults.

CRedit authorship contribution statement

Hanna Blanck: Conceptualization, Software, Investigation, Writing – original draft, Visualization. **Kristín S. Vogfjörð:** Conceptualization, Writing – review & editing, Supervision, Funding acquisition. **Halldór Geirsson:** Conceptualization, Writing – review & editing, Supervision, Funding acquisition. **Vala Hjörleifsdóttir:** Writing – review & editing, Supervision.

Declaration of Competing Interest

The authors declare that they have no known competing financial interests or personal relationships that could have appeared to influence the work reported in this paper.

Data availability

Routine locations since 1995 used as input are available at: <http://hraun.vedur.is/ja/viku/>. In near future, all input data will be available at Orfeus DMC (waveforms) and EPOS services (locations).

Acknowledgements

We thank Ragnar Slunga who provided the QuakeLook software and Philippe Jousset for providing valuable comments improving the manuscript. We also thank two anonymous reviewers for their thoughtful comments that further helped to improve this manuscript. This work has been supported by the Icelandic Research Fund, grant number 174377-051. Fig. 5 was produced with seaborn (Waskom, 2021). Figs. A3 and A4 were produced with matplotlib (Hunter, 2007). All other figures in this publication were produced with GMT (Wessel et al., 2019).

Appendix A. Supplementary data

Supplementary data to this article can be found online at <https://doi.org/10.1016/j.jvolgeores.2022.107666>.

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