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16 **Cyclical geothermal unrest as a precursor to Iceland's 2021 Fagradalsfjall eruption**

17 **Authors**

18 Ólafur G. Flóvenz (1), Rongjiang Wang (2,3), Gylfi Páll Hersir (1), Torsten Dahm (2,4), Sebastian
19 Hainzl (2), Magdalena Vassileva (2,5,9), Vincent Drouin (1, 10), Sebastian Heimann (2,4), Marius Paul
20 Isken (2,4), Egill Á. Gudnason (1), Kristján Ágústsson (1), Thorbjörg Ágústsdóttir (1), Josef Horálek
21 (6), Mahdi Motagh (2,5), Thomas R. Walter (2,4), Eleonora Rivalta (2,7), Philippe Jousset (2),
22 Charlotte M. Krawczyk (2,8), Claus Milkereit (2)

23 *1. ÍSOR, Iceland GeoSurvey, Reykjavík, Iceland*

24 *2. Helmholtz Centre Potsdam GFZ German Research Centre for Geosciences, Potsdam, Germany*

25 *3. Institute of Geophysics and Geomatics, China University of Geosciences, Wuhan 430074, China*

26 *4. University of Potsdam, Institute of Geosciences, Potsdam, Germany*

27 *5. Institute for Photogrammetry and GeoInformation, Leibniz University Hannover, 30167 Hannover,*
28 *Germany*

29 *6. Institute of Geophysics, Czech Academy of Sciences Prague, Czech Republic*

30 *7. Department of Physics and Astronomy, University of Bologna, Italy*

31 *8. TU Berlin, Institute of Applied Geosciences, Germany*

32 *9. Leibniz University Hannover, Institute of Photogrammetry and GeoInformation*

33 *10. Icelandic Meteorological Office, Reykjavík, Iceland*

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38 **Abstract**

39 Understanding and constraining the source of geodetic deformation in volcanic areas is an
40 important component of hazard assessment. Here, we analyse deformation and seismicity for
41 one year prior to the March 2021 Fagradalsfjall eruption in Iceland. We generate a high-

42 resolution catalogue of 39,500 earthquakes using optical cable recordings and develop a
43 poroelastic model to describe three pre-eruptional uplift and subsidence cycles at the
44 Svartsengi geothermal field, 8 km west of the eruption site. We find the observed deformation
45 is best explained by cyclic intrusions into a permeable aquifer by a fluid injected at 4 km depth
46 below the geothermal field, with a total volume of $0.11 \pm 0.05 \text{ km}^3$ and a density of 850 ± 350
47 kg/m^3 . We therefore suggest that ingress of magmatic CO_2 can explain the geodetic,
48 gravity, and seismic data although some contribution of magma cannot be excluded.
49 Our results demonstrate that inflation and deflation cycles with periods of several months over
50 scales of tens of square kilometres commonly observed in volcanic systems around the world
51 can be caused by the migration of deep magmatic fluids and gases into upper crustal
52 hydrothermal systems. It highlights the interaction of volcanic processes and the behaviour of
53 geothermal systems to address hazard assessments.

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57 Recent volcanic and seismic unrest with surface deformation at the Reykjanes Peninsula
58 (RP) plate boundary in SW Iceland, in and around the Svartsengi high-temperature (HT) field
59 (**Fig. 1**), reveals a previously unexplored cyclic interaction between tectonic spreading,
60 magmatic reservoirs, and supercritical magmatic fluids interacting with hydrothermal
61 reservoirs.

62 On January 22nd, 2020, an earthquake swarm started 3 km east of Svartsengi.

63 Simultaneously, uplift was recorded at the geothermal field close to the re-injection site¹ of the
64 power plant, followed by subsidence. Three such cycles of inflation and deflation occurred
65 until July 2020, where each inflation was followed by continuous deflation and diminishing
66 seismicity (**Fig. 2c**). Similar inflation started in August 2020 at the centre of the Krýsuvík HT

67 field, 20 km east of Svartsengi (**Fig. 1**). In February 2021, crustal extension and an intense
68 earthquake swarm revealed the formation of a NE striking magmatic dyke between the two
69 HT fields, followed by the Fagradalsfjall eruption 8 km east of Svartsengi on March 19th, 2021.
70 The spreading axis of the Mid-Atlantic Ridge comes on land at the SW corner of the RP.
71 There, it bends into a 60 km long N70°E striking oblique plate boundary, expressed by a 5-10
72 km wide seismic and volcanic zone, where large episodic earthquake swarms occur every 20-
73 40 years² with magnitudes up to M6, mostly on N-S trending strike-slip faults. Volcanic
74 eruptions have occurred at intervals of 800-1000 years during the past 4000 years, the last
75 one ending in 1240 AD³. Each volcanic episode might last for 1-3 centuries with basaltic lava
76 flows from N45°E trending fissures extending into the adjacent plates⁴. HT geothermal fields
77 with reservoir temperature of 240-330°C at 1-3 km depth⁵ have formed at the intersection of
78 the seismic zone and the main volcanic fissure swarms (**Fig. 1**). One well, IDDP-2 at the
79 Reykjanes HT field (**Fig.1**), was drilled to 4.6 km depth⁶ close to the brittle-ductile transition
80 (BDT) where bottom hole temperature is estimated to be about 600°C⁷ just beneath a
81 hydrostatic pressurized aquifer (~35 MPa). Corresponding results were obtained in the IDDP-
82 1 well at the Krafla HT field in North Iceland. The crust above the aquifers is in both cases
83 fully elastic and the fluid pressure in the rock is hydrostatic down to the BDT.
84 The upper crust of the RP is approximately 4.5 km thick^{8,9,10} and composed of basaltic
85 extrusives with a downward increasing alteration and a higher proportion of intrusives. The
86 lower crust down to Moho at ~15 km depth is thought to be made of intrusives with no
87 evidence of melt^{10,11,12}. The BDT is generally at 6-7 km depth beneath the RP, rising up to 4-5
88 km depth below the HT fields^{13,14,15} with an estimated temperature of ~600°C^{1,7,16,17}.
89 The 76 MW_e power plant at the Svartsengi HT field and the Blue Lagoon Spa (**Ext. data Fig.**
90 **1**) are the heart of the Geothermal Resource Park, providing electricity and hot and cold water

91 to 25,000 residents and local industries¹⁸. The average annual production rate is about 0.45
92 m³/s, of which about 0.3 m³/s are re-injected into the reservoir. These re-injection rates are an
93 order of magnitude smaller than the rates needed to explain the observed uplift in 2020.
94 The cyclic deformation and earthquake activity at two distinct HT fields at the RP plate
95 boundary and the time lag prior to a distant fissure eruption is unusual and was previously not
96 observed. Using a comprehensive modelling approach, we show that the ascent of magma-
97 derived volatiles into a sealed aquifer above the BDT beneath the Svartsengi HT field can
98 explain the uplift and subsidence cycles, and can be interpreted as a precursor to a coming
99 eruption.

100

101 **Transient deformation at the Svartsengi HT field**

102 Analysis of twelve months of satellite InSAR time-series data supported by GNSS data (see
103 methods), processed in both ascending and descending configuration, reveals an elliptical
104 area exceeding 80 km² affected by uplift and subsidence at Svartsengi, along with minor
105 horizontal displacements (**Fig. 2a & Ext. data Fig. 2**). The major axis of the elliptical area is
106 ~12 km long, striking N60°E and the perpendicular axis is 8.5 km long. The major axis follows
107 the dominant strike of the geothermal reservoir (**Ext. data Fig. 1**) but deviates both from the
108 N45°E strike of volcanic fissures and the N to NE trending strike-slip faults mapped at the
109 surface on the RP.

110 The duration of successive uplift episodes increased while the uplift rate decreased
111 correspondingly (**Ext. data Fig. 3**). The first episode had the highest uplift rate of 2.2 mm/day
112 and a total uplift of 66 mm over 30 days (**Fig. 2b**). It was followed by 18 days of subsidence,
113 totalling less than 10 mm. The second episode had an uplift rate of 1.1 mm/day and a total
114 uplift of 55 mm over 48 days. It was followed by a faster subsidence episode lasting
115 approximately 24 days, with a total subsidence of 16 mm. The final episode, with an uplift rate

116 of 0.5 mm/day, produced a total uplift of 32 mm over 60 days. On July 18th, an earthquake of
117 magnitude M4.1 occurred near the inflation centre¹⁹ (**Ext. data Fig. 1**). It was followed by a
118 period of subsidence visible on InSAR until mid-December with an average subsidence rate
119 of 0.2 mm/day and a total subsidence of 38 mm. The cumulative uplift observed at the centre
120 of displacement was slightly less than 150 mm and by December 2020 the actual uplift was
121 reduced to 90 mm.

122 **Free-air gravity changes**

123 Gravity provides additional insights, as changes in gravity depend on the mass of the intruded
124 fluids, while deformation depends on volume change. We measured gravity in four
125 consecutive campaigns (**Tab. S1**) at 10-12 existing permanent stations along an L-shaped
126 profile extending north and west from the centre of uplift (**Ext. data Fig. 4**).

127 The first campaign was conducted a week after the start of the first uplift episode and
128 repeated shortly after the end of the second one. Measured gravity changes between the first
129 and second campaign show a consistent free-air corrected gravity increase of 10-14 μGal
130 around the centre of uplift (**Ext. data Fig. 5 & Tab. S1**). A consistent free-air corrected gravity
131 decrease of 4-8 μGal is observed at the uplift centre from April to October, during the third
132 uplift episode and the following deflation. Between October 2020 and February 2021, the free-
133 air corrected gravity decrease continues at most stations. The maximum decrease averaged
134 over the three closest stations to the centre of uplift was 5 μGal .

135 To estimate the cumulative gravity increase, we need to account for the gravity changes
136 during the first week and during the third uplift episode that were not directly covered by the
137 gravity campaigns. This is done by assuming the same ratio of gravity increase to the uplift at
138 the inflation centre (170 $\mu\text{Gal}/\text{m}$) as measured during the first two uplift episodes. In addition,
139 we corrected for the long-term background gravity decrease caused by the geothermal
140 production. The resulting cumulative gravity increase of all three inflation episodes equals to

141 27 μGal at the centre of uplift. However, 13 months after the start of the unrest period, the
142 corrected net free-air gravity change was reduced to 14 μGal , implying that almost half of the
143 intruded mass had disappeared or migrated away from the inflation centre.

144 **Seismicity**

145 To evaluate the background seismicity and the rate changes during the transient deformation,
146 we used the national catalogue²⁰ of the Iceland Met Office (IMO). For a more detailed analysis
147 of the spatiotemporal patterns during the unrest period, we created a new catalogue for the
148 year 2020, using 26 seismic stations spaced up to 30 km around Svartsengi. For the first
149 time, we integrated distributed acoustic sensing (DAS) data from a 21-km-long fibre optic
150 telecommunication cable buried 80-90 cm below the ground from the southern tip of
151 Reykjanes to Grindavík, crossing the Svartsengi HT field²¹ (**Ext. data Fig. 4**). To detect and
152 locate the smallest earthquakes, we modified a waveform stacking and migration method²² to
153 combine seismic and continuous DAS data. The new and more complete catalogue covers
154 the period from February 1st to August 30th. We detected 39,500 earthquakes with magnitudes
155 of $M > -1$, i.e., a factor of 1.9 more events, and localised the majority automatically. The
156 locations have high quality, both laterally and vertically, since sensors and DAS cable were
157 located directly above the uplift zone, and the azimuthal distribution of all stations was
158 unusually good.

159 Seismicity shallower than 4 km depth occurs mainly within the elliptical uplift region, with the
160 shallowest events at the centre of uplift (**Fig. 3b & Ext. data Fig. 6**). This suggests that many
161 of these earthquakes are triggered by elastic bending stresses in the roof above the aquifer. .
162 The absence of earthquakes deeper than 4 km depth at the uplift centre supports the
163 hypothesis of an updoming BDT rising from 6-7 km depth (**Fig. 3**).

164

165

166 **Models of pre-eruptive processes**

167 The deformation pattern suggests an inflation source at depth. We first tested an isotropic
168 point-source (Mogi) model^{23,24} to retrieve information about the source location, depth and
169 strength. For the three uplift episodes, the best fitting parameters from inversion of the InSAR
170 data results in source depths ranging from 4.0 to 4.9 km with horizontal location difference
171 within 330 m. The inferred volume changes are estimated to be 4.70 , 3.55 and $2.75 \cdot 10^6 \text{ m}^3$,
172 respectively, and cumulatively explain over 90% of the observed deformation (**Suppl.**
173 **Information Figs. S1, S2 & Tab. S4**).

174 To better match the elliptical shape of the uplift pattern, we also tested rectangular dislocation
175 models²⁵. The uplift episodes can be modelled by three 10 m thick, 7-9 km long and 30-50 m
176 wide nearly horizontal intrusions at a depth between 3.7-4.4 km, with an average strike of
177 $N60^\circ E$. However, the subsidence rate after each uplift episode seems too fast and large (**Fig.**
178 **2c**) to be explained by a cooling-related volume reduction of a magma intrusion. Similarly, it is
179 difficult to explain the significant decrease of the free-air gravity observed in the central area
180 following the uplift episodes. Furthermore, the induced change in Coulomb failure stress is
181 concentrated directly around the presumed magma body and therefore cannot explain the
182 triggered seismicity, which has been observed over a wider area (**Fig. 4**). Anomalously low
183 V_p/V_s ratios are also observed in the lower crust beneath Svartsengi^{26,27}, indicating gas-
184 saturated porous rocks rather than partial melt. Analysis of ambient seismic noise prior to the
185 unrest²⁸ indicates a drop in S-wave velocities beneath Svartsengi, while a 1% temporary drop
186 in seismic velocities is also reported during the unrest period at Svartsengi²⁹. In both cases,
187 the observations can be explained by spatial or temporary changes in crack density without a
188 need for magma involvement.

189 As an alternative, we employ a poroelastic model considering a strongly coupled diffusion and
190 deformation process³⁰. The domain consists of a thin, permeable, porous aquifer layer at a

191 depth of approximately 4 km, embedded in a multi-layered poroelastic half-space (**Ext. data**
192 **Fig. 7**). The elastic parameters are adopted from the seismic reference model used for
193 earthquake locations (**Suppl. Information Tab. S3**). During the three uplift episodes, the
194 aquifer is pressurized by a fluid intrusion (injection) along a N60°E trending line source with
195 the strength of the fluid intrusion modelled by a Gaussian distribution with a standard
196 deviation of 2 km We inverted the InSAR data for the 12-months period of unrest, with fluid
197 inflow during uplift and no inflow during subsidence (**Fig. 5b**). The resulting intrusion rates
198 depend on the Skempton coefficient (B) used to describe how much of intruded fluid can be
199 accommodated by the aquifer through a given pressurization. The smaller B, the more the
200 intruded fluid volume. Based on previous interdisciplinary observations including realistic
201 estimate of temperature, pressure and porosity within the aquifer, we estimate $0.08 \leq B \leq 0.19$
202 as a realistic variation range for the assumed aquifer. Uncertainties in other parameters on
203 the volume and density estimations are negligible compared to the effect of the uncertainty in
204 the Skempton coefficient. (**Suppl. Information sheet IS3**). Accordingly, the total intrusion
205 volume can be estimated realistically to be 0.07- 0.16 km³.

206 The fluid intrusion causes poroelastic response resulting in surface deformation, fluid flow in
207 the aquifer, and the Coulomb stress changes on the fault planes. It explains the InSAR time
208 series, particularly the rapid subsidence following each cyclic uplift, the free-air gravity
209 changes and the seismicity variations in time and space.

210 The total mass of the intruded fluid, which best fits the campaign gravity data (**Fig. 2d and**
211 **Ext. data Fig. 5**), is about 82 Mt. It yields a fluid density in the range of 500-1200 kg/m³. We
212 estimate the range of pressure and temperature within the aquifer to be $110 < P < 150$ MPa
213 and $350 < T < 600$ °C. These values imply that the intruded fluid could be water (≤ 670 kg/m³),
214 carbon dioxide (≤ 740 kg/m³), supercritical sulfur dioxide, or a mixture of gas and magma
215 (2700 kg/m³) with maximum 33% volume ratio of the magma (**Suppl. Information sheet IS3**).

216 Although re-injection of fluid from the power plant might comply with the density value, the re-
217 injection rates are far too small to explain the observed volume and gravity change. Since
218 both H₂O and SO₂ remain dissolved in magma until close to the surface³¹, carbon dioxide is
219 practically the only possible fluid candidate. Therefore, we conclude that the injected fluid is
220 either pure CO₂ or in a CO₂ mixture with up to 33% magma volume.

221 The poroelastic model also explains the seismic rate changes relative to the background
222 activity. Specifically, we analysed the seismicity rate changes during the deformation, relative
223 to the average background rate between the years 2000 and 2020 in the region surrounding
224 the deformation centre. The observed seismicity rate changes were compared to those
225 predicted by a rate-and-state friction response³² based on the Coulomb failure stress changes
226 calculated from the poroelastic aquifer model. The predicted seismicity changes (**Fig. 4**) are
227 in good agreement with the observed spatiotemporal earthquake patterns, providing further
228 evidence for the poroelastic aquifer model.

229 We propose a feeder channel model that is consistent with the results from the poroelastic
230 model and fits the time scales of uplift and subsidence (**Fig. 5 & 6**). Finally, we propose a
231 conceptual model supported by geochemical observations of the erupted basaltic lava in
232 Fagradalsfjall (**Suppl. Information sheet IS2**).

233 It incorporates a magmatic reservoir containing crystal mush at 15-20 km depth, fed by slowly
234 upwelling currents of mantle derived magma (**Fig. 6 & Information sheet IS2**). We propose
235 that volatiles released from inflowing enriched magma into the sub-Moho reservoir migrated
236 upwards, possibly in fluid plumes modelled to explain the formation of porphyry-copper ore
237 shells³³. The volatiles were possibly trapped for weeks or months at the impermeable BDT at
238 ~7 km depth beneath Fagradalsfjall, generating overpressure, but not high enough to lift the
239 overburden (~220 MPa) and cause surface deformation. This might have triggered the
240 earthquake swarms that occurred within the brittle crust beneath Fagradalsfjall in 2017³⁴ and

241 late 2019. After reaching a certain limiting overpressure at the top of the fluid column, i.e.,
242 when a certain volume has accumulated, the magmatic volatiles were diverted upwards, just
243 below the BDT. The formed channel was episodically used by lithostatic pressurized fluid
244 batches to migrate toward the hydrostatic pressurized aquifer (~ 40 MPa) at 4 km depth at the
245 bottom of the convective HT field. They passed through the BDT and increased the pressure
246 sufficiently (>110 MPa) to cause the uplift.

247 The feeder channel model for the lower crust explains episodic inflow with exponentially
248 decreasing intensity and increasing recurrence times. Exponential decay is expected for
249 pressure-driven draining from a deep reservoir with finite volume³⁵ (see method section and
250 **Suppl. Information sheet IS1**). Episodic transport of fluid (CO₂) batches indicates a valve
251 mechanism in the channel, controlled e.g., by pressure-dependent permeability fluctuations³⁶
252 or strength/fracture toughness threshold mechanics³⁷. For B = 0.12, as an example, the three
253 constant intrusion rates would have been 17.1, 8.9 and 5.1 m³/s, respectively, yielding a total
254 intrusive volume of about 0.11 km³ (**Fig. 4b**). For smaller or larger B values, the intrusion
255 rates and volume can be scaled correspondingly. We infer that the first three fluid batches
256 were directed to the Svartsengi HT field, but that the fourth one in August was directed to the
257 Krýsuvík HT field where similar geothermal conditions are expected (**Fig. 6 & Ext. data Fig.**
258 **8**). Finally, in February 2021, the overpressure led to the fracturing of the brittle crust, forming
259 an ~8 km long magmatic dyke that ended with the Fagradalsfjall eruption in March.

260 Our model is consistent with the initial behaviour of the eruption that started March 19th, 2021.
261 In contrast to the violent initial phase of eruptions derived from a magma chamber, the
262 eruption rate was gentle during the first month. After that, it substantially increased with an
263 average production rate close to 11 m³/s based on the lava volume deposited at the surface³⁸.
264 Accounting for the porosity of the lava, it could correspond to 8-9 m³/s of magma outflow. In
265 our conceptual model, the volatiles from the degassing magma reservoir had already

266 detached from the initial magma batch prior to the eruption, rising from the upper mantle and
267 trapped at the BDT, where it eventually forced a different path through the ductile part of the
268 crust to the underpressurized geothermal reservoirs at the BDT. From the total mass of CO₂
269 injected into the roots of the HT fields, we can estimate that the minimum volume of degassed
270 magma beneath the present eruption site is at least of the order of 2-9 km³ (**Fig. 6 & Suppl.**
271 **Information sheet IS2**). Consequently, the available volume of magma is neither a limiting
272 factor for the longevity of the eruption nor the erupted volume.
273 This model helps to explain the role and importance of HT geothermal fields in the
274 complicated mechanism leading to volcanic eruptions.

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289
290 **Author contributions statement:**

291 ÓGF Co-ordination, writing, analysis of gravity, seismic and geothermal data, modelling
292 concepts.

293 RW Numerical and theoretical modelling, development of porous media deformation
294 approach and gravity modelling.

295 GPH Gravity measurements, data acquisition DAS, field work management.

296 TD Coordination, writing, modelling concepts, analytical model of cyclic channel flow.

297 SH1 Statistical analysis and modelling of seismicity.

298 MV Analysis of InSAR data, Mogi model inversion.

299 VD Analysis of InSAR data.

300 SH2 DAS processing, development of the Lassie stacking approach, automatic earthquake
301 detection/location for the combined network.

302 MI DAS data conversion and processing, earthquake detection/location, code
303 development.

304 EÁG Seismicity analysis, incl. focal mechanisms, data acquisition DAS, manual location of
305 earthquakes from the enlarged network.

306 KÁ Seismicity analysis, seismic network, contribution to field activity.

307 ThÁ Seismicity analysis, incl. focal mechanisms.

308 JH Contributed data from Czech temporary network.

309 MM Supervision and quality control InSAR analysis at GFZ.

310 TW Coordination InSAR analysis, contribution to manuscript drafting and writing.

311 ER Contribution to manuscript review, internal review of channel flow and fluid transport
312 models.

313 PJ Concept DAS measurements and coordination DAS data acquisition, coordination of
314 fieldwork in MAGIC at GFZ.

315 CMK Support of MAGIC HART initiative at GFZ, coordination of DAS data processing and
316 storage, manuscript review.

317 CM Installed temp. seismic stations, field work coordination, conversion of seismic data,
318 coordination of analysis and writing team at GFZ.

319

320 **Competing interest statement**

321 The authors declare no competing interests

322

323 **Figure legends**

324 **Figure 1. Overview of the tectonics and seismicity of the Reykjanes Peninsula.** The red
325 and black fault lines denote postglacial volcanic eruption fissures and opening fissures,
326 respectively. Yellow dots show the seismicity from September 2019 – May 2021. Yellow lines
327 show the extent of the high temperature geothermal fields on the peninsula, according to
328 resistivity measurements¹. Only the Reykjanes and Svartsengi geothermal fields are currently
329 operated. The blue stars show the centres of uplift in Svartsengi and Krýsuvík and the red
330 star the 2021 eruption site at Fagradalsfjall. The dashed green line shows the NE-striking
331 magmatic dyke intrusion, based on InSAR analysis and seismicity. Main roads are in black,
332 and topography is indicated as coloured background. Main landmarks referenced in the text
333 are shown on the map.

334

335 **Figure 2. Results from the poroelastic model. a.** Map of maximum cumulative uplift from
336 January 6th to July 17th, 2020 (Day 192). The colour map shows the vertical component
337 calculated by combining the ascending and descending LOS displacements, in comparison
338 with the predicted uplift (dashed contour lines) based on the poroelastic model for the same
339 time period. The GNSS time series from the station marked by the white with black dot circle

340 was used for the InSAR validation. The centre of uplift is obtained from the Mogi source
341 inversion. The dashed line shows the cross-section profile in Fig. 3a. The location of the
342 current Fagradalsfjall eruption is shown as a red triangle. **b.** Temporal snapshots of surface
343 uplift derived from the InSAR data, compared with the model predictions. The double model
344 curves for each snapshot show the values along the major and minor axes of the elliptic uplift
345 pattern. **c.** Comparison of the predicted and measured ascending and descending LOS
346 displacements at the pixel nearest to the uplift centre. The gray bars show the observed
347 seismicity rates within 5 km distance from the centre. **d.** Free-air gravity anomalies (blue
348 diamonds) from campaign measurements between January 27th and April 20th, 2020,
349 compared with the model predictions (solid curves) along the major and minor axes of the
350 elliptic uplift pattern. Estimated error is $\pm 5 \mu\text{Gal}$.

351

352 **Figure 3. Sketch of fluid migration paths and aquifer location compared to observed**
353 **seismicity and uplift. a.** Maximum uplift (solid black line) and topography (dashed gray line)
354 along the profile in Fig. 2a ($x=0$ at Mt. Thorbjörn). **b.** Density of micro-earthquakes between
355 February and August 2020 (gridded file and colour scale) estimated in a 1 km wide band
356 along the profile. The inferred brittle-ductile transition is indicated by the red dashed line. The
357 manually relocated, largest events from the earthquake swarms from November (green cir-
358 cles) and December 2019 (gray circles) and at the beginning of unrest at Svartsengi (blue
359 circles) are shown. The position and source intensity of the aquifer model are indicated.

360

361 **Figure 4. Comparison of observed $M \geq 1$ seismicity (grey/black) with model predictions**
362 **(coloured) based on induced crustal stress changes related to the geodetically**
363 **inverted source model.**

364 **a.** Histogram of earthquake numbers in the first 250 days of 2020 as a function of longitude.
365 **b.** Map showing the model numbers of predicted earthquakes per km² as contour lines, while
366 points indicate the observed events for the same period. Seismic stations are marked with
367 triangles (blue = stations of the national seismic network in Iceland (SIL) operated by IMO,
368 yellow = temporary network deployed by GFZ, green = seismic network operated by the
369 Czech Academy of Sciences together with ÍSOR, pink = joint ÍSOR-IMO stations, and the
370 DAS cable with a bold, blue line). **c.** Corresponding histogram of cumulative numbers as a
371 function of latitude. **d.** Time-versus-longitude plot comparing the temporal evolution, where
372 colours refer to the model's event density and magnitude-scaled points to observed
373 earthquakes. **e.** Temporal evolution of the spatially integrated numbers, where stars mark the
374 occurrence times of the three largest earthquakes.

375

376 **Figure 5. Cyclic fluid transport model.** **a.** Sketch of the deep geodynamic structures and
377 the assumed magmatic and fluid reservoir discontinuously pressurizing the aquifer at 4 km
378 depth. See the methods section for description of parameters and additional explanation. **b.**
379 Model flow rate (red line) at entry point of transport channel to explain the time, duration, and
380 magnitude of injections at the aquifer level (black rectangles). **c.** Time function of the predict-
381 ed cumulative volume (red solid line) at the entry point of the discontinuous transport channel
382 / fault is compared to the discontinuous volumes (black line) arriving at the aquifer level at 4
383 km depth. Red squares indicate the onset of the new batches of magmatic fluids. The open
384 squares indicate the batch arriving at the Krýsuvík HT field in August 2020.

385

386 **Figure. 6. A conceptual model of the processes leading to the uplift, gravity changes,**
387 **seismic unrest, and the volcanic eruption.** The model connects the datasets observed and
388 the dynamic processes through modelling with mathematical and physical approaches.

389 Magmatic volatiles, mainly CO₂, and possibly magma migrate upwards from a magma
390 reservoir containing crystal mush at 15-20 km depth, or just beneath Moho. When the gas
391 pressure or volume has reached a critical value at the BDT, batches of gas are driven just
392 below the BDT towards the hydrostatic pressurized geothermal aquifers where the gas
393 intrudes and inflates the overburden. The process resembles those described by Fourier in
394 1999⁵⁶. In February 2021, the pressure at the BDT managed to cause enough deviatoric
395 stress to break through the brittle crust, forming a magmatic dyke intrusion and finally the
396 eruption. The suggested amount of magma within the magma reservoir is based on the
397 degassing process (Information sheet IS2). Vertical and horizontal scales are the same.

398

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503

504 **Methods**

505 **InSAR data processing**

506 We exploited the Copernicus Sentinel-1A and 1B satellite SAR data, which are available over
507 Iceland in ascending (08/01/-15/12/2020) and descending (06/01/-13/12/2020) orbits with a
508 revisit time of 6 days for each track. By calculating a single interferogram for the cumulative
509 period between 20/01/-17/07/2020, we found four fringes in both geometries, representing a
510 shift of approximately 11 cm in the line of sight (LOS) direction (**Ext. data Fig. 2**). A more
511 detailed InSAR time series analysis shows three inflation and deflation episodes (**Ext. data**
512 **Fig. 3**): first uplift 19/01/-18/02/2020, second uplift 07/03/-24/04/2020, and third uplift 18/05/-
513 17/07/2020. Due to the 6-day revisit time of Sentinel1A/1B acquisitions, the exact timing of
514 inflation/deflation episodes may be shifted by a few days. The time-series analysis was
515 carried out using the Small Baseline Subset (SBAS) algorithm³⁹, implemented in SARscape,
516 which is based on the combination of interferograms characterized by small, normal and
517 temporal baselines, maximizing spatial and temporal coherence and therefore the quality of
518 the interferograms (maximum temporal baseline 18 days), resampling with a multi-look of 7:2
519 for range and azimuth direction, respectively (Goldstein filter⁴⁰ window size of 32 pixels,
520 coherence threshold of 0.2, ground resolution of 30 m). The topographic phase contribution
521 was subtracted using the ArcticDEM digital terrain model of 2 m spatial resolution. InSAR
522 time-series were compared to data from a GNSS station provided by the IMO and available
523 through the University of Iceland⁴¹, showing good agreement for both ascending and
524 descending datasets (**Ext. data Figs. 3a & 3b**). Ascending and descending LOS

525 displacement maps were combined to derive the vertical and the E-W components of the
526 ground motion (**Ext. data Figs. 3c & 3d**).

527

528 **Gravity data processing**

529 Each of the four gravity campaigns (**Ext. data Fig. 5 & Tab. S1**) lasted for a few days.

530 Elevation and gravity at the second to northernmost station (HS22) were used as a reference
531 (**Ext. data Fig. 4**), but free-air corrected by up to 2.6 μGal to account for small elevation
532 changes at this site between campaigns. The gravity value at the northernmost and
533 westernmost stations (HS16 and SNH25) were less constrained or had difficult local
534 conditions due to snow and ice and were therefore regarded as unreliable. Finally, the gravity
535 data were corrected for tidal and latitude effects and free-air gravity computed using elevation
536 changes obtained from InSAR measurements. The data were not corrected for ocean tides,
537 as these effects can be considered negligible.

538 The uncertainty in similar gravity campaigns is generally 10-15 μGal for individual data points.
539 However, these error limits are highly dependent on external conditions such as weather prior
540 to and during the measurements, local site conditions, nearby anthropogenic activity, or
541 seismic noise. We selected only quiet and calm days to minimize the measurement
542 uncertainty. Our dataset provides consistent maximum values close to the centre of uplift
543 which decrease with distance from the centre. However, a few data points are outliers, both
544 with respect to the general trend with time and the nearby stations. Apart from obvious
545 outliers, we estimate, based on the internal consistency of the data, that the error limits are
546 close to $\pm 5 \mu\text{Gal}$.

547 We corrected for long-term background gravity reduction caused by the net production of
548 geothermal fluid at Svartsengi⁴². The ratio of the total gravity decrease from 1976 to 2014 to

549 the corresponding net mass production gives a gravity decrease of $0.67 \cdot 10^{-9} \mu\text{Gal}/\text{kg}$. By
550 applying this to the estimated production of 2020 we get a value of $5.1 \mu\text{Gal}/\text{year}$ at the centre
551 of uplift.

552

553 **Joint DAS and seismic network processing**

554 The DAS strain-rate data were recorded using a Silixa iDAS (version 2) interrogator with a
555 sampling frequency of 1 kHz, a gauge length of 10 m and a spatial channel offset of 4 m
556 along the full length of the fibre. The DAS data have been included as 40 virtual single-
557 component stations by extracting the DAS signal at regular 500 m intervals along the fibre.
558 For each virtual station, the signal was integrated over 36 m (9 channels) along the cable to
559 improve the signal to noise ratio.

560 The employed automatic detection method⁴³ is based on an image function (IF) computed for
561 a grid of potential source positions and times. The IF is computed from the stacking of time-
562 back-shifted waveform attributes of P- and S-phases using the local velocity model (**Tab. S3**).
563 Waveform attributes of P-phases employ a smoothed STA/LTA function⁴³ calculated from
564 filtered seismograms (3-25 Hz). S-phase attributes are calculated from a smoothed squared
565 signal, which is more sensitive to high amplitudes. Smoothing causes the detector to be more
566 robust against errors in the assumed seismic velocities and reduces the computational cost.
567 Waveform attributes are normalized by a moving average with a duration longer than a typical
568 transient event. Seismic events appear as spatio-temporal peaks in the 4D image function. A
569 detection is registered when the IF exceeds a certain threshold value. The position of the
570 spatio-temporal peak is used to provide the origin time and location. The detector is
571 implemented in the Lassie software package, distributed under the Pyrocko framework⁴⁴
572 (**Tab. S2**).

573 **Poroelastic modelling**

574 The poroelastic model uses the semi-analytical tool POEL³⁰, which simulates strongly coupled
575 diffusion-deformation processes induced by injection (pumping) tests. Based on Biot's
576 poroelasticity theory, a poroelastic medium can be defined with 5 parameters; the shear
577 rigidity μ , the drained Poisson's ratio ν , the undrained Poisson's ratio ν_u , the Skempton
578 coefficient B and the hydraulic diffusivity D . We adopt the seismic reference model SIL
579 **(Suppl. Information Tab. S3)**⁴⁵. As the diffusion effect might be negligible in the seismic
580 frequency band, μ and ν_u can be derived simply from the density and seismic velocities
581 **(Suppl. Information Tab. S3)**. Except for a thin permeable aquifer layer representing the
582 medium saturated with compressible fluid, which is inserted into the layered half-space at the
583 depth z , all other layers are assumed to be nearly impermeable ($D \rightarrow 0$). We choose a
584 standard and uniform value for the drained Poisson's ratio ($\nu = 0.25$) of the whole layered
585 structure including the aquifer layer. Note that no diffusion process can happen in the
586 impermeable layers. Therefore, only the elastic property represented by the shear modulus
587 and the undrained Poisson's ratio is relevant for these layers.

588 Finally, there are only 4 free parameters (z , μ , B and D) characterizing the aquifer layer. We
589 optimize all these parameters so that the predicted surface deformation best fits the InSAR
590 data. Note that the thickness of the aquifer (here 10 m) is rather arbitrary, as the surface
591 deformation is not independently influenced by the thickness and hydraulic diffusivity, but by
592 their product determining the aquifer transmissivity.

593 The source is represented by episodic fluid injections into the thin and vertically confined
594 aquifer. The central location of injection coincides with the maximum surface uplift at
595 63.870°N/22.465°W. Each of the 3 uplift episodes is attributed to a constant injection rate.
596 The timings of injection are estimated directly by the uplift-subsidence turning points of the

597 InSAR time series with a resolution of 6 days; Episode 1 from day 12 to 42, Episode 2 from
598 day 60 to 108, and Episode 3 from day 132 to 192 in 2020. The surface uplift exhibits a
599 slightly elliptic pattern, which can be interpreted as preferential diffusion of intruded fluid along
600 the main strike of fracture systems under the Svartsengi HT field (**Ext. data Fig. 1**). For this
601 purpose, we use a Gaussian line source instead of a point source. The source line orientation
602 and its characteristic length are optimized, too. Note that the surface deformation is related
603 linearly with the 3 constant intrusion rates, but non-linearly with the aquifer parameters and
604 the source geometry. We optimize all these nonlinear parameters by a trial-and-error
605 approach. For each candidate set of nonlinear parameters, we estimate the 3 intrusion rates
606 by the least-squares method. The final optimal model has the minimum misfit between the
607 predicted and observed surface deformation derived from 139 pairs of ascending and
608 descending InSAR images.

609 An optimal choice of the parameter set is given by $z = 4.0$ km, $\mu = 7.55$ GPa, and $D = 0.10$
610 m^2/s (for the aquifer thickness fixed at 10 m). The line source has a characteristic length of
611 4.0 km (two sigma of the Gaussian distribution) and is oriented to N60°E. The normalized
612 variance of misfits between the predicted and observed InSAR data is about 5%. The surface
613 subsidence after each uplift episode is successfully explained by relaxation of the intrusion-
614 induced overpressure through fluid diffusion within the aquifer.

615 In comparison, the aquifer depth and the source geometry are well resolved and consistent
616 with those obtained from the Mogi and rectangle dislocation models. However, there is a
617 certain trade-off between μ and D . The InSAR data can be explained equally well using any
618 combination of the two parameters within a roughly elliptical trade-off zone covering about
619 $0.1 \leq \mu/\mu_s \leq 0.8$ and $0.03 \leq D \leq 0.12$ m^2/s (**Suppl. Information sheet Fig. IS3-1**), where
620 $\mu_s = 15.1$ GPa is the shear modulus of the top elastic layer above the aquifer. The estimated

621 fluid intrusion volume is mainly controlled by B , but it can also be slightly affected by about
622 $\pm 5\%$ through different choices of the μ - D combination, e.g., $40 \pm 2 \times 10^6 \text{ m}^3$ for $B = 0.3$ (**Suppl.**
623 **Information sheet Fig. IS3-2**). In fact, when fixing all other parameters, the same surface
624 deformation can be produced using different B values if the intruded volume remains
625 proportional to B^{-1} .

626 For the quasi-static poroelastic deformation, only the volume of intruded fluid, rather than its
627 mass, is relevant. The latter can only be constrained by the gravity data in the present study.
628 It is known that the internal deformation has no effect on the free-air gravity change for
629 spherical pressure sources in a homogeneous half-space⁴⁶. Based on this, and on predictions
630 from superposition of spherical pores⁴⁷, we infer that the effect of internal deformation is still
631 negligible in comparison to the primary effect of the intruded mass on the free-air gravity.
632 Thus, we use the POEL output for Darcy flux to calculate the spatio-temporal redistribution of
633 the intruded mass within the aquifer and its contribution to the free-air gravity change on the
634 surface. A simple comparison between the predicted and observed free-air gravity anomalies
635 yields an estimate of intruded fluid mass of about 82 Mt. The misfits of the model predictions
636 to the gravity data are all within the measurement uncertainties of about $\pm 5 \mu\text{Gal}$ (**Ext. data**
637 **Fig. 5**). In the present case, a realistic estimate of the Skempton coefficient is given by $0.08 \leq$
638 $B \leq 0.19$ (**Ext. data Fig. 9 & Suppl. Information sheet IS3**). Accordingly, the intruded volume
639 varying between 0.068 - 0.161 km^3 , the density of the intruded fluid is estimated to be between
640 500 - 1200 kg/m^3 .

641

642 **Seismicity modelling**

643 We applied the Coulomb-Rate-and-State model, widely used for natural, anthropogenic, and
644 volcanic activity^{32,48,49,50,51}. In this model, the earthquake rate as a function of space \vec{x} and
645 time t is given by:

$$646 \quad R(\vec{x}, t) = \frac{r(\vec{x})}{\dot{\tau}\gamma(\vec{x}, t)},$$

647 where r is the background seismicity rate, $\dot{\tau}$ is the background shear stressing rate, and γ is a
648 state variable. The evolution of γ depends on the stress history at the given location \vec{x} , the
649 background stressing rate $\dot{\tau}$, and the product $A \cdot \sigma$, where A is a constitutive parameter related
650 to the direct effect in the laboratory-derived rate-and-state friction law and σ is the effective
651 normal stress acting on the faults. Since neither A nor σ are well-constrained for natural faults,
652 the product $A \cdot \sigma$ is usually considered as a free parameter. Fits to various seismic sequences
653 yielded $A \cdot \sigma$ -values between 0.01 and 0.1 MPa⁵¹.

654 We estimated the CFS stress history by $-\tau_{max}(\vec{x}, t) + f \cdot p(\vec{x}, t)$ ², where f is the friction
655 coefficient typically in the range from 0.1 to 0.8, and τ_{max} and p are the induced maximum
656 shear and pore pressure. Here, we use the maximum shear value to account for variable
657 receiver mechanisms of the small to moderate magnitude earthquakes. We find that f only
658 slightly affects the results and set it to $f=0.7$. Based on the geodetically constrained
659 poroelastic model described above, we calculated CFS using the POEL code at grid points \vec{x}
660 in a spatial box with a dimension of 30 km in the NS- and 60 km in the EW-direction, centred
661 at the deformation source, and at six depth levels up to 3 km, with a grid spacing of 0.5 km.

662 The background rate $r(\vec{x})$ and the background stressing rate $\dot{\tau}$ were estimated by preceding
663 seismicity. For that purpose, we analysed the earthquakes in the IMO-catalogue that occurred
664 in this region between 2000 and the end of 2019. The catalogue is complete for $M \geq 1.5$ during
665 this period, while it is complete for $M \geq 1.0$ in 2020. The rate of $M \geq 1.5$ background events in the
666 selected region is found to be $r = 0.76 \text{ days}^{-1}$, which translates to $r = 2.41 \text{ } M \geq 1 \text{ events days}^{-1}$

667 assuming a Gutenberg-Richter magnitude distribution with $b=1$. The spatial distribution of $r(\vec{x})$
668 is estimated by laterally smoothing the epicentres using a Gaussian filter with a standard
669 deviation of 1 km and assuming a uniform distribution at depth. The uniform stressing rate $\dot{\tau}$ is
670 estimated using Kostrov's formula for our selected region^{52,53}, assuming a maximum
671 earthquake magnitude of M6 and an effective thickness of the seismogenic zone of 4 km,
672 resulting in $\dot{\tau}= 12$ Pa/day.

673 The single free model parameter $A \cdot \sigma$, determining the strength of the seismic activation due to
674 the stress changes, is set by matching the total observed number of earthquakes, yielding $A \cdot \sigma$
675 = 0.02 MPa, which is well within the range of previously observed values. The spatio-temporal
676 fits of the final model are presented in **Fig. 3** for earthquakes above the completeness
677 magnitude of M1.0 in 2020.

678 **Cyclic fluid transport modelling of magmatic volatiles**

679 The buoyancy-driven drainage of a deep reservoir by viscous channel flow resembles gravity-
680 driven outflow from a tank through a pipe. If viscosity is dominant, the flux follows an
681 exponential decay with $q(t) = q_0 e^{-t/t_0}$, where q_0 is the initial flux at time $t=0$ with $t_0 =$
682 $\pi \cdot r^2 \cdot h_2(t=0)/q_0$ (**Fig. 5a & Suppl. Information sheet IS1**). The ascent height h or the length L
683 of the inclined channel do not enter the problem. The modelling of the total influx after three
684 uplift episodes leads to a draining decay time of $t_0 = 95$ days and an initial flux of $q_0 = 4.1$ m³/s
685 (**Fig. 5b**).

686 The cyclic ascent of finite-volume fluid batches can be explained by buoyancy-driven fracture
687 movement. The fluid flux rate, $q(t)$, controls the filling of batches. The ascent of a fracture
688 through a channel and the detachment from the feeding system below is controlled by the
689 stress intensity at the upper and lower tip of the fracture⁵⁴. The intrusion batches arrive at the
690 sealed aquifer at 4 km depth with estimated volumes of $\sim 45 \cdot 10^6$ m³, $\sim 38 \cdot 10^6$ m³ and $\sim 27 \cdot 10^6$

691 m³ for the first, second, and third batch, respectively (**Fig. 5b**). The ascent velocity of the fluid
692 batch depends on several factors, including the properties of the channel, the alteration of
693 rocks³⁶ and the enclosed fluid volume controlling the internal overpressure⁵⁴. A constant
694 ascent velocity is supported by laboratory experiments of finite volume fluid ascent in
695 gelatine⁵⁴.

696 The fracture model explains the injected volumes and the onset and inter-event time of the
697 three uplift cycles (**Fig. 5b & c**). The fit has been obtained by matching the accumulated
698 volume at the end of phase 3, $\Sigma V = V_1 + V_2 + V_3$, and the onset of the first injection phase. The
699 model is also consistent with the onsets of uplift phases 2 and 3 and perhaps a fourth phase
700 at Krýsuvík in August 2020 (**Figs. 5c & 6, Ext. data Fig. 8**). The model cannot resolve the
701 ascent times of the fluid batch to the injection point at the aquifer at 4 km depth. The onset of
702 an uplift phase in **Fig. 5b** and **c** represents the arrival time at 4 km depth. In contrast, the
703 filling of the first fluid batch is estimated for the December 18th, 2019 (**Fig. 5c**), which
704 correlates with the occurrence of the strong earthquake swarm on December 15th-20th, 2019,
705 at 5-6 km depth and about 7 km east of the centre of uplift. We speculate that the inflow
706 traversed from this location. This would also link the unrest beneath Svartsengi to the eruption
707 that started in March 2021 just above the location of the December 2019 earthquake swarms
708 (**Ext. data Fig. 6**). The increasing duration of the injection into the aquifer may imply that the
709 pressure difference between the injection source and the sealed aquifer decreased with the
710 arrival of each new over-pressurized batch.

711

712 **Data availability**

713 The basic data used in this paper consist of InSAR, seismic and gravity data.

714

715 InSAR data: The radar data acquired by the Copernicus Sentinel-1A and 1B satellites are
716 available at no cost from the European Space Agency's Copernicus Open Access Hub
717 (<https://scihub.copernicus.eu/>) and can be interferometrically processed using the freely
718 available Sentinel 1 toolbox (<https://step.esa.int/main/toolboxes/sentinel-1-toolbox/>). The
719 InSAR data generated from the Copernicus Sentinel-1A and 1B satellites and as published in
720 the paper are available on Zenodo,

721 <https://zenodo.org/record/6344933>, or <https://doi.org/10.5281/zenodo.6344933>

722

723 Seismic data: The earthquake catalogue based on waveform stacking from DAS and local
724 seismic stations presented in this paper is available on Zenodo,

725 <https://zenodo.org/record/6337788> , DOI: <https://doi.org/10.5281/zenodo.6337788> .

726 Seismic data from the permanent national seismic network in Iceland are available from the
727 open database of the Iceland Met Office, <https://skjalftalisa.vedur.is>

728 For data from temporary stations from the Czech Academy of Sciences, Czech Republic,
729 restrictions apply to the availability, which were used under licence for the current study, and
730 so are not publicly available. These data are however available from the authors upon
731 reasonable request and with permission of the Czech Academy of Sciences.

732 The seismological data from temporary GFZ stations analysed during the current study are
733 available in the GEOFON repository at <https://geofon.gfz-potsdam.de/waveform/archive> under
734 code 9H and Dahm, T., Jousset, P., Heimann, S., Milkereit, C., Hersir, G.P., Magnússon, R.
735 (2020): MAGIC - Seismic network MAGma in Iceland,
736 <https://doi.org/10.14470/4U7575229166> (restricted until 01/2026).

737 Seismological waveform data from the glass-fibre cable monitoring in SW Reykjanes

738 Peninsula are available from the GEOFON data centre, under network code 5J. This dataset

739 comprises a selection of waveforms recorded along an optical fibre of 21 km length. The
740 subset consists of 40 channels at 100 Hz (spatially stacked 9x). Specific full data set is
741 available upon request to the authors. Jousset, P., Hersir, G.P., Krawczyk, C., Wollin, C.,
742 Lipus, M., Reinsch, T., Isken, M., Heimann, S. (2020): MAGIC (MAGma in Iceland). GFZ Data
743 Services. Other/Seismic Network. DOI: <https://doi.org/10.14470/0W7575244885> .

744 The preliminary seismic catalogue used in this study has been derived by combining
745 horizontal strain DAS and local seismic stations in a migration-based detection and location
746 approach using the open-source Lassie software (<https://git.pyrocko.org/pyrocko/lassie/>).

747
748 The gravity data are available on Zenodo: <https://zenodo.org/record/6344613>
749 and DOI 10.5281/zenodo.6344613

750
751 All other data generated or analysed during this study are included in this published article
752 and its supplementary information files in form of tables, or available from the corresponding
753 author on reasonable request.

754 All general request regarding the paper, the geological interpretation, the conceptual model,
755 and the gravity data should be directed to the corresponding author, Ólafur G. Flóvenz, ÍSOR,
756 Iceland GeoSurvey, ogf@isor.is

757 Requests regarding other specific parts of the paper should be directed as follows:

758 Seismic data processing and interpretation: Sebastian Hainzl, hainzl@gfz-potsdam.de

759 Poroelastic modelling: Rongjiang Wang, wang@gfz-potsdam.de

760 InSAR data processing and modelling: Thomas Walter, twalter@gfz-potsdam.de

761 Cyclic fluid transport modelling: Torsten Dahm, dahm@gfz-potsdam.de

762

763 **Code availability**

764 The seismic recordings were processed with Pyrocko⁴⁴. The software package Lassie
765 (<https://git.pyrocko.org/pyrocko/lassie.git>) was used for detection and localisation of earth-
766 quakes.

767 Recorded DAS data was converted using iDAS Convert (Isken, M., Wollin, C., Heimann, S.,
768 Quinteros, J., Jäckel, K.-H., Jousset, P. (2021). DAS Convert - Convert distributed acoustic
769 sensing data. V. 1.0. GFZ Data Services. <https://doi.org/10.5880/GFZ.2.1.2021.005>). The
770 seismic software frameworks are publicly available at <https://pyrocko.org>. The seismic
771 software code is open, and methods have been reviewed publicly.

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