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1 **Evolution of deformation and stress changes during the caldera**
2 **collapse and dyking at Bárðarbunga, 2014-2015: Implication for**
3 **triggering of seismicity at nearby Tungnafellsjökull volcano**

4

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19 **Highlights**

- 20 • Constrained multi-source model reproduces both the near- and far-field
21 volcanic deformation
- 22 • Temporal evolution of source parameters determined throughout unrest and
23 eruption

24 • Bárðarbungna caldera collapse triggers earthquakes at nearby
25 Tungnafellsjökull volcano

26

27 **Abstract**

28 Stress transfer associated with an earthquake, which may result in the seismic
29 triggering of aftershocks (earthquake-earthquake interactions) and/or increased
30 volcanic activity (earthquake-volcano interactions), is a well-documented
31 phenomenon. However limited studies have been undertaken concerning volcanic
32 triggering of activity at neighbouring volcanoes (volcano-volcano interactions). Here
33 we present new deformation and stress modelling results utilising a wealth of diverse
34 geodetic observations acquired during the 2014-2015 unrest and eruption within the
35 Bárðarbunga volcanic system. These comprise a combination of InSAR, GPS,
36 LiDAR, radar profiling and optical satellite measurements. We find a strong
37 correlation between the locations of increased seismicity at nearby Tungnafellsjökull
38 volcano and regions of increased tensile and Coulomb stress changes. Our results
39 suggest that stress transfer during this major event has resulted in earthquake
40 triggering at the neighbouring Tungnafellsjökull volcano by unclamping faults within
41 the associated fissure swarm. This work has immediate application to volcano
42 monitoring; to distinguish the difference between stress transfer and new intrusive
43 activity.

44

45 **Keywords**

46 Joint inversion of geodetic observations; Holuhraun eruption; Bárðarbunga;
47 Tungnafellsjökull; stress triggering; subsurface fault mapping.

48 **1 Introduction**

49 An earthquake can increase rates of seismicity on surrounding faults by altering the
50 shear and normal stress fields (e.g., King et al, 1994; Stein et al., 1992; Harris and
51 Simpson, 1992) via both increased shear and unclamping, which are related to
52 changes in Coulomb failure stress (Stein, 1999). Abrupt changes in pressure within a
53 magmatic system may also be induced by nearby earthquakes, ultimately leading to
54 an eruption (e.g. Walter and Amelung, 2007; Manga and Brodsky, 2006; Hill et al.,
55 2002; Marzocchi et al., 2002; Linde and Sacks, 1998). Conversely, volcanic activity
56 may trigger earthquakes on near-by faults (e.g. Hill et al., 2002; Roman and Heron,
57 2007; Jónsson, 2009; Green et al., 2015; Heimisson et al., 2015). Numerous studies
58 have been undertaken on stress transfer during earthquake-earthquake interactions
59 (e.g. Freed 2005; Árnadóttir et al., 2003; Stein, 1999 and references therein) along
60 with earthquake-volcano interactions (e.g, Walter and Amelung, 2007; Parsons et al.,
61 2006; Díez et al., 2005; Toda et al., 2005; La Femina et al., 2004; Hill et al., 2002;
62 Nostro et al., 1998). However, fewer studies have investigated volcanic stress
63 triggering at neighbouring volcanoes (e.g., Albino and Sigmundsson, 2014;
64 Gonnermann et al., 2012).

65 In this work we study the interaction between two closely spaced (25 km)
66 central volcanoes in Iceland, Bárðarbunga and Tungnafellsjökull. The volcanoes are
67 very different. Bárðarbunga is one of the most powerful volcanoes in Iceland, with at
68 least 26 eruptions in the last 1100 years, including four where the magma volume was
69 between 1 and 4 km³. Its associated fissure swarms extend 115 km to the southwest
70 and at least 55 km to the north of the volcano (Larsen and Gudmundsson, 2015;
71 Hjartardóttir et al., 2016). Tungnafellsjökull, on the other hand, has only had two
72 minor eruptions in the Holocene, and its fissure swarms are immature, extending only

73 25 km to the southwest and 15 km to the north of the volcano (Einarsson, 2015;
74 Björnsdóttir and Einarsson, 2013). Compared to the Bárðarbunga volcano,
75 Tungnafellsjökull has not been very seismically active over the last 26 years of digital
76 seismic recording in Iceland, with a yearly average of two recorded earthquakes of
77 $M \geq 2$. However, seismicity at Tungnafellsjökull increased at the same time as major
78 unrest, dyke propagation, slow caldera collapse and eruption occurred within the
79 Bárðarbunga volcanic system, Iceland (Figure 1). We can gain insight into volcano
80 interactions during these events by the joint analysis of earthquake and ground
81 deformation data.

82 Major activity in the Bárðarbunga volcanic system was initially identified by
83 the onset of an intense earthquake swarm on the 16th August 2014 and concurrent
84 movement registered at several nearby continuous GPS (cGPS) stations. Over the
85 following weeks additional cGPS stations were installed, campaign GPS sites were
86 reoccupied and interferograms formed using X-band and C-band satellite images.
87 Data were analysed in near real-time and used to map ground displacements
88 associated with the initial dyke emplacement and propagation (northeast of
89 Bárðarbunga caldera), responsible for the sudden unrest (Sigmundsson et al 2015).
90 The main fissure eruption commenced on the 31st August, characterised by lava
91 fountaining and the extrusion of extensive lava flows. In addition to the lava effusion,
92 a slow collapse of the Bárðarbunga caldera began on the 20th August 2014
93 (Gudmundsson et al., 2016) and continued throughout the eruption, resulting in a
94 maximum subsidence of 65 m by the end of the eruption on the 27th February 2015.
95 Shortly after the onset of unrest at Bárðarbunga (within 24 hours), an increase in
96 seismicity was observed at the neighbouring Tungnafellsjökull volcano situated ~25
97 km northwest of the Bárðarbunga caldera (Figures 1 and 2). The seismicity however

98 abruptly increased during unrest in Bárðarbunga; to 10 times the average in 1996 and
99 to 60 times the average in 2014.

100 Possible mechanisms for increased seismicity at Tungnafellsjökull during the
101 2014-2015 Bárðarbunga unrest and eruption include:

- 102 1) A new intrusion beneath Tungnafellsjökull volcano,
- 103 2) Pressure change within a resident magma body beneath Tungnafellsjökull,
- 104 3) Dynamic stress transfer related to the passing of seismic waves from large
105 Bárðarbunga earthquakes,
- 106 4) Static stress transfer related the Bárðarbunga unrest and eruption.

107 Activity in the magmatic system of the receiver volcano (Tungnafellsjökull)
108 would be responsible for possibilities (1) and (2), whereas the origin of the seismicity
109 would be related to activity in the source volcano (Bárðarbunga) for cases (3) and (4).
110 In order to consider case (4), three distinct deformation processes need to be
111 considered in the Bárðarbunga volcanic system: i) dyke propagation from
112 Bárðarbunga caldera to the Holuhraun eruption site (deformation associated with this
113 emplacement continued from 16th August 2014 to mid September 2014), ii) magma
114 withdrawal from a reservoir $\sim 12 \pm 4$ km beneath the caldera (Gudmundsson et al.,
115 2016) and iii) slow caldera collapse. Processes ii) and iii) continued throughout the
116 eruption.

117 Previous deformation and stress modelling studies have been undertaken in the
118 vicinity of Bárðarbunga during the 2014-2015 activity. Sigmundsson et al. (2015)
119 originally modelled the complex deformation field, associated with the lateral dyking
120 event and the associated caldera collapse, during the initial period of activity from
121 16th August to 6th September 2014. A Markov chain Monte Carlo (MCMC) approach
122 was used to estimate the multivariate probability distribution of opening and shearing

123 on a segmented dyke, extending from the southeast edge of the caldera to the eruption
124 site, slip on two caldera faults, and deflation of either a spherical or flat-topped
125 magma chamber at depths of ~1.4 and 3.5 km BSL respectively. Riel et al. (2015),
126 modelled the caldera subsidence utilising InSAR observations spanning the first few
127 weeks of the eruption. The magma chamber was modelled as a deflating horizontal
128 circular crack in an elastic half-space. The authors concluded that one of the most
129 likely physical processes responsible for the large number of CLVD (compensated
130 linear vector dipole) dominant M5 earthquakes accompanying the collapse, was slip
131 on caldera ring faults. They compared deformation associated with a M4.9
132 earthquake, in the northern sector of the caldera on the 14th September (derived from
133 1-day interferograms), with a static model of slip on a 1 km wide ring fault segment at
134 a depth of 2 km. Green et al. (2015) investigated the static stress changes associated
135 with dyke emplacement and deflation of a source at 17 km BSL beneath Bárðarbunga
136 caldera. They demonstrated a stress shadow effect within regions to the north and east
137 of the volcano. More recently Gudmundsson et al. (2016), used a combination of
138 distinct element method numerical modelling and seismic data to run a series of
139 forward simulations of the magma chamber and ring fault system, which were
140 compared to a NNW-SSE trending caldera subsidence profile (collected using radar
141 altimetry and corrected for ice-flow). They determined that the subsidence at depth
142 was controlled by the reactivation of steeply dipping ring faults.

143 Our study incorporates a similar ring fault and magma reservoir geometry as
144 Gudmundsson et al. (2016) combined with the dyke geometry from Sigmundsson et
145 al. (2015). We undertake a joint inversion of multiple geodetic datasets (caldera
146 subsidence grids, GPS and InSAR observations) using a Bayesian approach and
147 present a new constrained model of the subsurface processes within the Bárðarbunga

148 volcanic system. Our model reproduces the complete cumulative deformation field for
149 six individual time periods between 16th August 2014 and 10th April 2015 (Figure 2).
150 In addition, we generate a time series of cumulative stress changes throughout the
151 period of volcanic unrest and eruption, and compare our results to earthquake
152 relocations and mechanisms at the neighbouring Tungnafellsjökull volcano, to
153 ultimately determine the most likely cause of the increased seismicity in that area.

154 **2 Data and methods**

155 The input data to our modelling comprises a combination of various seismic and
156 geodetic observations. We utilise a more diverse dataset than those presented in
157 previous Bárðarbunga deformation studies. This consists of both near- and far-field
158 deformation measurements spanning an extended time period (16th August 2014 to
159 10th April 2015). These data include both campaign and GPS observations from 33
160 stations with good azimuthal coverage around Bárðarbunga (Figure 1 and Table S1 in
161 Supplementary Material), subsidence grids covering the caldera (derived from a
162 combination of LiDAR, radar altimetry and SPOT datasets), and X-band
163 interferograms (see Table S2 in Supplementary Material). Detailed descriptions of the
164 methods used to process these data are outlined in section S3 (Supplementary
165 Material). Information regarding the earthquake hypocenter relocations, deformation
166 modeling and stress modeling is included in the following section.

167 **2.1 Earthquake hypocenter relocations**

168 High-precision relocations of earthquakes within the Tungnafellsjökull central
169 volcano, recorded by the Icelandic national digital seismic network - South Iceland
170 Lowland (SIL) network - (Böðvarsson et al., 1996; 1999), were used to locate the
171 active areas within the volcano during the 2014-2015 Bárðarbunga unrest and

172 eruption, and where possible, to map the active faults and fractures and their slip
173 directions. Over 750 earthquakes in the Tungnafellsjökull area between 2008 and
174 March 2015 are relocated using the relative, double-difference method of Slunga et al.
175 (1995) and the standard SIL velocity model generally used for initial routine locations
176 of earthquakes in Iceland (Stefánsson et al., 1993).

177 The relocated event distribution displayed in Figure 1c shows that the activity is
178 mostly distributed in a southwest-northeast direction, extending from the central part
179 of the Tungnafellsjökull caldera to the area north of it and in a small cluster in
180 Vonarskard caldera. Almost 90% of events are at depths < 9 km, 66% at depths < 5
181 km. Source depths of event clusters appear to increase towards north; from a
182 maximum of 4 km in the south to 8.5 km in the north. The bulk of the activity at
183 Tungnafellsjökull also moves with time from south to north during the Bárðarbunga
184 unrest, while the cluster in Vonarskard is active throughout the period. Absolute
185 accuracy of events strongly depends on the availability of observations from close-in
186 stations, therefore events prior to the installation of the Vonarskard station in 2011
187 and during times of its outages in December 2014 through January 2015 have less
188 constrained source depths. This outage appears to affect the depth accuracy of the
189 latest events in the Vonarskard cluster – concentrated at 7-8 km depth, while the
190 previous events cluster at 3-4 km depth – and the clusters north of Tungnafellsjökull,
191 which all have depths extending down to around 8 km. The absolute depths of these
192 northernmost clusters are therefore uncertain and likely to be overestimated by 2–4
193 km.

194 Many small lineaments and fractures are illuminated by the relocated
195 seismicity. The 10 main fractures outlined by the event distribution are displayed in
196 Figure S4 (Supplementary Material). These are jointly interpreted with the

197 distribution of individual focal mechanisms to define the strike, dip and approximate
198 slip direction on each fault (Table S5 in Supplementary Material). Most of the
199 mapped faults strike in a similar direction as the long-axis of the caldera and most
200 faults within the caldera dip around 55 degrees to northwest, whereas faults at the
201 northern margin and north of the caldera are closer to vertical and have a more varied
202 strike. Slip directions are mostly normal or close to normal, except for a near-vertical
203 fault at the eastern caldera rim, active in 2012, where slip direction is close to left-
204 lateral strike slip. This fault was not reactivated in 2014-2015. The northernmost fault
205 segment (Supplementary Material Figure S4, segment nr. 5) is just southeast of a
206 surface fault mapped from aerial photographs (Björnsdóttir and Einarsson, 2013) with
207 a similar strike. The general consistency of the orientation and slip direction of the
208 fault plane solutions suggests that in spite of the rather large azimuthal gap in the
209 northwest quadrant of Tungnafellsjökull, the mechanisms are rather well constrained.

210 **2.2 Deformation modelling**

211 We generated a series of deformation models covering the period 16th August 2014 to
212 10th April 2015. Each of these models comprise i) a sill, at depth beneath Bárðarbunga
213 caldera, representing the magma reservoir; ii) a caldera ring fault system and iii) a
214 dyke, which extends from the southeast edge of the caldera to the eruption site. We
215 utilise a similar ring fault and magma reservoir geometry as employed by
216 Gudmundsson et al. (2016). In addition we also incorporate the dyke geometry
217 defined by Sigmundsson et al. (2015). We model the caldera ring fault, sill and dyke
218 as a series of rectangular dislocations (patches) in an elastic half-space (Okada, 1992).
219 We employ Bayes' Theorem to calculate the posterior probability distribution for slip
220 on the ring fault and opening of the sill and dyke, assuming a uniform *a priori*
221 probability over a range of possible values (Mosegaard and Tarantola, 1995;

222 Sigmundsson et al., 2015). We jointly solve for the opening along the dyke (24
223 patches in azimuth and 5 in depth), slip on each segment of the ring fault (24 patches
224 in azimuth and 8 in depth) and the closing of the sill (85 patches).

225 Cumulative deformation modelling is carried out over six time periods (Table
226 S2 in Supplementary Material) to determine the evolution of these multiple source
227 parameters over the course of the eruption. To maximize the amount of input data
228 used we undertook the modeling using a two-step process for the last three time
229 periods (between 4th September 2014 to 10th April 2015) where cGPS /InSAR data
230 was limited in the vicinity of the dyke from mid-September 2014 onwards. We ran
231 separate cumulative deformation models for the periods 20140904-20141023,
232 20140904-20141218 and 20140904-20150410. Then summed the opening/slip
233 derived from each of their median posterior distributions (for the ring fault, sill and
234 dyke) to those computed for the 20140816-20140904 model, as this model
235 incorporated the largest amount of input data for quality control purposes (see Table
236 S2 in Supplementary Material).

237 Although an inflating dyke and deflating sill (or spherical source) have been
238 incorporated in previous modelling studies (e.g. Sigmundsson et al., 2015, Riel et al.,
239 2015, Green et al., 2015) these sources do not fully account for the near- and far-field
240 deformation and/or agree with depth estimates of the melt, derived from petrological
241 analysis (Gudmundsson et al., 2016). In order to satisfy these independent
242 observations a large amount of fault slip is required at the caldera, combined with
243 deflation of a reservoir at a depth below the caldera floor of between 8-12 km
244 (Gudmundsson et al., 2016). We satisfy the above criteria by incorporating a
245 reactivated ring fault system at the caldera, extending from 1 to 10 km, connected to a
246 deflating sill at its base. Constraints on the geometry of the caldera fault system are

247 described by Gudmundsson et al. (2016). During each time period analysed in this
248 study (Table S2 in Supplementary Material) both dyke emplacement and caldera
249 subsidence were occurring, so each model comprises all three components (ring fault,
250 sill and dyke). The depth of the sill was here set to 10 km. This depth was determined
251 from a probability density function (PDF) generated from one million iterations of a
252 MCMC inversion of a point source (Mogi, 1958) in an elastic half-space, constrained
253 by post-rifting cGPS and InSAR observations in the far-field (Gudmundsson et al.,
254 2016). The peak of this PDF is at 10 km beneath the caldera floor with a 95%
255 confidence interval of 8-12 km. Although the magma source must have a finite
256 volume, which we model as a horizontal sill, a point source is a reasonable
257 approximation for observations far away. The location and geometry of the upper
258 edge of the caldera ring fault is determined from modelling a 1-day interferogram,
259 which demonstrated good coherence on the ice-covered volcano, and coincided with
260 three >M4, and a M5.3 earthquake which occurred on the 18th September 2014,
261 within the northern sector of the caldera. We assumed the fault on the northwest side
262 dipped at 10 degrees to the northwest (Gudmundsson et al., 2016).

263 **2.3 Stress modelling**

264 We calculate the tensile and Coulomb failure stress change for each of the six
265 time periods using our cumulative deformation models corresponding to each of these
266 intervals. The stresses are calculated using a receiver fault orientation representative
267 of the general orientation of the mapped surface and subsurface faults. Absolute stress
268 state is difficult to attain, however, we compute based on seismic observations, the
269 likely rake and planes of failure. Thus, our stress changes can be considered to either
270 add or subtract from the absolute Coulomb stress. Given the strike (θ) and dip (δ) of a
271 fault plane, the components of \mathbf{v} , the normal to that plane, are given by:

272 $v_1 = \sin \delta \cos \theta$ (1)

273 $v_2 = \sin \delta \sin \theta$ (2)

274 $v_3 = \cos \delta$ (3)

275

276 If the components of the stress tensor σ_{ij} , represent the change in stress at each point
277 on the fault, then the change in traction $\Delta \mathbf{T}$ on a fault with normal \mathbf{v} is given by:

278
$$\Delta T_i = \Delta \sigma_{ij} v_j$$
 (4)

279

280 The change in normal stress is then given by:

281
$$\Delta \sigma_n = \Delta T_i v_i$$
 (5)

282

283 The magnitude of change in shear stress $\Delta \tau$ in the direction of a vector $\boldsymbol{\rho}$ in the plane
284 of the fault is given by:

285
$$\Delta \tau = \Delta T_i \rho_i$$
 (6)

286

287 The change in Coulomb stress $\Delta \sigma_c$ is then given by:

288
$$\Delta \sigma_c = \Delta \tau + \mu \left(\Delta \sigma_n - \frac{B}{3} \Delta \sigma_{kk} \right)$$
 (7)

289

290 Where $\Delta \tau$ is the change in shear stress, μ is the coefficient of friction, $\Delta \sigma_n$ is the
291 change in normal stress, B is the Skempton coefficient and $\Delta \sigma_{kk}$ is the volumetric
292 stress change ($\Delta \sigma_{kk} = \Delta \sigma_{xx} + \Delta \sigma_{yy} + \Delta \sigma_{zz}$) (Cocco and Rice, 2002). We set $\mu = 0.6$ and
293 $B = 0.5$ (after Árnadóttir et al., 2003). We calculate $\Delta \sigma_c$ in an elastic half-space and
294 use a Poisson's ratio of 0.25 and a shear modulus of 30 GPa. We follow the
295 convention that tensile stress is positive; therefore μ is also positive.

296 In accordance with the dip on most of the mapped subsurface faults defined by
297 the relocated seismicity in the Tungnafellsjökull caldera, we assume that the receiver
298 faults have a strike of 220 degrees and dip 55 degrees (to the west). This orientation
299 agrees with the strike of most of the mapped subsurface faults, as well as the mapped
300 surface faults/fractures within the Tungnafellsjökull fissure swarm (Björnsdóttir and
301 Einarsson, 2013). This is also in general agreement with the strike of recent fault
302 movements, observed in the field in the summer of 2015, along graben faults west and
303 north of Tungnafellsjökull. To test the sensitivity of the results to variations in strike
304 and dip, we ran a series of stress models assuming receiver faults with a strike of 230
305 degrees and dip of 90 degrees. We found that this did not significantly alter the final
306 results, however predicts higher Coulomb stress changes in the northern part of the
307 swarm where faults are closer to vertical.

308 **3 Deformation and stress modeling results**

309 Based on the analysis of geodetic and seismic data acquired throughout the unrest and
310 eruption, the primary mechanism for the increased seismicity at Tungnafellsjökull
311 may be evaluated. PS-InSAR analysis, spanning the entire period of unrest and
312 eruption, displays no evidence of local deformation in the vicinity of
313 Tungnafellsjökull volcano (Figure 3 and Figure S6 in Supplementary Material), while
314 major deformation is observed within the Bárdarbunga volcanic system – related to
315 both the slow caldera collapse and dyke emplacement. The lack of deformation at
316 Tungnafellsjökull suggests that either no new intrusions/increase in pressure occurred
317 here during the observation period, or the pressure changes related to any such event
318 were too small to produce observable deformation at the surface. To determine the
319 size of possible intrusions/increases in pressure within a resident chamber that may go

320 undetected, a series of models were run using a Mogi point-source (Mogi, 1958)
321 beneath Tungnafellsjökull for both a shallow and deeper magma chamber (depths of 2
322 and 7 km respectively, Figure S7 in Supplementary Material). These models suggest
323 that any such intrusion or pressure increase within an existing reservoir within this
324 depth range could not exceed a volume change of $\sim 4 \times 10^6 \text{ m}^3$, which would produce
325 one fringe ($\sim 1.5 \text{ cm}$) of deformation in the satellite's line-of-sight (LOS). This
326 deformation is not observed on the PS-InSAR interferogram (Figure 3). We
327 nevertheless consider these volumes maximum bounds, and determine whether the
328 stress changes related to such an increase in volume could induce triggered seismicity
329 at Tungnafellsjökull. Both tensile and Coulomb stress changes were calculated for a
330 Mogi source at a depth of 7 km, since this model corresponds to a larger volume
331 change and broader deformation signal. The results indicate that an inflationary
332 source beneath Tungnafellsjökull could result in an increase in tensile/Coulomb stress
333 here, however the area affected would be quite limited and would not account for the
334 earthquake activity to the northeast or east of Tungnafellsjökull (Figure S8 in
335 Supplementary Material). For these reasons, we consider a pressure increase under
336 Tungnafellsjökull an unlikely cause for the earthquake activity. A deflation source
337 beneath Tungnafellsjökull would have the opposite effect; clamping faults in the
338 Tungnafellsjökull fissure swarm and reducing the likelihood of earthquakes (also
339 Figure S8 in Supplementary Material).

340 Stress transfer may result from either small permanent static stress changes or
341 larger transient dynamic changes (the latter of which is produced by the propagation
342 of seismic waves). However, temporal correlation between the occurrences of large
343 earthquakes at Bárðarbunga and earthquakes at Tungnafellsjökull is not observed (see
344 section S9 in Supplementary Material). Many of the largest events do not induce

345 almost instantaneous nucleation of earthquakes at Tungnafellsjökull, which would be
346 expected to be the case for dynamic triggering, since the stress perturbation lasts a
347 very short time. Furthermore, from Figure 2 we can see that the rate of seismicity
348 went well above background levels before any large events occurred within the
349 Bárðarbunga caldera and continued well above that level after the M5+ events had
350 ceased. Therefore dynamic triggering of earthquakes at Tungnafellsjökull volcano is
351 not the explanation for our observations.

352 In order to determine whether static stress changes associated with the 2014-
353 2015 Bárðarbunga unrest and eruption could be responsible for the observed
354 seismicity at Tungnafellsjökull, a robust model is required that is capable of
355 reproducing both the near- and far-field deformation observed during this extended
356 period. A series of optimal models were calculated for six time intervals throughout
357 the unrest and eruption (Figure 2 and Table S2 in Supplementary Material). Each of
358 the models incorporated a caldera ring fault system connected to a sill at its base (at a
359 depth of 10 km) and a four-segment dyke extending from Bárðarbunga caldera to the
360 eruption site. The optimal models derived for each of the intervals provides an
361 excellent fit to the input data. An example of the results of the MCMC inversion to
362 determine the optimal deformation model for the period 16th August – 4th September
363 2014 is displayed in Figure 4. A comparison between the data and the optimal models
364 is displayed in Figure 4(a-f). Figure 4(g-i) display the median posterior probability of
365 slip on the ring fault, closing of the sill and opening of the dyke respectively. By the
366 end of August 2016 the dyke had reached its full length yet subsidence at the caldera
367 amounted to ~0.5 m/day. This fast subsidence was related to magma withdrawal from
368 the reservoir below, feeding the Holuhraun eruption where the effusion rate was high,
369 200-250 m³/s in the first few weeks (Gudmundsson et al., 2016). Our model spanning

370 16th August to 4th September 2014 indicates that slip on the ring fault during this
371 period (Figure 4g) averaged 16 m. The average closing of the sill was 19 m (Figure
372 4h), with larger amounts of closing observed in the eastern part of the sill. The
373 average opening of the dyke was 1 m (Figure 4i), with greater opening towards the
374 northeastern segments in the uppermost 4 km. The asymmetric closing of the sill is
375 likely related to increased magma withdrawal from the eastern section. This pattern is
376 consistent throughout the modelled time series and may be related to connectivity
377 within the reservoir, or because of increased faulting in this region. The increased
378 opening towards the end of the dyke is consistent with that observed by Sigmundsson
379 et al. (2015) in the vicinity of the eruption site.

380 This model varies from that presented in Sigmundsson et al. (2015), in that it
381 incorporates the caldera subsidence grid (from Gudmundsson et al., 2016) but more
382 importantly a revised model geometry, comprising an enclosed caldera ring fault
383 system connected to a sill at 10 km beneath Bárðarbunga. However the opening
384 observed in the dyke is very similar to that presented by Sigmundsson et al. (2015).
385 During the entire period analysed (20140816-20150410) our model indicates an
386 average slip on the ring fault of 40 m, average closing of the sill of 50 m and average
387 opening of the dyke of 1.5 m (Figure S10 in Supplementary Material).

388 The time series of slip multiplied by area (seismic potency) for the ring fault
389 and volume change for the sill and dyke are displayed in Figure 5. The curves show
390 an initial sharp increase in seismic potency until the 23rd October 2014, followed by a
391 more gradual increase until the end of the eruption. The volume change curves for the
392 sill and dyke appear to be more correlated. In general there is a significant increase in
393 closing of the sill and opening of the dyke until the 13th September 2014, after which
394 the volume changes become more gradual. However a smaller incremental closing of

395 the sill between the 28th August-4th September 2014, appears to correlate to a larger
396 incremental dyke opening during the same period. The final volume changes
397 calculated up to the end date in the time series (10th April 2015) are $-1.9 \pm 0.1 \text{ km}^3$ for
398 the deflation of sill, $8.2 \pm 0.5 \text{ km}^3$ for the slip \times area of the ring fault and 0.7 ± 0.04
399 km^3 for the inflation of the dyke (Figure 5). The volume change associated with the
400 deflation of the sill agrees with estimates of the collapse volumes calculated by
401 Gudmundsson et al (2016) ($1.8 \pm 0.2 \text{ km}^3$). The volume intruded into the dyke for the
402 model ending on the 4th September 2014 (0.6 km^3), is larger than that presented in
403 Sigmundsson et al. (2015). This is due to the different caldera model being used,
404 leading to more opening at the corner of the second dyke segment (Figure 4i). This
405 was corrected for by replacing the patches in this column with the median value of
406 those in the adjacent column. The calculated final volume on the 4th September 2014,
407 using this corrected corner opening is 0.5 km^3 – the same as that determined by
408 Sigmundsson et al. (2015). At the end of the eruption the final corrected volume for
409 the dyke is 0.6 km^3 , suggesting that some continued widening/post-rifting inflation of
410 the dyke did occur during this period.

411 To assess the effect of the caldera collapse and dyke emplacement in terms of
412 its potential to modify the existing stress field, we calculated the stress changes
413 resulting from the modelled deformation sources within the Bárðarbunga magmatic
414 system. This was undertaken for the same six time periods (spanning the entirety of
415 the unrest and eruption), utilised in the deformation modelling (Figure 2 and Table S2
416 in Supplementary Material). The time series of Coulomb stress changes from the 16th
417 August 2014 until the 10th April 2015, for the 0-4 km depth interval, are displayed in
418 Figure 6 (this interval contained the largest number of earthquakes). The tensile stress
419 changes are displayed in Figure S11 (Supplementary Material) along with the

420 Coulomb and tensile stress changes for the 4-8 km interval (Figures S12 and S13 in
421 Supplementary Material). The stress field below this depth is not considered, due to
422 the fact that the majority of the relocated events are at depths shallower than 9 km and
423 the observation that the deeper events more commonly occur during periods of
424 absence of data from the closest seismic station (at Vonarskard), making these source
425 depths less reliable. In each of these figures tensile stress is positive.

426 The time series of Coulomb stress changes for the depth interval 0-4 km,
427 shows an initial increase in Coulomb stress trending from Bárðarbunga towards the
428 northeastern edge of both Tungnafellsjökull and Vonarskard – the magnitude of
429 which increases throughout the observation period (Figure 6). Earthquakes initially
430 occur in zones of elevated Coulomb stress <0.05 MPa within the Vonarskard caldera
431 and on the northeastern edge of Tungnafellsjökull. Maximum Coulomb stress changes
432 in the vicinity of Tungnafellsjökull are reached by 18th December 2014 (~ 0.25 MPa).
433 The tensile stress changes display a similar trend of increased stress towards the
434 northeast of Tungnafellsjökull and Vonarskard (Figure S11 in Supplementary
435 Material), with broader regions of increased stress observed after the 4th September
436 2014. These areas tend to be slightly more localised than the Coulomb stress changes,
437 but still encapsulate the majority of the earthquakes.

438 The Coulomb stress changes for the deeper interval (4-8 km, Figure S12 in
439 Supplementary Material) display a more complex evolution through time than the 0-4
440 km interval, but with the largest increases in Coulomb stress still observed at
441 Tungnafellsjökull and Vonarskard after 4th September 2014. The tensile stress
442 changes at this interval display a similar trend including an apparent reduction in
443 stress in the vicinity of both Tungnafellsjökull and Vonarskard during the 28th
444 August-4th September 2014 (Figure S13b in Supplementary Material). In general, the

445 calculated stress changes (both normal and Coulomb) for these two depth intervals are
446 in good agreement with both the spatial and temporal variation in seismicity,
447 indicating that earthquakes are occurring in regions of increased tensile and Coulomb
448 stress.

449 To determine the dominant source influence of these observed variations in
450 normal and Coulomb stress, the calculations were rerun for the deformation model
451 spanning the 16th August 2014 to the 10th April 2015, using model components which
452 contribute to the observed deformation field separately (e.g. a separate calculation
453 was undertaken for i) slip on the ring fault and closing of the sill and ii) opening of
454 the dyke. The results are displayed in Figure S14 (Supplementary Material), and
455 demonstrate that the closing of the sill/slip on the ring fault system is the primary
456 cause of the large positive stress changes observed in the vicinity of
457 Tungnafellsjökull, whereas the dyke influences the stress changes in this region by
458 reducing the tensile and Coulomb stress in the northeast sector of Tungnafellsjökull.
459 This suggests the stress changes observed in this area result from a complex
460 interaction between those induced by the deflating magma reservoir/caldera collapse
461 and the emplacement of a long dyke. This interplay is likely the result of the apparent
462 reduction in stress observed between the 28th August-4th September, as during this
463 period the modeling indicates a small incremental closure of the sill but a larger
464 opening of the dyke.

465 **4 Discussion**

466 We present a compelling example of how a major slow caldera collapse and
467 associated dyke intrusion triggered seismicity within the fissure swarm of a
468 neighbouring volcanic system, some 25 km away. Earthquakes occurred in regions of

469 positive changes in tensile and Coulomb stress over a series of depth and time
470 intervals. The region of positive changes in tensile and Coulomb stress extend over a
471 wide area from the Bardarbunga caldera to Tungnafellsjökull. However, the triggered
472 earthquake activity is mostly confined to a part of the Tungnafellsjökull fissure
473 swarm. The tensile and Coulomb stress changes in this area are between <0.05 - 0.25
474 MPa. Our interpretation is that faults in this area were previously close to failure, and
475 accordingly, unclamping of faults within the Tungnafellsjökull central volcano are
476 responsible for the observed increase in seismicity rates in this area. Although stress
477 changes related to a hypothetical undetected intrusion/inflation event directly beneath
478 Tungnafellsjökull could produce a similar magnitude stress change, the region of
479 increased tensile/Coulomb would be restricted to a small area (Figure S8 in
480 Supplementary Material), and this would not account for the spatial distribution of
481 relocated earthquakes that occurred throughout this event. Therefore we consider it
482 unlikely.

483 Understanding volcano triggering effects is important for evaluation of
484 volcanic hazards. Modelling such as that presented here, can directly help evaluate
485 hazards during an eruption and thus facilitate effective eruption response. In addition
486 to the activity at Tungnafellsjökull, it was unclear if small clusters of earthquakes,
487 occurring within the Vonarskard caldera, between Tungnafellsjökull and Bárðarbunga
488 (Figure 1c) were related to a local intrusion. The stress modelling results show that
489 these earthquakes coincide with a lobe of elevated positive stress changes (Figures 6
490 and S11), suggesting that they were the result of unclamping and movement on pre-
491 existing faults – primarily triggered by the magma withdrawal and subsidence at the
492 neighbouring Bárðarbunga volcano.

493 Our results can be compared to triggering studies at other volcanoes. Both stress
494 changes (e.g., Lind & Sacks 1998; Rowland et al., 2010; Palladino and Sottili; 2012),
495 and hydraulic connections such as a dyke intrusion intersecting a magma reservoir
496 beneath a neighbouring volcano, tapping of multiple reservoirs, and pore pressure
497 diffusion within the asthenosphere (e.g., Hildreth, 1991; Eichelberger and Izbekov,
498 2000; Ebinger et al., 2008, Gonnermann et al., 2012) have frequently been attributed
499 as the cause of coupled activity. In Iceland, the only other closely spaced volcano pair
500 studied for triggering effects are the Eyjafjallajökull and Katla volcanoes in South
501 Iceland. These reside at a similar distance from each other as Bárðarbunga and
502 Tungnafellsjökull. In that case, intrusive and eruptive activity at Eyjafjallajökull is
503 inferred to have had minor influence on the magmatic system of Katla (Albino and
504 Sigmundsson, 2014). Magma moving in the roots of the Eyjafjallajökull volcano was
505 an order of magnitude less than at Bárðarbunga, explaining the lack of triggering
506 effects.

507 A case of lateral dyke intrusion from one volcano intersecting a magma body
508 within the domain of another volcano is inferred to be the cause of the 1996 Gjalp
509 eruption in Iceland. There, geochemical and seismic data suggest that a dyke
510 propagated from Bárðarbunga and intersected a magma body in the domain of the
511 nearby Grimsvötn central volcano, located midway between the two volcanoes (Pagli
512 et al., 2007). The erupted material has an isotopic signature comparable to that of
513 eruptive products of the Grimsvötn volcano (Sigmarsson et al., 2000). There is no
514 evidence that a dyke propagated from Bárðarbunga to Tungnafellsjökull during the
515 recent eruption.

516 Gonnermann et al. (2012) modelled recent contemporaneous uplift at
517 neighbouring Kilauea and Mauna Loa volcanoes (a few tens of kilometers apart), via

518 pore pressure diffusion within an asthenospheric layer of melt accumulation.
519 Although the combined uplift could be accounted for using this model, the authors
520 also noted that the heightened intrusive/eruptive activity at Kilauea during this period
521 likely inhibited an eruption at Mauna Loa. The deformation pattern during episodes of
522 unrest at a series of volcanoes residing within the Kenyan Rift were examined by
523 Biggs et al. (2016). In the northern part of the Kenyan Rift, Paka volcano displayed
524 variable deformation between 2006-2010 related to the inflation/deflation of four
525 distinct sources. During the same time interval neighbouring Silali volcano (less than
526 25 km away) displayed a continued long-term subsidence (1-2 cm/yr), likely related
527 to cooling/crystallisation of a shallow magma chamber. In the southern part of the
528 Kenyan Rift, Logonot volcano exhibited a phase of rapid uplift between 2004-2006,
529 followed by a period of gradual subsidence (2007-2010). However, nearby volcanoes
530 Olkaria and Suswa (15 and 30 km away respectively from Logonot) showed no
531 indication of deformation during this period. The authors suggest that for small
532 increases in melt supply (causing unrest), interaction is limited to magma sources <10
533 km apart. Tungnafellsjökull and Bárðarbunga are ~25 km apart and reside within
534 separate, distinct fissure swarms. The lack of any observed GPS or InSAR
535 deformation of local origin at or near Tungnafellsjökull volcano, during this major
536 eruption within the Bárðarbunga system, indicates that the two subsurface plumbing
537 systems are not connected. There is no evidence of a decrease in pressure beneath
538 Tungnafellsjökull, which would be indicative of either pore pressure diffusion or
539 magma tapping – so no evidence for a hydraulic connection between the volcanoes.
540 However, rapid stress changes associated with major eruptions (10^9 - 10^{11} km³), may
541 trigger activity >25 km away. In this case, the stress changes can cause simultaneous
542 eruptions e.g. at Suswa and Longonot volcanoes in the Kenyan Rift (Scott and Skilling,

543 1999) and Rotorua and Ohakuri calderas in New Zealand (Bégué et al, 2014). Both of
544 these examples were major caldera forming eruptions. The synchronous caldera
545 collapse between Suswa and Logonot is thought to have resulted from lateral magma
546 movement along rift floor tension fractures combined with decompression of shallow
547 reservoirs. Bégué, et al. (2014) proposed that the 240 ka simultaneous eruption of
548 over 245 km³ of rhyolitic magma from Rotorua and Ohakuri calderas (30 km apart) is
549 linked to the existence of a continuous intermediate mush zone beneath the region,
550 comprising isolated batches of magma. Stress perturbations rather than magma
551 recharge has been suggested by the authors as the triggering mechanism for this super
552 eruption – with the evacuation of an initial batch of magma leading to the activation
553 of regionally linked faults, triggering the eruption of juxtaposed melt lenses within the
554 mush layer.

555 Our geodetic and seismic data are consistent with triggered seismicity on faults
556 in the Tungnafellsjökull fissure swarm. Faults in these areas were inspected in
557 summers of 2015 and 2016, and the primary observation is that a number of them
558 showed fresh sinkholes, with some evolution between years. Fresh sinkholes
559 identified in the summer of 2016 indicate that the movements appear to be continuing
560 following the end of the eruption. The sinkholes were all associated with faults, and
561 yet not all faults showed fresh sinkholes. This supports the conclusion that the fresh
562 sinkholes are due to fault movements on the faults that show them. The associated
563 fault displacements are inferred to be small (less than several centimetres) since they
564 were not detected on InSAR – but the overlying sediment cover of fissures may widen
565 considerably, forming localised sinks when the surface sediments trickle into them,
566 possibly facilitated by precipitation or the summer melting of the winter snow.

567 This is not the first time that triggered activity has been observed within the
568 Tungnafellsjökull's fissure swarm. During the 1996 Gjalp subglacial eruption (20 km
569 south of Bárðarbunga) and associated unrest at Bárðarbunga, Pagli et al. (2007)
570 identified three linear deformation signals, less than four km long, to the northwest,
571 north and northeast of Tungnafellsjökull volcano using a series of co-eruptive
572 interferograms. A sharp lineation to the north was the most prominent, corresponding
573 to ~14 mm of motion in the satellite's LOS. The authors determined that the
574 deformation observed to the north of Tungnafellsjökull was consistent with either a
575 slipping fault or a dyke injection, however, fault slip was considered the most likely
576 mechanism since no deformation was observed at Tungnafellsjökull. Björnsdóttir and
577 Einarsson (2013) also reported earlier fault movements (sinkholes) within the
578 Tungnafellsjökull fissure swarm, as recent as 2010, in the same areas as fault
579 movements mapped using InSAR during the 1996 Gjalp eruption (Pagli et al., 2007).
580 The abnormally high ratio of geodetic moment to seismic moment during 1996-2010
581 suggested to Björnsdóttir and Einarsson (2013) that there might have been a
582 magmatic component to these events, following the argumentation of Pedersen et al.
583 (2007). The moment ratio was however, within the normal range for the recent events.
584 The observed surface effects of triggered activity in the Tungnafellsjökull fissure
585 swarm, in relation to both the 1996 and 2014-2015 eruptions, are consistent with our
586 conclusion that faults in this area are close to failure and thus produce triggered
587 seismicity when subject to positive changes in tensile and Coulomb stress.

588 **Conclusions**

589 Between the 16th August 2014 – 27th February 2015, ~1.9 km³ of magma was
590 withdrawn from a chamber beneath the Bárðarbunga central volcano and transferred

591 through a 48 km long dyke, that in turn fed a fissure eruption at the Holuhraun plain.
592 Our modelling results demonstrate that the combination of magma withdrawal from a
593 reservoir at 10 km beneath Bárðarbunga, reactivation of the caldera ring fault system
594 and inflation of a segmented dyke can reproduce the entire deformation field (both
595 near- and far-field) observed during the 2014-2015 eruption. The total volume
596 changes calculated during the deformation modelling are $-1.9 \pm 0.1 \text{ km}^3$ for the
597 deflation of sill, $8.2 \pm 0.5 \text{ km}^3$ for the slip \times area (seismic potency) of the ring fault
598 and $0.6 \pm 0.04 \text{ km}^3$ for the inflation of the dyke (corrected volume). The computed
599 volume change in the dyke up until the 4th September 2015 is 0.5 km^3 . However, our
600 dyke volume after the end of the eruption is 0.1 km^3 greater, indicating some
601 continued widening during the eruption or post-rifting inflation of the dyke. Our
602 computed volume change associated with the deflation of the sill is comparable to the
603 total combined volume of magma intruded into the dyke ($0.6 \pm 0.04 \text{ km}^3$, from this
604 study) and the volume of erupted lava ($1.4 \pm 0.2 \text{ km}^3$, Gudmundsson et al., 2016).

605 The excellent spatial and temporal correlation between regions of
606 unclamping/increased Coulomb stress and relocated earthquakes, and the predominant
607 normal mechanisms, combined with the absence of detectable deformation in the
608 vicinity of Tungnafellsjökull, is indicative that the increased earthquake activity
609 observed at this volcano was triggered by the recent unrest and eruption within the
610 Bárðarbunga volcanic system. The cumulative seismicity plot also corroborates this,
611 as the onset of increased seismicity rate coincides with the start of the caldera collapse
612 (20th August 2014) and flattens off near the end of the eruption, when caldera
613 subsidence rates were minimal. This strongly suggests that the primary effect causing
614 the seismicity must be stress changes due to the caldera subsidence.

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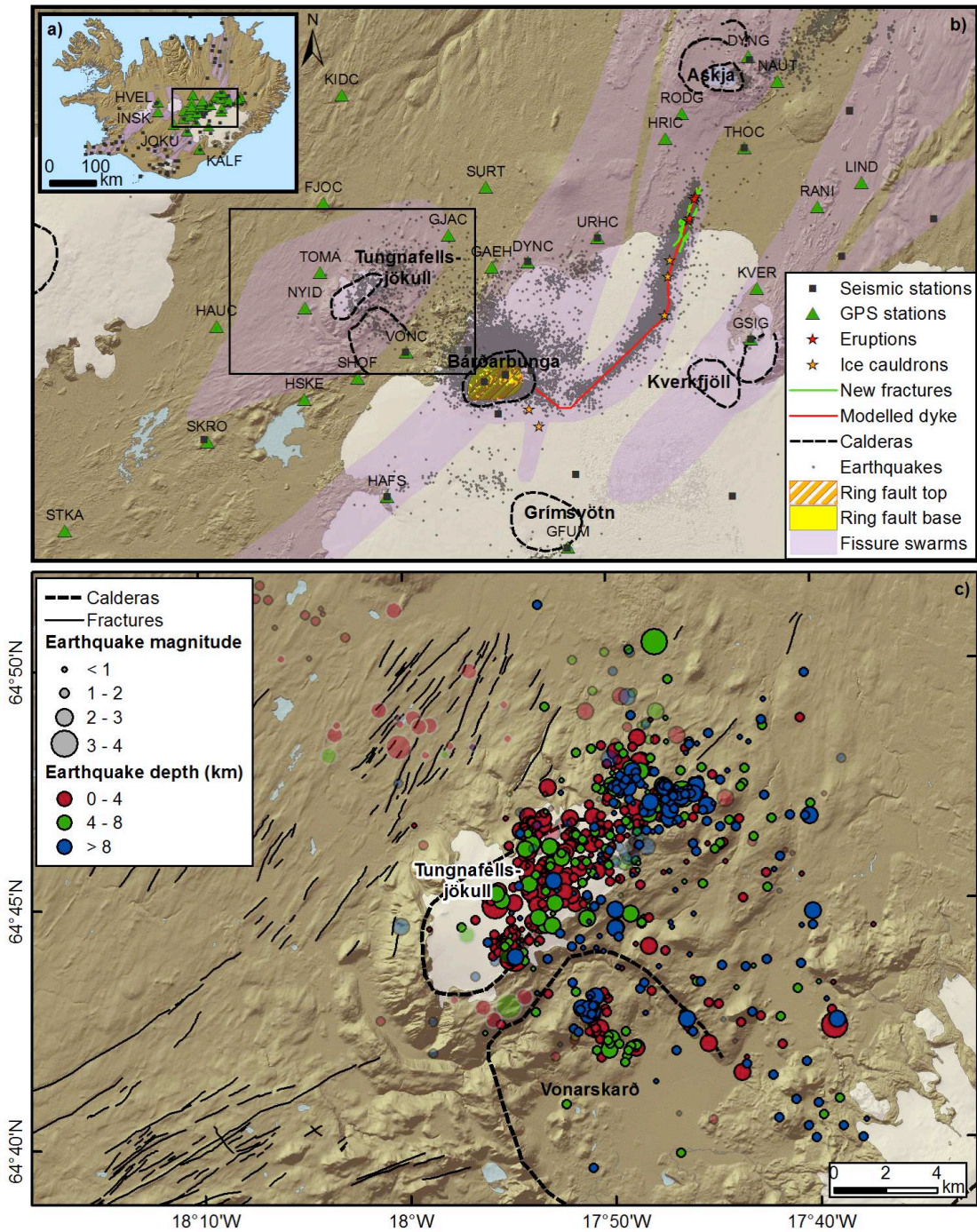
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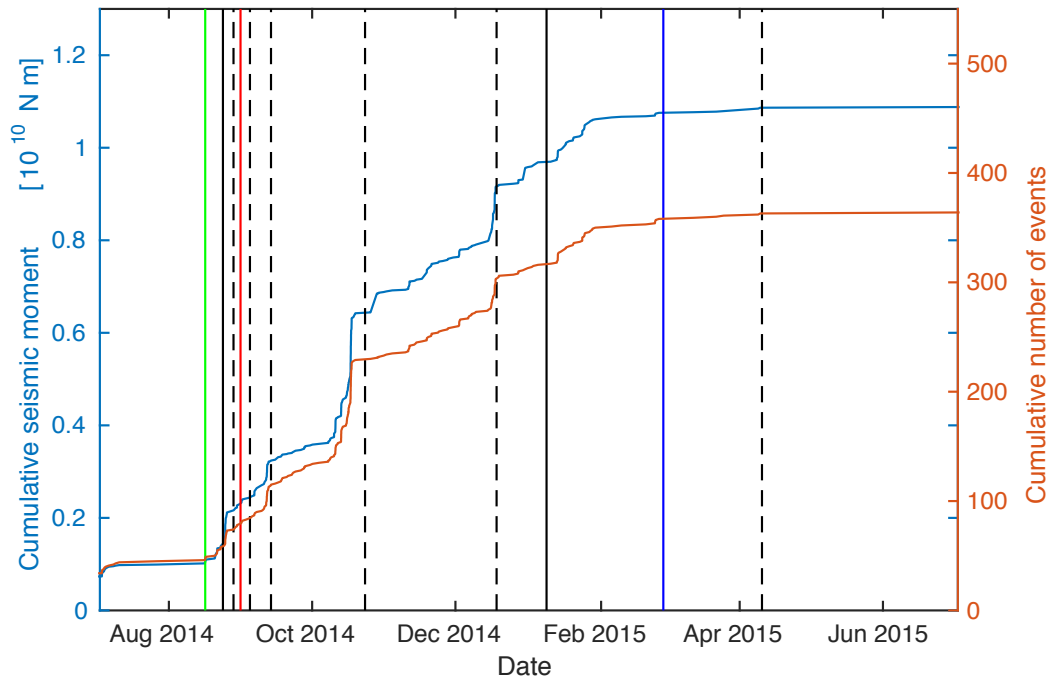
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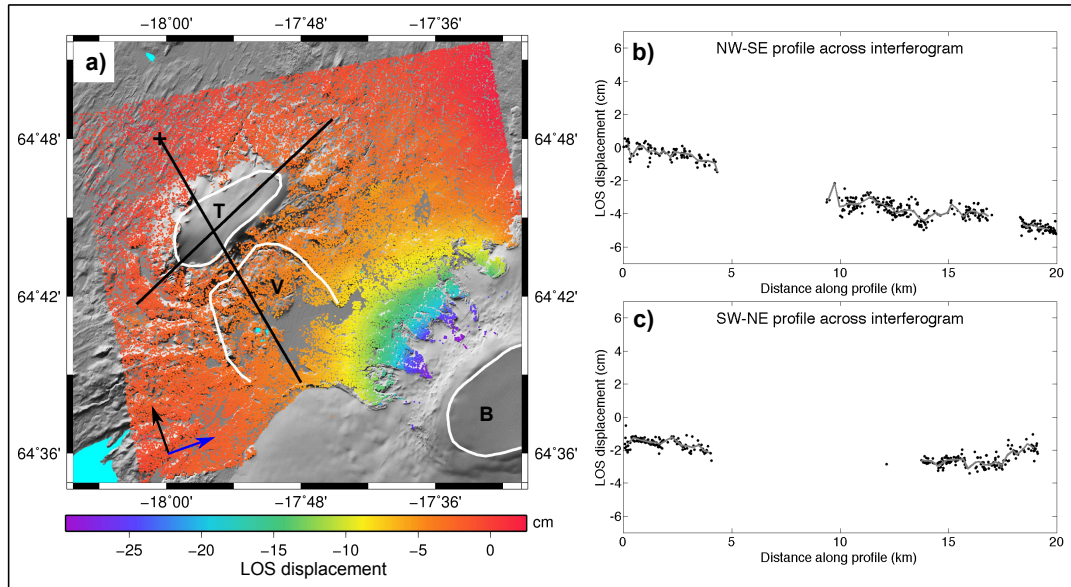


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 839 **Figure 1.** Location map. a) Map of Iceland displaying fissure swarms from Einarsson
 840 and Sæmundsson (1987), outline of study area (black box) and data utilised in this
 841 study. b) The Bárðarbunga and Tungnafellsjökull volcanoes. Earthquakes epicenters
 842 from 15 August 2014-10 April 2015 (Icelandic Met Office, 2016) are displayed along
 843 with eruptive sites and the location of the modelled ring fault, sill and dyke. c)
 844 Relocated earthquakes that occurred in the vicinity of Tungnafellsjökull during the
 845 period 5th October 1996 – 8th March 2015. Opaque earthquakes displayed with black
 846 outlines occurred between 15th August 2014 and 8th March 2015. Transparent
 847 earthquakes with black outlines occurred between 12th April 1998 – 14th August 2014.
 848 Those with light-grey outlines occurred between 5th of October 1996 and 11th April

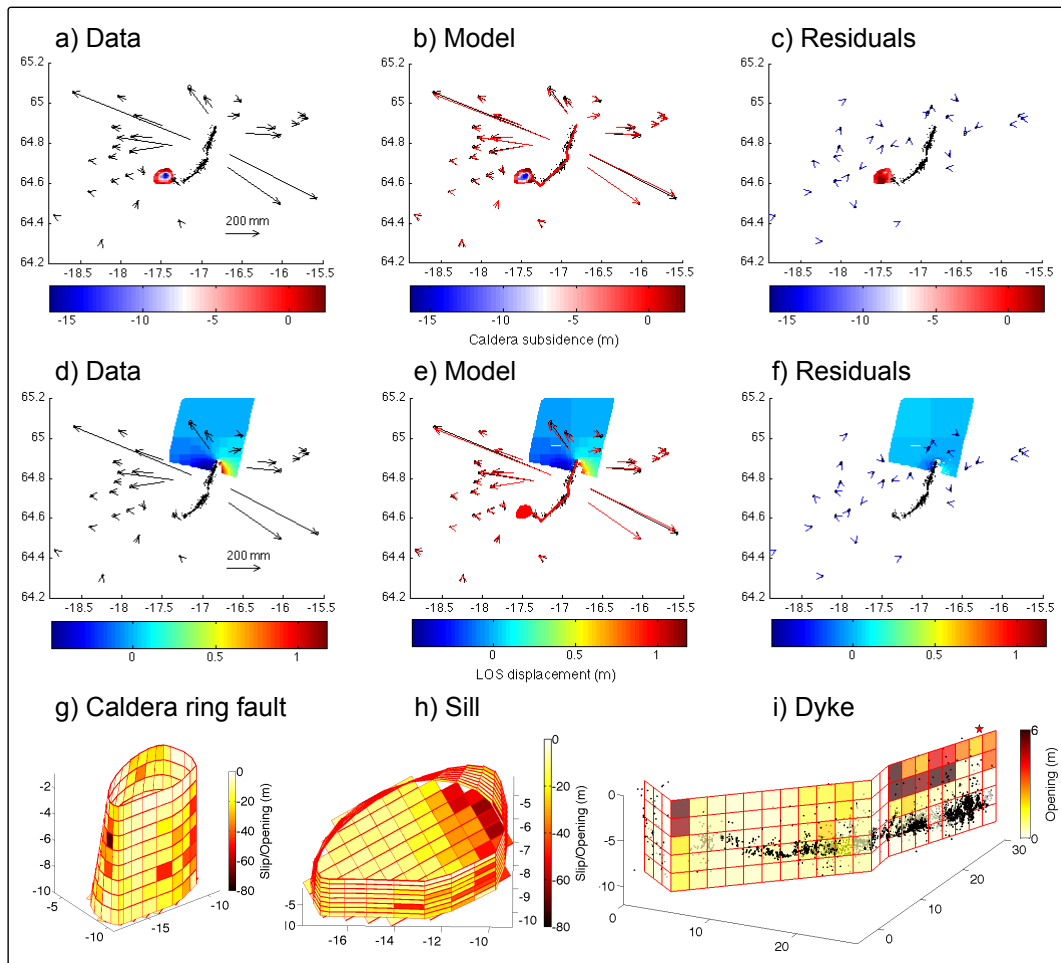
849 1998. During this period there is larger error in the hypocenter locations due to gaps
850 in the seismic network. Mapped surface fractures are from Björnsdóttir and Einarsson
851 (2013). The intermediate TanDEM-X digital elevation model provided by DLR is
852 displayed in the background of figures b and c.
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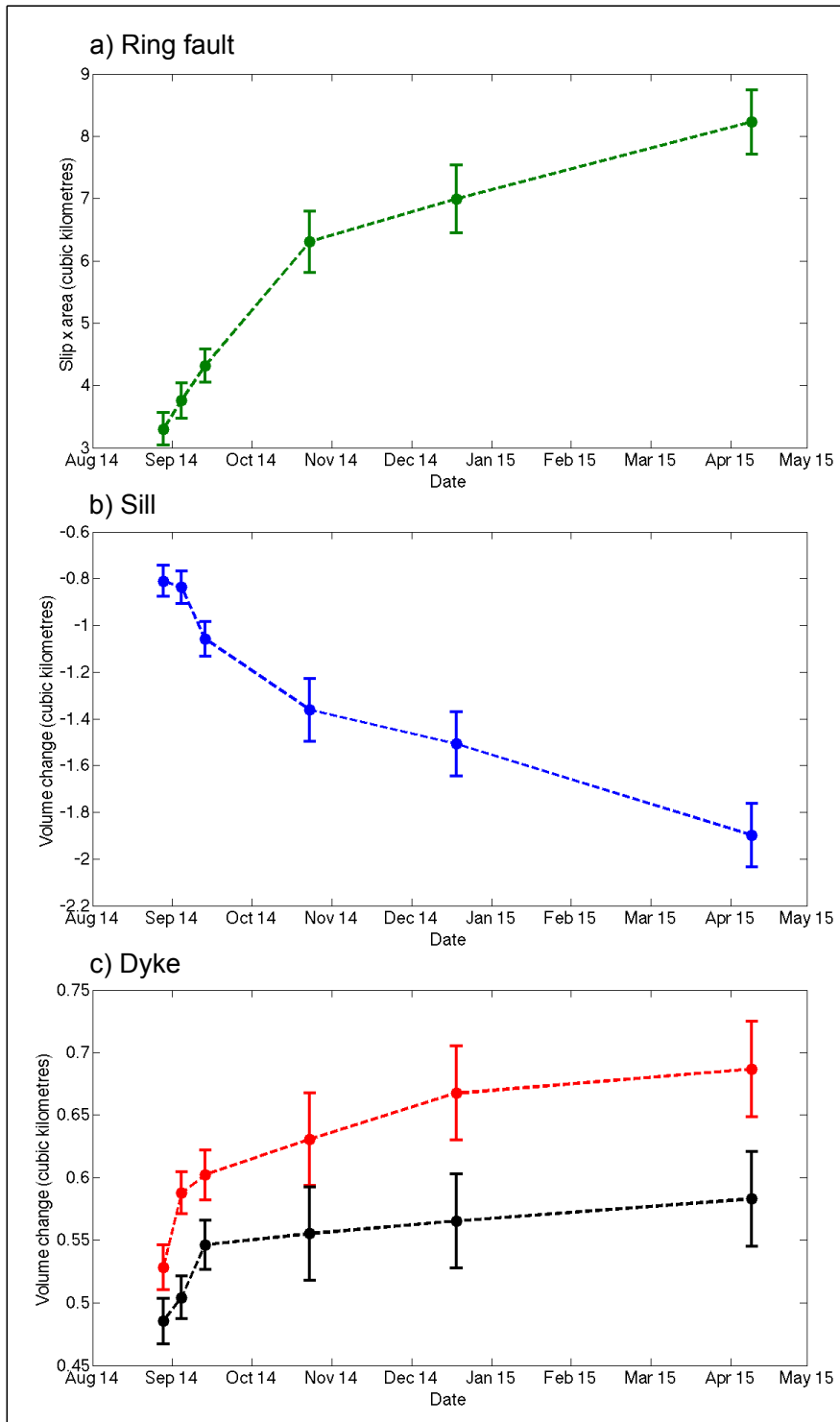
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855 **Figure 2.** Cumulative number of earthquakes/seismic moment at Tungnafellsjökull
856 since mid 2014. The vertical green line marks the start of unrest on the 16th August
857 2014, the vertical red line the start of the main eruption (31st August 2014) and the
858 vertical blue line the end of the eruption on the 27th February 2015. The solid black
859 lines indicate the first and last of the >M5 earthquakes which occurred within
860 Bárðarbunga caldera. The black dashed lines indicate the end dates for each of the six
861 time periods modelled. The figure shows that much of the activity occurred in swarms
862 of a few days duration. The region used corresponds to the black box displayed in
863 Figure 1b. Earthquakes are filtered by the magnitude of completeness (M_c) 1.5, to
864 remove any artifacts caused by changes in network sensitivity. The zero reference for
865 the y-axis is the beginning of 2010.



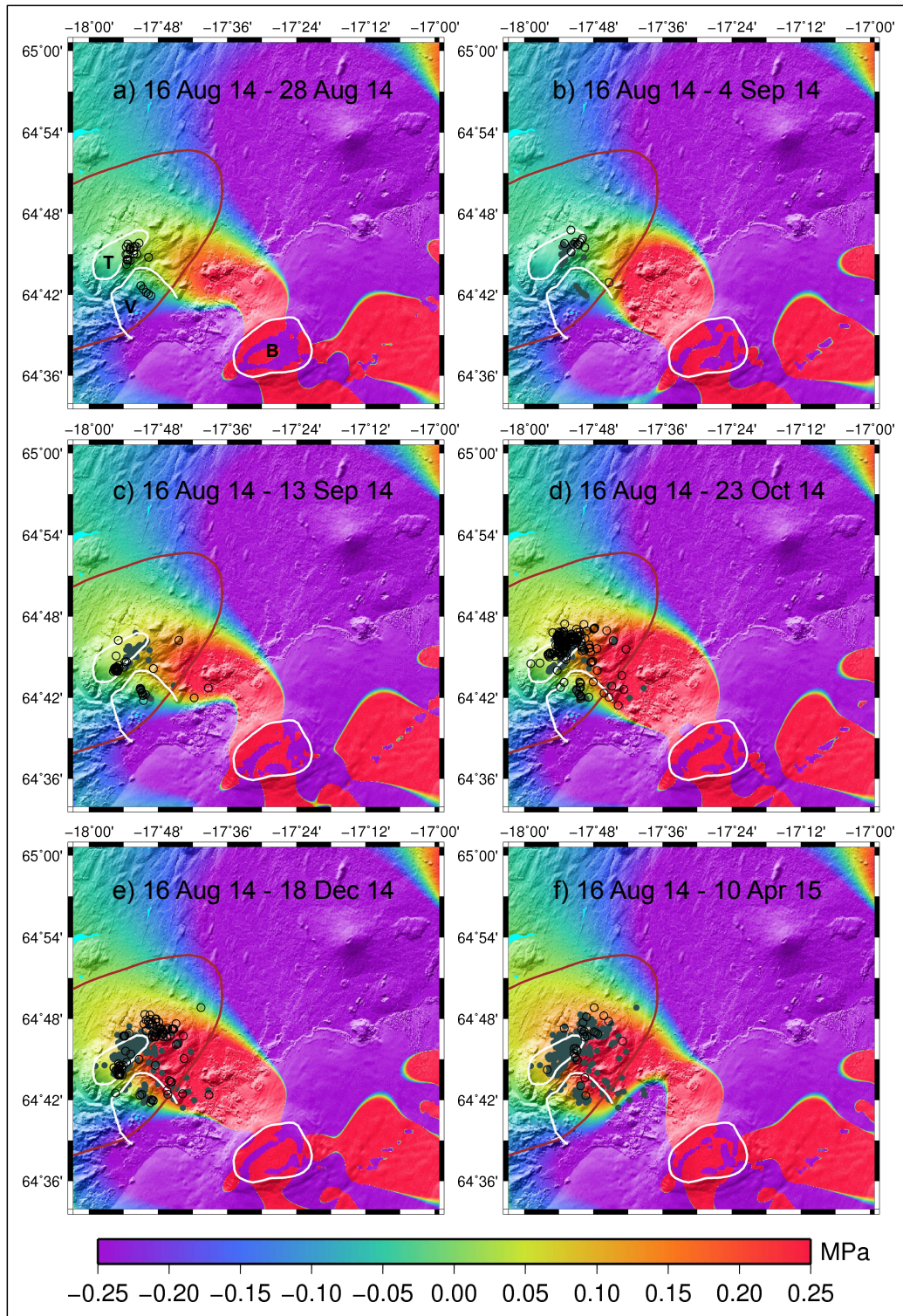
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 867 **Figure 3.** PS-InSAR analysis at Tungnafellsjökull. a) PS-InSAR cumulative line-of-
 868 sight (LOS) deformation map for the period 28th August 2014 to 29th July 2015, for
 869 ascending COSMO-SkyMed (CSK) SAR data. Underlying relief grid is the
 870 intermediate TanDEM-X DEM. Caldera outlines for Tungnafellsjökull (T),
 871 Vonarskard (V) and Bárðarbunga (B) are displayed in white. Satellite heading and
 872 look direction are displayed as black and blue arrows respectively. LOS
 873 displacements are positive towards the satellite. The reference point is displayed as a
 874 black cross. The movement away from the satellite (e.g. blue and purple amplitudes)
 875 displayed northwest of Bárðarbunga results from magma withdrawal beneath
 876 Bárðarbunga during the 2014-2015 eruption. b) Northwest-southeast profile across
 877 interferogram displayed in a). c) Southwest-northeast profile across interferogram
 878 displayed in a). In both b) & c) black dots represent individual PS-InSAR points
 879 extracted along profiles and solid grey line is the average LOS displacement.



880
881 **Figure 4.** Cumulative deformation model computed in this study, for the period 16th
882 Aug – 4th Sep 2014. Figures a-f display input data (a & d), optimal models (b & e)
883 and residuals (c & f). In figures (a-f), cGPS vectors are displayed with black arrows,
884 modeled vectors in red and residuals in blue. In (a-c) the colour scale represents the
885 cumulative caldera subsidence from the 16th August 2014 – 5th September 2014. In
886 (d-f) the colour scale represents the line-of-sight displacement (LOS) derived from a
887 descending TSX interferogram covering the period 2nd August 2014 – 3rd September
888 2014. LOS displacement is positive towards the satellite. Figures (g-i) show the
889 calculated median posterior slip on the ring fault (looking towards the northeast) (g),
890 closing of the sill (looking towards the northeast) (h) and opening of the dyke
891 (looking northwest) (i). Negative slip in (g) represents downward displacement and
892 negative opening in (h) represents closing of the sill.



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 894 **Figure 5.** Time series of a) slip \times area on the ring fault, b) closing \times area of the sill
 895 and c) opening \times area of the dyke. The red dashed line in c) represents the volume
 896 change computed from the optimal models. The blacked dashed line in c) represents
 897 the corrected volume change (see section 3 for details). The error bars represent the
 898 95% confidence interval (± 2 standard deviations of the total volume change from
 899 10,000 iterations).



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Figure 6. Cumulative Coulomb stress changes ($\Delta\sigma_c$) for depth interval 0-4 km. Red regions represent areas of increased $\Delta\sigma_c$, assuming receiver faults that are normal faults with a strike of 220 degrees, dipping 55 degrees to the west. Underlying relief grid is the intermediate TanDEM-X DEM. Caldera outlines for Tungnafellsjökull (T), Vonarskard (V) and Bárðarbunga (B) are displayed in white. Outline of Tungnafellsjökull's fissure swarm is displayed in brown. New earthquakes (since end date in previous panel) are displayed as black open circles. Previous earthquakes (since the 16th August 2014) are displayed as filled grey circles.