# **Ground deformation due to steam cap processes at Reykjanes, SW-Iceland: Effects of geothermal exploitation inferred from interferometric analysis of Sentinel-1 images 2015-2017**

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# SUMMARY

The Reykjanes geothermal system is a high-temperature seawater system situated in SW-Iceland. Interferometric analysis of the Sentinel-1 satellite synthetic aperture radar (InSAR) data has been used to determine a time series of ground deformation induced by geothermal utilization between April 2015 and October 2017. Surface displacements have been estimated at coherent pixels, indicating a steady and linear subsidence within a sub-circular bowl centered on the well field at a maximum near-vertical rate of about 25 mm/yr, together with horizontal contraction. The average line-of-sight (LOS) displacement rates from ascending and descending tracks are inverted to determine the characteristics of the deformation source at depth, modeling the geothermal reservoir as a body of simple geometry within an elastic half space. The results indicate a deformation source at about 1 km depth contracting at a rate of (0.7-0.9)×105 m3/yr during the 2015-2017 period. Using pressure and temperature monitoring data at 900 m depth as well as an analysis of the reservoir structure and rock properties, we infer that the recent estimated volume change can be attributed to steam cap processes involving a combination of compaction under pressure decrease and/or thermal contraction due to cooling of the rocks within or near the steam cap situated in the topmost part of the geothermal reservoir, in the 800–1200 m depth range. This steam cap was formed in response to a sudden pressure drop resulting from the increase in extraction of geothermal fluids for a new power plant in 2006.

**Key words**: Radar interferometry; Time-series analysis; Inverse theory; Hydrothermal systems; Creep and deformation; Mechanics, theory and modeling

# INTRODUCTION

Studies of ground deformation induced by geothermal utilization have been applied to many high-temperature geothermal systems worldwide, using geodetic methods such as GPS-geodesy or levelling. Since the 1990s, interferometric synthetic aperture radar (InSAR) satellite imaging methods have been increasingly used to characterize and monitor changes within geothermal reservoirs, for example at Krafla, Iceland (Drouin *et al.* 2017), Cerro Prieto, Mexico (Sarychikhina *et al.* 2011), Coso (Fialko & Simons 2000) and Salton Sea Geothermal Field, California (Eneva *et al.* 2014), and at the Taupo Volcanic Zone, New Zealand (Bromley *et al.* 2009). Data provided since the launch of the Sentinel-1A (April 2014) and Sentinel-1B (April 2016) satellites offers new opportunities to measure ground deformation and create improved time series as a result of more frequent image acquisitions, every six days (González *et al.* 2015; Mora *et al.* 2016; Zhou *et al.* 2017; Mellors *et al.* 2018). Several studies have used the Sentinel Terrain Observation by Progressive Scans (TOPS) data for measuring ground displacement over geothermal fields (e.g. Xu *et al.* 2017). Here we use Sentinel-1 datasets to extend the time series of deformation at the Reykjanes geothermal system, SW-Iceland (Fig. 1), for the 2015-2017 period. Deflation due to geothermal utilization (subsidence and horizontal contraction towards a well field) has been measured there since 1992 using GPS-geodesy and since 2003 using InSAR analysis of ENVISAT and TerraSAR-X SAR data (Michalczewska *et al.* 2014; Keiding *et al.* 2010; Parks *et al.* 2018).

Figure 1

The Reykjanes geothermal system is a high temperature geothermal system located where the Mid-Atlantic Ridge emerges on the southwestern tip of the Reykjanes Peninsula in Iceland (Fig. 1) (Bjornsson *et al.* 1970; Franzson *et al.* 2002). The circulating fluid has a high salinity due to the sea-water recharge of the system. The geological structure consists of a highly fractured superimposition of volcano-sedimentary strata typical of a submarine environment (Saemundsson & Einarsson 1980; Franzson 2004) intersected by sub-aerial Pleistocene lava flows (Friðleifsson *et al.* 2014). The upper part of the series is dominated by shallow water lithologies including Pleistocene phreato-magmatic hyaloclastite tuffs and breccias, intersected by thin layers of reworked shallow marine fossiliferous sediments mostly between 400 and 800 m depth (Friðleifsson & Richter 2010). Below 1100 m depth, the sequence is dominated by crystalline pillow basalts and breccias formed in a deep marine environment. Intrusions can be found with increasing density from 1.5 km depth, dominating the series below 2.8 km depth (Franzson 2004; Friðleifsson *et al.* 2014). An interval of mixed coarse-grained to fine-grained gabbroic intrusions and extrusive rocks between 1.5 and 3.2 km has been identified as a transition zone between the overlying extrusive volcanic rocks and an underlying sheeted dyke complex, interpreted as the heat source for the geothermal system (Friðleifsson *et al.* 2017). In January 2017, the deepest drillhole in Iceland, the Icelandic Deep Drilling Project well IDDP-2/RN-15 reached supercritical conditions at 4.5 km depth at Reykjanes (Friðleifsson *et al.* 2017). The temperature recorded at the bottom of the hole reached 426°C after 6 days of measurements for a pressure of 34 MPa.

An area of intense hydrothermal alteration and surface activity of about 1.5-2 km² has been delineated on the surface at Reykjanes around the Gunnuhver thermal area (Palmason *et al.* 1985), interpreted to reflect a main up-flow zone from a central part of a geothermal reservoir, where the permeability and temperatures are the highest (Fig. 1). The total volume of the estimated 1500 m thick productive reservoir was estimated to be approximately 3 km3 on the basis of these surface manifestations (Fridriksson *et al.* 2010, Axelsson *et al.* 2015). This is smaller than the total volume of 19 km3 estimated from the lateral extent of a 8-10 km2 low resistivity cap at 800-1000 m depth, representing the outer boundary of the system (Fig. 1c), identified using TEM resistivity surveys (Karlsdóttir & Vilhjálmsson 2014).

The Reykjanes geothermal system is located within a highly oblique rift on the Reykjanes Peninsula, where a natural subsidence of the plate boundary area has been observed prior to utilization at a rate of about 6 mm/yr from levelling (Eysteinsson 2000), InSAR (Vadon & Sigmundsson 1997), GPS (Hreinsdottir, 2001) and lithological studies (Friðleifsson & Richter 2010). Tectonic studies suggest that the system is confined to a NE/ENE aligned graben structure controlled by the Litla-Vatnsfell and the Skálafell faults (Fig. 1b). The highest temperatures (up to 20°C higher than anywhere else in the reservoir below 1 km depth) were measured in well RN-10, situated 600 m to the west of RN-12, in the central part of the graben (Franzson *et al*. 2002). The graben constitutes a narrow corridor channeling seawater recharge towards the well field from the southwest. Recharge is however limited by a low-permeability area around wells RN-17b and RN-30 in the south (Fig. 1c) and by an impermeable WNW-striking barrier near RN-16 (Axelsson 2012b; Khodayar *et al.* 2016). Some productive layers have been identified within porous formations at 800–1200 m depth, but most of the fracture intrusion-related feed zones are irregularly distributed. The largest are associated with sub-vertical fractures along or near dykes between 1.9 and 2.3 km depth (Franzson *et al.* 2002). Good permeability was also found in the basalt and dolerites in several locations below 3.2 km depth in the IDDP-2 drill hole, with the main feed points identified at around 3300, 4000, 4300 and 4500 m depth (Friðleifsson *et al.* 2017).

Extensive fracturing has allowed intense water-rock interaction responsible for high hydrothermal alteration of the system (Sigurdsson 2010). A high-temperature alteration profile is clearly revealed in MT resistivity models (Karlsdóttir & Vilhjálmsson 2014). A low resistivity cap (<10 Ωm), interpreted as a 8-10 km2 up-domed area (2.5 km × 3 km) elongated in the ENE direction (Fig. 1c), has been associated with a conductive smectite zone. Between 300 and 500 m depth in the central part of the system, smectite is replaced within a mixed-layer clay zone by more resistive chlorite minerals that become dominant at 500 m depth. A high resistivity core (10-30 Ωm) at deeper level reflects pore fluid conduction within the chlorite-epidote zone extending from 500 to 1200 m depth in the geothermal reservoir. It reaches the shallowest level at the Gunnuhver fumarole (100-200 m b.s.l.) and is followed by the epidote-actinolite zone down to 2.5-3 km depth (Friðleifsson & Elders 2005). Finally, the basalts and dolerites found below 3 km depth are affected by alteration ranging from the upper greenschist to the amphibolite facies (Friðleifsson *et al.* 2017). Marine sediments found within the hyaloclastite tuffaceous sequence between 390-472 and 550-800 m are intensely affected by chlorite-smectite and illite alteration, respectively, that also contain secondary crystallization of quartz, calcite and precipitated anhydrite. This intense alteration tends to have transformed these layers into cap rocks to the geothermal reservoir, extending from about 400 m down to 700 m in the centre and 900 m in the periphery of the system (Franzson *et al.* 2002; Friðleifsson *et al.* 2014). Calcite was found in abundance near aquifers at 1000-1100 m depth in well RN-10. This was taken as an indication of the occurrence of boiling conditions before production started, in the depth interval situated between the cap rock (500-800 m depth) and the beginning of the convecting zone at about 1300 m depth (Franzson *et al.* 2002).

Geothermal fluid extraction at Reykjanes started in 1970 although the first well was drilled in 1956. Large-scale utilization however only began in May 2006 with the commissioning of a 100 MWe power plant. The sudden increase in the yearly average production rate from 50 to 800 kg/s resulted in a pressure drop of about 3.0 MPa between May 2006 and May 2009 (Fig. 2), in well RN-12 located in the central part of the system (Fridriksson *et al.* 2010). Expansion of the pre-existing boiling zone in response to this pressure drawdown was associated with an increase in both surface geothermal activity and in the average discharge enthalpy of the producing wells, from 1210-1400 kJ/kg in 2006 to 1450-1950 kJ/kg in 2010 (Fridriksson *et al.* 2010; Friðleifsson *et al.* 2011; Axelsson *et al.* 2015). While the main feed zones at 1.9-2.3 km depth are still liquid dominated, those situated between 800 and 1200 m depth began to supply the wells with saturated steam, with an enthalpy of 2700 kJ/kg in 2008 (Sigurdsson 2010). That year, two relatively shallow wells, RN-27 and RN-28 (Fig. 1c), were drilled in the central part of the system down to 1225 m and 960 m depth, respectively, to produce directly from the steam cap (Fridriksson *et al.* 2010).

Figure 2

The rate of pressure drop at 1625 m b.s.l. between 2009 and 2015 was lower than in 2006-2009, about -0.1 MPa/year in the center of the system and -0.07 MPa/yr at its periphery. A cumulative drawdown of -3.8 MPa was reached at this depth in 2015 relative to 2005 and between 2015 and 2017 a minor increase in pressure of 0.3 MPa was reported (Fig. 2). Not only has pressure declined in the liquid part of the reservoir, but also in the steam cap where relatively constant pressure decline at a rate up to -0.2 MPa/yr was measured at 925 m b.s.l. between the end of 2008 and 2017, resulting in an additional pressure drawdown of about 1.7 MPa when compared to the observations at 1625 m b.s.l.

In 2016, 13.5 Mt of geothermal fluid was extracted by 15 deep production wells and the two shallow dry steam wells RN-27 and RN-28 (Þorvaldsson & Arnaldsson 2017). Some of the deep wells are cased down to the bottom of the steam zone to produce from the liquid zone only and therefore generally display higher mass extraction rates (Ómar. Sigurðsson, HS-Orka, personal communication, 2018). The increase in steam fraction in the fluid produced by deep two-phase wells however contributed to a reduction in the yearly average mass production rate from 800 kg/s in 2008 to 430 kg/s in 2016 (Khodayar *et al.* 2016). In addition, reinjection of colder separated brine at 15 kg/s was initiated in July 2009 at about 2.5 km depth into well RN-20b to counterbalance the initial pressure drop (Flovenz *et al.* 2015). Until 2017, reinjection was performed at irregular rates into five wells, averaging to 80 kg/s over the whole time period with a maximum of 146 kg/s in 2016.

Geodetic measurements of subsidence at Reykjanes have been carried out through a combination of GPS-geodesy (Hreinsdottir *et al*. 2001; Sturkell *et al.* 1994; Keiding *et al.* 2010; Magnússon 2009, 2013, 2015, 2016) and InSAR data analysis (Michalczewska *et al.* 2014, Keiding *et al*. 2010, Parks *et al.* 2018). High rates of ground deformation of about 30 mm/yr were measured during the two years following the start of the production, along a 4×3 km elliptically-shaped subsidence bowl elongated in the NE-SW direction (Keiding *et al.* 2010, Parks *et al.* 2018). Since 2008, the subsidence zone has narrowed to an area just above the central part of the reservoir, displaying a circular shape where the maximum line-of-sight (LOS) deformation rate reduced down to about 25 mm/yr. Inferred cumulative subsidence is about 26 cm in the center of the deformation field between 2005 and 2016 (Parks *et al.* 2018).

The parameters of the source responsible for the observed ground surface deformation since production started have been found by inversion of InSAR data using analytical models (Parks *et al.* 2018). These models assume that ground subsidence is the result of a pressure decrease in a body of simple geometry within a homogeneous and isotropic elastic half space representing the Earth. For the 2005-2008 period, the best fit was obtained for a near-horizontal ellipsoidal source at about 2.2 km depth displaying a rate of volume change of -7.3×105 m3/yr. Deformation modeling for the period 2009-2016 indicated a decrease in the rate of volume change down to -1.5×105 m3/yr, for a best fitting point pressure source situated at about 1 km depth. This is in accordance with the independent modeling based on GPS data from 2008–2014 (Magnússon 2015, 2016). In addition, modeling of observed gravity changes corrected for estimated elevation changes, during the 2008-2010 period were interpreted in terms of mass renewal of the reservoir fluid in the range of 30-50%, representing a recharge rate of about 250 ± 60 kg/s (Magnússon 2009, 2013, 2015; Guðnason *et al.* 2015). The data were modeled considering mass-change in a spherical volume, with best fitting depth of about 1.3-1.7 km depth (Axelsson *et al.* 2015). Smaller yearly changes in micro-gravity after 2010 show greater renewal than up to 2010, indicating an increase in the rate of mass recharge to the system since production started (Gudnason *et al.* 2018).

The objective of this study is to pursue the analysis of the ground deformation at Reykjanes through time series analysis of the new Sentinel-1 data from 2015 to 2017 and further constrain the nature of the physical processes responsible for the observed deformation at present, using an evaluation of geological conditions. Analytical models are used to fit observed surface deflation, assuming the reservoir is a contracting source at depth. The volume changes inferred from the best fitting models are then compared to the monitored reservoir pressure and temperature considering the presence of a steam cap in the upper part of the system. Linkage to the production history is made to understand the relationship between the deformation revealed by our InSAR data for the 2015-2017 period and the poro-elastic and thermo-elastic processes within the steam cap.

# INSAR DATA AND ANALYSIS

Synthetic Aperture Radar (SAR) images of Reykjanes from the Sentinel 1-A and 1-B satellites, collected in Interferometric Wide (IW) swath mode in the period 2015-2017 have been utilized to study ground motion over the geothermal reservoir. InSAR analysis determines the phase shift between SAR images acquired from approximately the same location overhead but at different times (e.g. Liu *et al.* 2017). Interferograms are formed by comparing an initial “master” SAR image with a second “slave” image. As the satellite is “side-looking”, the relative displacement value of each pixel is recorded as phase change, representing range change in the LOS direction towards the satellite (Massonnet & Feigl 1998) proportional to the radar wavelength of about 56.6 mm for Sentinel data. In the absence of errors, it corresponds to the projection of the three-dimensional (3D) displacement field onto the unit vector pointing from the ground to the satellite:

(1)

We initially analyzed the total set of 104 and 107 SAR images from Sentinel-1 ascending Track 16 (T16) and descending Track 155 (T155) respectively, covering the Reykjanes Peninsula from the 21/11/2014 to the 22/01/2018 (see Fig. 1 for coverage).

Ascending and descending interferograms with an initial resolution of about 5 m × 20 m in range and azimuth were formed using the ISCE (InSAR Scientific Computing Environment) software by processing the common bursts in the master and slave images overlapping the study area, together with the generation of coherence maps (Rosen *et al.* 2015). LOS unit vectors are and for T16 and T155, respectively, over the study area. A Tandem-X digital elevation model (DEM) was used to assist the co-registration of the slave images on the master image and correct for topographic phase. The interferometric phase was unwrapped using the SNAPHU MCF algorithm (Chen & Zebker 2001), with a mask applied to water surfaces. Before unwrapping, a Goldstein-Werner power spectral smoothing filter (Goldstein & Werner 1998) with a strength of 0.5 and a 8 × 2 multi-look operation were applied in range and azimuth to increase the signal to noise ratio. The unwrapped interferograms with resolution of 40 m × 40 m on the ground were finally geocoded using the 12-m resolution DEM grid, and cropped over the region of interest (Fig. 3).

Short term interferograms with a 12-day temporal baseline were initially formed to analyze the quality of the full set of images, skipping the unwrapping and geocoding steps to reduce processing time. As no significant deformation was expected in such a short time interval, all the SAR images resulting in spatially correlated interferometric fringes due to a high level of atmospheric phase delays (unstable atmosphere) were removed from the datasets based on visual inspection. Many winter images also led to incoherent interferograms, due to snow cover/drift, and these were also removed from the final time series analysis.

We formed stacked two-year interferograms utilizing images from both tracks, to initially visualize the total cumulative displacement between 2015 and 2017 (Receveur 2018). We then evaluated the temporal evolution of the deformation over the study period by generating a time series of deformation for both ascending and descending tracks. The selected SAR images were co-registered to a single master image in each track, acquired on 20 August 2016 in T16 and 30 August 2016 in T155, chosen to minimize the temporal decorrelation (e.g. Maghsoudi *et al.* 2018). Good orbital control of the Sentinel satellites (Appendix A) and the relatively flat topography in the Reykjanes area, with the highest point situated at less than 200 m height, ensure no significant influence of topography in the interferograms after removing topographic effects.

Having generated the time series of interferograms with respect to common master images, interferograms still containing atmospheric signals or affected by unwrapping errors were removed prior the final analysis. This procedure resulted in a total of 39 and 46 geocoded interferograms for T16 and T155, respectively, covering the period from April 2015 to October 2017 (Appendix A). For each track, average LOS velocity maps over the Reykjanes-Svartsengi area were generated (Drouin *et al.* 2017) using pixels with coherence higher than 0.8, as estimated by the ISCE software, during this time period. Displacement rates were determined pixel by pixel, using linear regression of the LOS displacement values within the whole time series of selected interferograms, relative to a reference area situated to the east of the Reykjanes geothermal system (Fig. 3).

Figure 3

The velocity maps clearly show a deformation signal at Reykjanes characterized by a sub-circular LOS increase, with highest rates in the center of the field (about -18 mm/yr in T16 and -25 mm/yr in T155 relative to the reference area). A signal of lower amplitude but larger spatial extent can also be seen at the Svartsengi geothermal field. We also measure a persistent signal at the location of the Stóra-Sandvík fault, north of the Reykjanes area (Clifton & Schlische 2003). This signal, clearly visible on the LOS displacement maps north the Stampar crater row, is characterized by subsidence at a near-vertical rate of ~4 mm/yr and a contraction toward the subsidence center at a near-east rate of ~4 mm/yr.

We created time series of LOS displacements for the average phase value of a set of pixels situated in the area of highest deformation at Reykjanes (Fig. 4). This approach was applied in order to reduce potential effects from individual pixels with significant noise. The average ascending and descending LOS velocities for the 2015-2017 period were determined for pixels having a coherence higher than 0.3 within the selected areas of each interferogram within the time series (Fig. 5). The analyses reveal a steady and linear trend over the study period, with an average LOS velocity of -16 mm/yr and -20 mm/yr in the assumed zones of maximum deformation in T16 and T155, respectively.

Figure 4

Components of the 3D deformation field in the area can be visualized by forming a linear combination of the LOS velocities measured in the ascending and descending tracks and , respectively. By adding (eq. 2) or subtracting them (eq. 3), and scaling the outcome considering the unit vectors for each track (eq. 1), an approximate estimate of the vertical (“near-vertical”) and the east component (“near-east”) of the displacement field can be derived (see e.g. Keiding *et al.* 2010). We utilized equations 2 and 3 for the decomposition, where represents the east, north and up components of the LOS vector:

(2)

(3)

Uncertainty in the decomposition approach can be approximately inferred by evaluating the relative contribution of the different displacement components. If the displacement components are of similar magnitude (horizonal and vertical displacements about equal), then equations 2 and 3 show that the component has about 0.246/1.617 (~15%) contribution from east displacement, and has about 0.043/1.149 (~4%) contribution from the up component. Overall uncertainty on and using this approach is estimated 15%. The decomposition results are displayed in the lower panels in Fig. 5. Fig. 5(c) indicates a sub-circular subsidence bowl centered on the Reykjanes geothermal field in the area of extensive geothermal alteration. The maximum near-vertical displacement rate, located in the central part of this bowl, is about -25 mm/yr relative to the InSAR reference area situated to the east of the geothermal field. The near-east displacement field (Fig. 5d) is characterized by a contraction towards the center of this zone of highest deformation, with an eastward displacement of about 5 mm/yr of the western part of the field and a westward motion of about -10 mm/yr of the easternmost part.

Figure 5

Velocity profiles across the geothermal field are shown in Fig. 6, for the T16 and T155 velocities as well as the inferred near-vertical and near-east displacement rates.

Figure 6

# GEODETIC MODELING

We invert the average LOS displacement rates obtained from the time series analyses to estimate the parameters of a contracting source at depth that best replicates the observations. A set of models representing deformation sources embedded within a uniform elastic half-space with a Poisson’s ratio are considered: a point pressure source (Mogi 1958), a finite sized spherical pressure source (McTigue 1987), a planar horizontal rectangular and square sill with uniform closing (Okada 1985), and a horizontal penny shaped crack (Fialko *et al.* 2001). These sources are assumed to correspond to a part of (the steam zone), or the total volume of the geothermal reservoir, causing the observed deformation. The inversions were performed using two different codes, both based on a Bayesian optimization approach, described in detail in Receveur (2018).

In both approaches, the LOS deformation rate is calculated relative to a reference point at (-22.564°E; 63.814°N), which corresponds to the center of the selected reference area for the InSAR data (Fig. 3). This point is situated near the STAD GPS station where a natural subsidence at a rate of about 6 mm/yr was measured before production began in 2006 (Hreinsdottir *et al.* 2001). Subsidence of this station has continued at a similar rate after the production began (Parks *et al.* 2018). Using this point as a reference, we reduce the contribution of other natural deformation signals compared to those induced by geothermal utilization.

First, inversions for a point pressure source, a finite spherical pressure source and a horizontal sill were performed using scripts from Drouin *et al.* (2017), developed to study ground deformation in North Iceland. A series of 1000 bootstrap inversions were run for each model to construct *a posteriori* distributions of the model parameters (Drouin *et al.* 2017). For each bootstrap inversion, the non-linear inverse problem is solved using a simulated annealing algorithm. The T16 and T155 average LOS velocity maps used as input were initially resampled to a larger regular grid to reduce the number of observations and keep the computational time reasonable (Drouin *et al.* 2017). A uniform LOS standard deviation of 0.5 mm/yr was assumed for each sub-sampled data point and used to weight equally the observations during the inversion. The set of model parameters that best fit the input data was determined by minimizing the chi-square (Chi²) allowing for a constant offset in the InSAR velocity fields, , where is the vector of residuals (difference between observation and prediction) and the data variance.

Results from the inversions are summarized in Table 1, including the best fitting values and the 95% confidence interval for each model parameter. All the models indicate a source at about 1-1.4 km depth contracting by an amount of about 0.9×105 m3/yr. The weighted root mean square (WRMS) values reported in Table 1 for the models inferred using the approach of Drouin *et al.* (2017) show that the best fits are obtained for the Okada square and rectangular sills (similar WRMS value).

Table 1

The Okada sill characterizes a source with horizontal dimensions much higher than the vertical extent. This is also the case of the penny shaped crack (PSC) model, that we inverted using the open-source Geodetic Bayesian Inversion software (GBIS v1.0 ©2017 Marco Bagnardi), developed at the University of Leeds. This model was used to invert for pressure change *∆P* instead of a volume change *∆V,* derived from the vertical closing at a constant rate of the Okada sills. Inversion using the Mogi point pressure source was also performed to allow a comparison of the GBIS results with the best fitting parameters estimated from the approach of Drouin *et al.* (2017). In GBIS, inversions are based on the Markov-chain Monte Carlo (MCMC) algorithm, incorporating the Metropolis-Hasting algorithm to sample the posterior probability distribution of each model parameter (Hastings 1970; Mosegaard & Tarantola 2002). This distribution, drawn from the histograms of retained solutions, provides an estimate of the probability density function of each parameter, including its most probable value and 95% confidence interval. One million iterations were run for each model to adequately sample the probability distributions of the parameters (*i.e.* Malinverno 2002), from the joint inversion of the T16 and T155 average LOS displacement rates. Prior to the inversion, the T16 and T155 velocity maps were sub-sampled using a quadtree approach integrated within the software.

Modeling results obtained from the GBIS software are in accordance with those obtained from the approach of Drouin *et al.* (2017), regarding the location of the source. The rates of volume change obtained from both methods are also consistent, except for the penny shaped crack where the volume change *∆V* depends on the shear modulus *µ,* as the model invert for the ratio *DP/µ* (see eq. 7).

We display in Fig. 7 the unwrapped data, modeled deformation and residuals for the best fitting horizontal square sill and the penny shaped crack source. We only consider these two models in the following discussion due to their good fit to the data and their simplicity (only 5 model parameters). Posterior distributions of the parameters for both models, displaying the best-fit solution as well as the 95% confidence interval, are shown in Fig. 8. Results for the other models and associated histograms are shown in Appendix B.

Figure 7

Figure 8

The locations and the dimensions of the square sill and the penny shaped crack are shown in Fig. 9 along with the estimated near-vertical velocity field, together with some of the most productive wells in 2016 (*i.e.* RN-11, RN-26, RN-12, RN-23, RN-25, RN-27). The center of both deformation sources is clearly situated in the area of maximum production, slightly to the southeast of the hottest part of the system (around RN-10) and north of Gunnuhver fumarole. The lateral dimension of the deformation source is about 1.5 km for the Okada square, and the diameter for the penny shaped crack is about 1.4 km, according to modeling results. This represents an average areal extent of about 1.9 km², which coincides with the value of 2 km² independently found by Axelsson *et al.* (2015) when estimating the area of the central part of the reservoir. The dimensions of the sources also appear to coincide with the zone of the main up-flow, where surface activity and alteration are the greatest.

Figure 9

An additional model based on using 17 point pressure sources with a volume change proportional to the injection and extraction rates at each well has also been evaluated (see Appendix B), in a similar manner as applied to the study of the deformation at Krafla by Drouin *et al.* (2017). This model, which indicates sources at 2 km depth, gives a significantly higher misfit to the data. We therefore conclude that the volume change during the period 2015-2017 cannot be directly related to the production rates and water extraction from the deep liquid dominated zone of the reservoir.

# RELATIONSHIP BETWEEN VOLUME CHANGE AND PHYSICAL PROCESSES

## Comparison with previous results and geological profile

Comparison of our modeling results for the 2015-2017 period with the results from earlier periods (Parks *et al.* 2018) indicates a continued decrease in the rate of volume change since 2006 (Fig. 10). However, the decline in volume contraction in 2015-2017 relative to the period 2009-2016 is small, and contrasts with the sharp decline observed between the 2005-2008 and 2009-2017 periods. In addition, a similar depth of about 1 km was found for all the best fitting deformation sources obtained from inversion of InSAR data between 2009 and 2017, contrasting with the depth of 2.2 km inferred from the inversion of 2005-2008 data (Parks *et al.* 2018) and suggesting a migration of the deformation processes toward shallower depth.

Figure 10

The pattern of the decline in the rate of volume change and the cumulative volume change appears to correlate well with the decrease in the rate of pressure drop measured at 1625 m b.s.l in the reservoir until end of 2015 (Fig. 2). Between 2015 and 2017, minor pressure increase is however reported at 1625 m b.s.l. in the liquid dominated part of the reservoir, while a continued pressure decline up to 1.7 MPa between 2009 and 2017 was measured at 925 m b.s.l. in the upper steam dominated part of the reservoir.

We compare the depth of the deformation sources for each time period with the geological structure of the Reykjanes geothermal system in order to evaluate what geological formations are involved. The NW-SE geological cross-section (Fig. 11) indicates that at the depth of the inferred deformation sources during the 2009-2017 period, the structure is dominated by rather fresh and porous breccias and pillow basalts only slightly compacted, with an average density ρ = 2500 kg/m3 (Hefu 2000; Friðleifsson *et al.* 2017). In addition, the center of the best fitting deformation sources is located in the 1-1.4 km depth range, which coincides with the inferred location of the steam-water boundary during the production period. The boiling zone in the initially liquid dominated reservoir would have expanded as a result of the pressure drawdown caused by the high production rates between 2006 and 2009, lowering the water table and resulting in the development of a 300-400 m thick steam zone (Appendix D). From 2008 and until summer 2016, feed zones located between 800 and 1200 m depth supplied wells with dry steam (Ómar Sigurðsson, HS-Orka, personal communication, 2017).

Figure 11

The 2.2 km deep ellipsoidal source estimated to be responsible for the deformation during the period 2005-2008 is located within a similar rock complex as the 2009-2017 deformation sources, but containing however, a greater density of dykes. The depth to its center moreover coincides with the location of the deep liquid-dominated aquifer.

We use the results from laboratory experiments performed on Reykjanes rock samples from wells RN-17b, RN-19 and RN-30 (Fig. 11) to obtain an insight into the average porosity of the reservoir and evaluate the sensitivity of the geological formations involved in the deformation to changes in effective stress and temperature (see detail in Appendix C). Using a reservoir areal extent of 1.9 km², we infer a total pore space of about 0.6 km3 based on the cumulative thickness and the average porosity of each rock type situated within the productive part of the reservoir, between 800 and 2800 m depth. This corresponds to an average micro-scale porosity of 15% for the 3.8 km3 reservoir and constitutes a lower bound on the real macro-scale reservoir porosity. The greatest volume of pore space, about 0.4 km3, is contained within the pillow basalt and dolerite dyke sequence where fractures provide additional large-scale porosity. These rock types have been interpreted to be the most sensitive to change in pressure and temperature relative to hyaloclastites and are therefore likely to compact under increasing effective stress, especially when altered with chlorite (Reinsch *et al.* 2016). A total closure of the estimated micro-scale pore space within this complex would therefore induce a volume change of -0.4 km3 concentrated below 1 km, where the basaltic lava and dyke sequence dominates the volcano-sedimentary succession. This value, which thus represents a lower bound on the maximum volume change due to pore closing, is about two orders of magnitude higher than the cumulative volume change ∆*Vtot* of -3.9×106 m3, estimated at Reykjanes since 2005 (Fig. 10b). We therefore infer that only minor (on the order of 1%) closure of pore space and rock compaction has occurred in the reservoir between 2006 and 2017.

## Deformation and physical processes

Ground deformation above geothermal reservoirs is often attributed to poro-elastic or thermo-elastic processes, associated with pressure change or cooling of the reservoir rock. Change in specific volume, , relates to change in pressure, *P*, and temperature, *T*, through the following equation:

(4)

where is the volumetric coefficient of thermal expansion and *c* the total system compressibility of the fluid saturated material. The isothermal compressibility describes the change in a volume/thickness of rock due to a change in pressure under isothermal conditions. It is defined by a complex equation including the compressibility of the solid rock matrix and the compressibility of the pore fluid, which depends on the relative proportion of liquid water and steam phases (Grant & Sorey 1979; Brock 1986). Depending on the nature of the system (*i.e.* confined, unconfined, liquid-dominated, steam-dominated, two-phase system), the compressibility also controls the mass of fluid that can be stored or released per unit volume of rock under a unit pressure change, referred to as storativity (Axelsson 2012a).

In a liquid-dominated system, the reduction of pore pressure as a result of the depletion of fluid storage may lead to the compaction of water bearing deposits under increasing effective stress. Volumetric contraction can also be induced by cooling of the rock matrix under natural recharge or reinjection of cooler fluid as well as water vaporization, process requiring the transfer of heat energy from the rock to the fluid (*e.g.* Im *et al.* 2017, Ali *et al.* 2016). Thermal and poro-mechanical processes are however generally coupled in two-phase systems and can occur on various time scales that can be complex to separate. At Reykjanes, the migration of the modeled source of deformation from 2.2 to 1 km depth together with the change in both the subsidence pattern and the deformation rate since the end of 2008 clearly suggests an evolution of the responsible deformation mechanisms. We first consider in detail the effects of pressure changes, exploring as well possible change in material properties over time (i.e. compressibility), and then evaluate the effects of temperature changes. We infer the deformation at Reykjanes within our study period results from a combined action of pressure and temperature change within a steam zone.

## Relationship between volume and pressure changes

Volume change within a geothermal reservoir can be attributed to the compaction of unconsolidated layers in response to utilization. Compaction processes are related to pressure change through the one-dimensional Terzaghi poro-elastic consolidation theory (Terzaghi 1925). According to this theory, the reduction in thickness of a layer results from a closure of the pore space in response to slow drainage of the pore water from the stressed deposits, followed by a reduction of the grain size under inter-granular transfer of these stresses. In absence of temperature change, eq. (4) can be transformed into:

(5)

where *h* is the amplitude of the compaction and is the thickness of an unconsolidated layer. Compaction therefore mainly depends on the total compressibility of the combined pore fluid and rock matrix *c* (Geertsma 1957), on the amount of pressure drop but also on the initial porosity and permeability of the reservoir rock and its pre-consolidation stress (Bull 1964; Grant *et al.* 1982; Zhang *et al.* 2012; Mondoni *et al.* 2013). With compaction, the porosity of the volume of rock involved is expected to decrease, causing a progressive decline in the total system compressibility and thus in the rate of deformation (Zhang *et al.* 2009). Grant *et al.* (1982) moreover showed that for the same reservoir volume, the total compressibility of a confined two-phase system (10-6 Pa-1) is generally one order of magnitude higher than the compressibility of dry steam (10-7 Pa-1) and three orders of magnitude higher than liquid compressibility (10-9 Pa-1). Change in compressibility is therefore likely to occur with a modification of the pore fluid resulting from boiling or re-saturation processes within the reservoir, caused by pressure variations during production. This will impact the storage and deformation mechanisms of the rock that, in return, control the reservoir pressure and temperature (Bromley *et al.* 2015).

We explore the relationship between the estimated volume change for each time period and the measured pressure change at 1625 m b.s.l () and 925 m b.s.l () to evaluate possible change in the system compressibility. This would mostly be attributed to an increase in the pore fluid compressibility resulting from the replacement of liquid water by steam in the steam zone, developed between 2006 and 2009. In absence of temperature change, eq. (4) can be expressed as:

(6)

We assume that is the pressure change in the liquid dominated part of the system of initial volume *Vr* during the period 2005-2008, while is representative after 2009 of the pressure change within the steam zone of volume *Vcap* formed in the upper part of the reservoir.

We use *Vcap* = 0.6-0.8 km3 (see details in Appendix D) considering the thickness of the steam zone (300-400 m) and the estimated average areal extent of the modeled deformation source of 1.9 km². The ratio / and / estimated for each time period are summarized in Table 2.

Table 2

Considering a uniform pressure drop within the whole reservoir of total volume *V=Vr* between 2005 and 2008, we find a value for the total compressibility of about 2×10-10 Pa-1 using /This value is assumed to represent the compressibility of the liquid dominated system and is in accordance with the total compressibility estimated from injection in the deep well RN-30 (Kajugus 2015). We then use the / ratio from 2016-2017 together with the volume range *V*=*Vcap*, to estimate the compressibility of the steam saturated material that is assumed to contribute to deformation in the upper part of the system within this time period. We find a compressibility in the order of (8-10)×10-10 Pa-1, with the higher estimate obtained for *Vcap =*0.6 km3, which is in accordance with the total system compressibility estimated from injection tests in the shallow well RN-32 (Kajugus 2015). It is also one order of magnitude higher than the compressibility of the liquid dominated zone, as inferred in general by Grant *et al.* (1982).

The inferred values of compressibility indicate that the geothermal reservoir should not be considered as a uniform volume and that heterogeneous compaction might occur at different depth. We can use these values to estimate the amount of compaction that may have occurred as a result of the cumulative pressure drawdown within the liquid dominated part of the reservoir and within the steam zone, throughout the different time periods. Between 2005 and 2015, a drawdown of 3.8 MPa has been measured at 1600 m depth in the central part of the system, in well RN-12, and of 2.5 MPa at the periphery, in well RN-16 (Ómar Sigurdsson, HS-Orka, personal communication, 2018). Using an average pressure drawdown of 3.2 MPa over the whole 1.9 km² area (main up-flow zone in Fig. 1) , an average compaction of -1.3 m is estimated for a compressibility equal to 2×10-10 Pa-1 and a reservoir thickness of 2 km (Fig. 12a). This would produce a volume change ∆*Vr* = 2.6×106 m3 which is somewhat lower than the cumulative volume change of about -3.5×106 m3, inferred between 2005 and 2015 from InSAR (Fig. 10b).

We thus examine the amount of compaction that would result from the continued pressure decline at the depth of the modeled deformation source from 2009 and until 2017. As mentioned earlier, this compaction may result from the additional 1.7 MPa pressure drop monitored during that time period in wells RN-27 and RN-28 at 925 m b.s.l. (Fig. 2), where water has been replaced by steam, resulting in higher system compressibility of about 10×10-10 Pa-1. Assuming that the steam cap has a thickness of 400 m, we find that a vertical compaction of -0.7 m might occur at shallow depth (Fig. 12a), corresponding to a volume change in the steam zone of *∆Vcap*= -1.3×106 m3. The compaction within the liquid dominated part of the reservoir together with the compaction within steam zone does together produce the observed total volume change ∆*Vtot* over the whole production period 2005-2017 (Fig. 10b). Comparing the total amount of compaction inferred in this manner and the estimated maximum volume of pore space in the reservoir, we can estimate the overall change in porosity of the reservoir to be about 0.7%.

We can alternatively use the results from the penny shaped crack model for the period 2015-2017 to determine the relation between pressure and the volume change that would occur in an oblate spheroid, considering it to represent the steam cap. In the penny shaped crack model, assuming a Poisson’s ratio equal to 0.25, the pressure change is related to a volume change by (e.g. Rivalta & Segall 2008):

(7)

The radius estimated in the inversion is well constrained to and the estimated ratio of yearly pressure change over shear modulus is /µ = -9.96×10-5, which corresponds to a volume change of about -0.7×105 m3/yr using eq. (7). The value of the effective shear modulus *µ* in geothermal areas is uncertain. We therefore compare the pressure changes inferred for a range of *µ* to the average yearly pressure change = -0.15 MPa/yr measured in 2015-2017. If we consider the volume change obtained from the penny shaped crack model, the pressure change observed in 2015-2017 is obtained for a shear modulus of 1.5 GPa. This is more than one order of magnitude lower than the value of shear modulus used for regional geodynamical models (often 30 GPa). Using a shear modulus of 10 GPa (defined at Reykjanes by Keiding *et al.* 2010), a pressure change of about 1 MPa/yr would be necessary to produce the inferred volume change and thus, poro-elastic compaction would only contribute to a maximum of 15% of the observed deformation.

Given the uncertainty in the actual value of the shear modulus, we also explore the possibility of cooling within or above the steam cap to explain the volume change during the 2015-2017 period in the following section.

Figure 12

## Cooling of a horizontal layer

In addition to the recent shallow depressurization, cooling trends have been suggested close to and within the steam cap, based on the monitored temperature in the same shallow wells RN-27 and RN-28 (Guðmundsdóttir 2016). The 2016 conceptual model of the Reykjanes geothermal area by ISOR - Iceland GeoSurvey suggests a decrease in temperature from 270°C in 2008 to 240°C in 2015 at 925 m b.s.l, representing a cooling rate of about 4-5°C/yr, within a volume of rock situated between about 600 and 1200 m depth (Khodyar *et al.* 2016). However, no signs of cooling have been detected in the convective liquid dominated part of the reservoir, below 1500 m (Fig. 12). There a slight increase in the average temperature from 270-280°C to 280-290°C has been inferred in association with the increase in enthalpy between 2006 and 2010 (Fridriksson *et al.* 2010). In the absence of pressure change, eq. (4) can be written as:

(8)

if a scaling coefficient, , is introduced (see below). This equation relates the amount of contraction in the vertical dimension of a rock body of initial height due to a change in temperature . Considering that rock contraction under cooling is a volumetric process, Werner *et al*. (2017) used to relate the linear coefficient of thermal expansion to change in elevation, with being referred to as the effective vertical thermal expansivity. If the volumetric contraction is only accommodated by contraction in the vertical direction due to cooling from above or below and no change in the horizontal dimension is induced, then = 3. Estimates for the volumetric coefficient of thermal expansion of rocks, , is generally in the range of 1-5×10-5 °C-1 (Fialko & Simons 2000; Ali *et al*. 2016; Im *et al*. 2017), with values of 2×10-5 °C-1 in basalt-like composition rocks (Robertson 1988) and as low as 3×10-6 °C-1 suggested by Peck (1978) for Alae Lava Lake in Hawaii.

Using eq. (8), we estimate the temperature change that would be necessary to produce the same volume change as modeled from InSAR, if cooling occurs near or within the steam zone. One can use the inversion results for a horizontal Okada sill of 1.5 km side contracting by a constant amount of -4 cm/yr, corresponding to a rate of volume change of -0.9×105 m3/yr (Table 1). This model is chosen for its direct estimation of vertical closing, considered to represent at best the vertical contraction of a horizontal layer due to cooling. We first assume that the sill represents the 400 m thick layer of rock within the steam zone. Using a range of effective vertical thermal expansivity values (0.1-2) × 10-5 °C-1, we find that the temperature change required to reproduce the inferred volume change during the period 2015-2017 ranges between -100°C/yr and -5°C/yr. For a temperature change of -4°C/yr, a contracting layer of about 500 m could produce the observation for 2×10-5 °C-1 (Fig. 12b).

These results indicate that the thermal contraction of rocks resulting from a 4-5 °C/yr cooling within a 400-500 m thick layer can produce the same volume change as a pressure decrease, only with an effective vertical thermal expansivity of about 2×10-5 °C-1. This thickness is consistent with the vertical extent of the cooling rocks modeled between 600 and 1200 m 100 m depth in the upper part temperature profiles of the 2016 conceptual model (Khodayar *et al*. 2016).

The analysis above shows that both poro-elastic and thermo-elastic processes, in or near the steam cap zone in the Reykjanes geothermal reservoir, are likely to contribute to the observed deformation during the period 2015-2017. Temperature and pressure changes are indeed closely related under boiling conditions in two-phase systems. Furthermore, explaining the deformation only in terms of pressure change or temperature change requires values for effective shear modulus and effective vertical thermal expansivity at an extreme end of the likely value for these parameters. Therefore, deformation mechanisms may be expected to result from the combined action of temperature and pressure decline, considering more realistic values for the shear modulus and thermal expansivity.

# DISCUSSION

We have explored two main physical deformation processes that may explain the volume change of the Reykjanes geothermal reservoir during the 2015-2017 period, based on the measured surface deflation with InSAR methods and its comparison with previous deformation studies for earlier time periods (2005-2016). The possibility of compaction of reservoir rocks under pressure decrease within a steam cap of higher compressibility (Grant & Sorey 1979) was considered as well as thermal contraction of the rocks in the upper part of the reservoir due to cooling within or near the steam cap.

The hypothesis that the observed subsidence between April 2015 and October 2017 does originate from a shallow steam zone was suggested from the comparison of the inferred depth of the best fitting deformation source (1-1.4 km) with geological, temperature and pressure profiles, together with an analysis of the production history of the reservoir. The lateral extent of the modeled depressurized sources indicates that this steam cap covers a surface area of about 2 km², in accordance with the extent of the geothermal manifestation on surface (Palmason *et al*. 1985). Some residuals elongated in the NE-SW direction in the models however suggest that all the 2015-2017 deformation cannot be explained by the inferred shallow source and that NE-SW faults of the fissure swarm have some influence on the deformation field at Reykjanes. These residuals might be partly explained by the rectangular Okada source striking N46°E that offers a good fit to the data (Appendix B). Similar strike and elongation of the subsidence bowl was observed with InSAR at Reykjanes for the period 2005-2008 and best modeled by an ellipsoidal source at 2.2 km depth. We suggested that this initial deformation was due to the immediate compaction of basaltic rocks and dolerites in response to a large 3.0 MPa pressure drop in the central part of the reservoir in 2006-2009. Enhanced pressure diffusion along deep permeable NE-SW striking faults would have controlled compaction in the deepest part of the reservoir, along this direction.

The velocity residuals, after accounting for our model, show also evidence of an additional signal in the area between the Stampar crater row and the Stóra-Sandvík fault (Figures 1, 3 and 7). Landscape in this area suggests structures of a rift valley and important normal fault (the Stóra-Sandvík Fault) bounding the area to the northwest, suggestion eventual long-term focus of deformation in the area. Deformation may also be reflecting adjustment to prior events in the area, including dike emplacement in association to the formation of the Stampar crater row in 1210-1240 AD, or seismic activity in recent years. The area is at the northern boundary of long-term seismicity on Reykjanes (Keiding *et al.* 2009; Björnsson *et al.* 2018). Another possibility is that it relates to the geothermal activity or utilization. The area is north of well RN-34 were fluids have been injected and additional cooling may have taken place. The signal could also relate to general drop in pressure or pressure in the steam cap, if tectonic lineaments would link this area to the main geothermal reservoir. Further research is needed to evaluate each of these different possibilities.

The continued subsidence at lower rate since 2009 can partly be explained by compaction (i.e. reduction of the pore volume) within the inferred shallow steam zone, due to the continuation of pressure decrease at that depth. We also suggest the existence of a slow or delayed diffusion of the effect of this pressure decline, influenced by layers of lower permeability close to or within the inferred steam and boiling zones, where the pore fluid compressibility is higher. Such non-linear relationship between pressure change and volume change of the deformation sources has been explored for example at the Ohaaki and Wairakei–Tauhara geothermal fields in New Zealand, where subsidence is inferred to be due to a reduction in pore pressure in a shallow steam zone as a result of geothermal production (Allis 2000; Koros *et al*. 2016). The delayed subsurface compaction at gradually falling rates was explained by a slow drainage of shallow boiling sandstone layers containing local low-permeability and highly porous (about 50% porosity) interbeds (White *et al*. 2005; Bromley *et al*. 2009; Hole *et al*. 2007; Brockbank *et al*. 2011). Creep deformation was there attributed to the high compressibility of the fined-grained clay mineralogy such as smectite, present at 5-30% in the altered Huka Falls Foundation mudstone aquitards. It was estimated to have about two orders of magnitude higher compressibility than typical for 5-10% porous basement rock (Bromley *et al*. 2009), increasing their possibility to compact under increasing normal stresses (Bromley & Reeves 2013). At Reykjanes, such creep deformation within the sequences weakened by smectite/chlorite alteration (i.e. above 1200 m depth) might however be buffered by the presence of denser layers consolidated by secondary mineralization (*i.e.* by quartz, anhydrite, calcite).

Thermal contraction can then be considered as a deformation mechanism taking over poro-elastic compaction (Mossop & Segall 1997), as observed at many systems when pressure equilibrium is reached after a few years of geothermal production (Im *et al*. 2017). Cooling in geothermal system can result from heat exchange between the host rock and colder recharge water. Such thermal effects have been suggested to be the cause of the subsidence observed in the Krafla geothermal field in Iceland (Drouin *et al*. 2017). At Reykjanes, no significant temperature change has been measured in the convective zone (below 1500 m depth) since production started. In addition, a slight but continuous decrease in the discharge enthalpy between 2010 and 2017 (Khodayar *et al.* 2016) suggested that the steam cap was no longer expanding in 2015-2017. Vaporization, which is generally responsible for the cooling of host rocks as it requires transfer of heat energy to the vaporized fluid, is therefore not expected during the study period. We thus consider combined temperature and pressure related processes linked to the two-phase nature of the geothermal system to explain the cooling of the rock observed in the steam and boiling zones (between 700-800 and 1200-1300 m depth).

Pressure within a steam zone may indeed rise or fall depending on the relative displacement of the water-steam boundary. When cold water is reinjected in a steam zone, steam condenses to provide the system with the energy needed to heat up the inflowing water toward reservoir temperature. The volume of steam condensed is generally higher than the volume gained as liquid water, resulting in pressure decrease and a loss of volume that can induce surface subsidence (Grant *et al.* 1982). Reinjection in the liquid dominated part of a reservoir can also, in spite of providing deep pressure support, reduce boiling and thus the generation of steam, providing less mass and heat recharge to a steam zone (Bromley *et al*. 2009). This latter mechanism has been inferred to be responsible for continued subsidence at Spa Bowl, New Zealand (Bromley *et al.* 2015). It could also explain the gradual pressure decline together with the cooling measured in the steam zone at Reykjanes since reinjection of cold water started at 2 km depth in 2009.

Porosity change caused by secondary mineral precipitation may influence processes in the area. Sulfide assemblages were deposited in the reservoir as a result of extensive boiling during the first three years of production (Hardardottir *et al.* 2010). Cooling around reinjection well RN-20b was also responsible for amorphous silica precipitation in 2009 (Berehannu 2014). However, geochemical analysis indicate equilibrium between the produced and the re-injected geothermal fluids since 2010 (Óskarsson *et al.* 2015), suggesting limited contribution of the geochemical alteration of the reservoir rock on the deformation within the two-year time scale of our study. Further analysis would be necessary to characterize the impact of porosity change caused by secondary mineral precipitation on rock compressibility and pressure diffusion in the reservoir since production started in 2006 and identify how such consolidation would affect the deformation patterns.

Models of deformation processes have been realized based on the monitored pressure and temperature at 925 and 1625 m b.s.l. and compared to the inferred rates of volume change. We suggest that a combined effect of temperature and pressure decrease in the upper part of the reservoir is responsible for the observed volume change during the period 2015-2017, considering realistic values for the shear modulus and coefficient of thermal expansion of the rock. The increase in the extraction of steam directly from the steam zone between 2009 and 2016 and the increase in reinjection rate until February 2017 would have resulted in steam condensation, together with a reduction in steam recharge, suggesting the decline of a steam zone during the period 2015-2017.

The free access of the Sentinel-1 data with a 6-day acquisition interval has allowed us to create a time series of ground surface deformation with a dense temporal and spatial resolution from April 2015 to October 2017. The linear ascending and descending LOS increase of about 16-20 mm/yr have been associated with an average volume decrease in a shallow steam zone of about 0.9×105 m3/yr, due to both thermal contraction and compaction of the reservoir rock. Despite their simplicity, the use of analytical model to simulate InSAR observations appears to be a valuable tool to monitor change and improve the understanding of sub-surface processes occurring in utilized geothermal systems. Numerical modeling software such as the non-isothermal multiphase multicomponent fluid flow simulator TOUHGH2 (Pruess *et al.* 1999) or the parallel finite element code for modeling crustal deformation Defmod (Ali, 2014) could also be used to support further investigation of the mass and heat transfers within the two-phase system and their impact on ground subsidence, taking into account the heterogeneities of the reservoir (i.e. permeability, compressibility, thermal properties). Better constraints of the shear modulus and coefficient of thermal expansion of the host rock would be helpful to perform still improved simulations of deformation. This could be achieved through additional laboratory experiments on core samples. Given the known production history of the reservoir, long-term monitoring of pressure and temperature in observation wells together with measures of ground subsidence may provide constraints on the time scale necessary for pressure and temperature to diffuse across the reservoir and on the resulting amplitude of the deformation. This would allow a still improved view of the influence of poro-elastic and thermal processes on ground deformation in the area.

# CONCLUSION

Ground deformation at the Reykjanes geothermal area during the 2015-2017 period is well constrained by by interferometric analysis of Sentinel-1A and 1B radar images. LOS changes average to 16-20 mm/yr in the satellite LOS in the selected area of maximum deformation in both ascending and descending satellite tracks, with a maximum vertical displacement of about 25 mm/yr. When results from the two tracks are combined, the observations reveal a subsidence bowl centered on the well field together with a horizontal contraction toward the center of the deforming area. This subsidence can be fit with different sources of simple geometries, including a penny shaped crack model and a horizontal square layer, contracting at an average rate of (0.7-0.9)×105 m3/yr at about 1200 m depth. Models of deformation mechanisms reveal that the volume change during the 2015-2017 period can be explained by steam cap processes in the upper part of the geothermal reservoir, above 1200 m depth. These are characterized by a combination of a compaction of the estimated 400 m thick steam cap under a pressure decrease of 0.15 MPa/yr and by thermal contraction related to a 4-5°C/yr cooling of rocks within or close to the steam cap.

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