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Mass transfer processes in a post eruption hydrothermal system: Parameterisation of microgravity changes at Te Maari craters, New Zealand

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A R T I C L E I N F O

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A B S T R A C T

Fluid transfer and ground deformation at hydrothermal systems occur both as a precursor to, or as a result of, an eruption. Typically studies focus on pre-eruption changes to understand the likelihood of unrest leading to eruption; however, monitoring post-eruption changes is important for tracking the return of the system towards background activity. Here we describe processes occurring in a hydrothermal system following the 2012 eruption of Upper Te Maari crater on Mt Tongariro, New Zealand, from observations of microgravity change and deformation. Our aim is to assess the post-eruption recovery of the system, to provide a baseline for long-term monitoring. Residual microgravity anomalies of up to 92 ± 11 μGal per year are accompanied by up to 0.037 ± 0.01 m subsidence. We model microgravity changes using analytic solutions to determine the most likely geometry and source location. A multiobjective inversion tests whether the gravity change models are consistent with the observed deformation. We conclude that the source of subsidence is separate from the location of mass addition. From this unusual combination of observations, we develop a conceptual model of fluid transfer within a condensate layer, occurring in response to eruption-driven pressure changes. We find that depressurisation drives the evacuation of pore fluid, either exiting the system completely as vapour through newly created vents and fumaroles, or migrating to shallower levels where it accumulates in empty pore space, resulting in positive gravity changes. Evacuated pores then collapse, causing subsidence. In addition we find that significant mass addition occurs from influx of meteoric fluids through the fractured hydrothermal seal. Long-term combined microgravity and deformation monitoring will allow us to track the resealing and re-pressurisation of the hydrothermal system and assess what hazard it presents to thousands of hikers who annually traverse the volcano, within 2 km of the eruption site.

1. Introduction

Locally hazardous, phreatic eruptions can occur with little warning from pressurisation of a hydrothermal system following even small intrusions of magma, or from release of volatiles caused by earthquake shaking (e.g. Christenson et al., 2007). While hydrovolcanic eruptions account for around 5% of the eruptions listed by the Global Volcanism Program, they are responsible for around 20% of the deaths related to historic eruptions (Sano et al., 2015). As such, knowing the current pressurisation state of a volcano hydrothermal system is critical for monitoring, hazard assessment, and risk mitigation.

Microgravity is a useful tool to monitor hydrothermal systems (e.g. Carbone et al., 2017), as it can detect mass changes that can be interpreted in terms of fluid flux or phase changes (condensation, etc.), as well as distinguish between fluids of magmatic or hydrothermal origin (Battaglia et al., 2006). While deformation measurements alone can locate pressure sources, interpreting those sources in terms of fluid movement, or thermal changes, in poroelastic medium is more challenging (Fournier and Chardot, 2012). In hydrothermal systems, microgravity allows reinterpretation of pressure based deformation models, in terms of fluid transfer in a porous medium, using analytic (e.g. Allis and Hunt, 1986) or numerical models (Rinaldi et al., 2011; Coco et al., 2016; Currenti and Napoli, 2017). Additionally, the combination of deformation and microgravity can

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determine if a single source model can explain both sets of observations or if there are multiple processes occurring, requiring a combination of sources (e.g. Miller et al., 2017). Hence, tracking changes in both deformation and microgravity sources over time provides a richer understanding of the system dynamics.

Mount Tongariro Volcanic Centre hosts a shallow, vapour dominated hydrothermal system (Walsh et al., 1998), as mapped from the extent of demagnetised and electrically conductive rocks (Hill et al., 2015; Miller and Williams-Jones, 2016). The hydrothermal system is likely confined to more permeable volcanic material above a low permeability greywacke basement surface, modelled at a depth of around 0 m a.s.l. beneath the Te Maari area (Miller and Williams-Jones, 2016). Upper Te Maari crater, on Mt Tongariro, erupted just before midnight (NZST) on the 6th August 2012 (Crouch et al., 2014), after 24 days of unrest (Hurst et al., 2014) that followed 125 years of quiescence (Scott and Potter, 2014). On the 21st November 2012, upper Te Maari erupted again at 1 pm, this time without warning. Ballistic blocks from the August eruption punctured the roof and floor of an unoccupied hiking hut 2 km away. Dyke injection into the base of the hydrothermal system (Christenson et al., 2013) increased pore pressures within mechanically weak, hydrothermally altered rocks (Montanaro et al., 2016). A landslide triggered the eruption by unloading the system by ~0.5 MPa (Procter et al., 2014) resulting in rapid boiling and expansion of the over-pressurised hydrothermal fluids. The intruded volume of magma was small enough that no local deformation was detected prior to eruption (Jolly et al., 2014), and no traces of fresh magma were found in the erupted deposits (Pardo et al., 2014). The eruption enlarged the existing Upper Te Maari crater and created a fissure 30 m deep and 400 m long (Fig. 1), south of the main crater.

Hamling et al. (2016) modelled a post eruption subsidence trend at Te Maari from persistent scatter InSAR data, and suggested it was due to de-pressurisation of the hydrothermal system. They modelled the deformation as the closing of a horizontal sill-like crack, located at around 500 m depth beneath Upper Te Maari. We extend that work with new microgravity, InSAR and fumarole temperature measurements taken after the eruption, to better understand the processes driving the observed subsidence and gravity increases. We perform an extensive suite of analytic solutions to test the likely source parameter distribution and to determine if subsidence and gravity change sources are co-located. We initially model the gravity and deformation data independently to determine the optimal solutions for each dataset. We then investigate joint inversion, to see if a single source, and hence process, can explain both sets of observations. We use a numerical, finite element, forward model to test if the effects of topography and geological heterogeneities are significant. Finally, we semi-quantitatively describe the likely combination of processes occurring within the hydrothermal system that contribute to the observed data. Our aim is to gain a fuller understanding of post-eruption hydrothermal processes, to better inform volcano monitoring efforts and hazard assessments at Te Maari and similar volcanoes.

2. Data

2.1. Gravity measurements

In February 2014 we established a network of 14 gravity benchmarks on accessible ground to the north and east of Upper Te Maari crater, at a spacing of 100 to 300 m (Fig. 1). Steep ground to the south of the Upper Te Maari crater limits data coverage in this area. We installed benchmarks by driving stainless steel rods to refusal into the pyroclastic deposits from the 2012 eruptions. We installed a reference benchmark, TGKB, off the volcano, approximately 5 km to the north, adjacent to highway 46 (Fig. 1). Access to the network is by helicopter, and then on foot. We used a single LaCoste and Romberg gravity meter (G106) for all surveys, with each benchmark surveyed independently 2 or 3 times in separate loops, from local base benchmark TGM03. We repeated the network at 11 to 12-month intervals in January 2015, January 2016 and December 2016 to determine mass changes over time. We refer to the December 2016 measurement as '2017' in places.

Using Gtools (Battaglia et al., 2012), we corrected daily measurements for Earth tide and ocean loading, along with linear drift corrections, to determine residual gravity changes relative to TGM03. TGM03 is tied to the reference benchmark TGKB each day so that all final gravity changes are relative to TGKB. We combined multiple repeat readings to a single value through a least squares adjustment, with standard deviations and standard errors calculated for each benchmark (Table 1). After 2015, benchmark TGM10 was buried by a small landslide. Because the mountainous terrain may affect the global free air gradient, we used a locally measured free air gradient of $-0.3047\pm\ 0.002$ mGal/m to correct for the observed subsidence (Battaglia et al., 2008). The locally measured gradient is only slightly different to the global average of -0.3086 mGal/m and for the small (0.01 to 0.04 m) vertical displacements measured at Te Maari, the resulting difference in the free air correction is less than the precision of the gravity meter ($<1\ \mu$Gal).

2.2. Deformation measurements and models

Hamling et al. (2016) presented a time series of COSMO-SkyMed InSAR deformation measurements at Te Maari from April 2012 to April 2015 (covering the first gravity measurement interval), and described the best fit model as a flat lying sill dislocation. To extend the timeseries to cover the last two gravity measurement intervals (January 2015 to December 2016), we use long temporal baseline interferograms using ALOS-2 InSAR data acquired between March 2015–March 2016, June 2015–June 2016 and from June 2016–June 2017.

We processed ALOS-2 data from two descending tracks in stripmap mode to measure the deformation over Tongariro using the InSAR Scientific Computing Environment (ISCE) software (Rosen et al., 2012). Topographic corrections were made using the 30 m SRTM data (Farr et al., 2007). The SRTM DEM pre-dates the eruption and around the vent area where we observe deformation there is estimated to be maximum 10 m of topographic elevation change. Based on the perpendicular baselines for each of the interferograms (fortuitously <15 m) the maximum contribution from topographic changes in the region would be 0.7 mm. The interferograms are filtered using a power-spectrum filter (Goldstein et al., 1988), and unwrapped using SNAPHU (Chen and Zebker, 2002). To model the data, the line-of-sight displacements are resampled onto a 120 m grid in the vicinity of the eruption site and every ~ 1 km in the far-field. Initially we ran the inversion based on a quadtree subsampled dataset but short wavelength regions of noise away from the crater area with high variance tended to get oversampled and the inversion generally didn't converge on a solution. For that reason we chose to sample the crater area at a higher resolution on a uniform grid. Despite the two 2015–2016 interferograms covering different time periods the deformation is almost the same. Taking the difference gives residuals over the vent area of <4 mm suggesting that the deformation is nearly constant over that period.

We follow the same modelling procedure as described in Hamling et al. (2016) and assume that the deformation can be approximated by the contraction of a horizontal tensile dislocation. We solve for the position, depth, length, width and amount of contraction for each time period, fixing the strike to 34° (Hamling et al., 2016). Parameter uncertainties are estimated using a bootstrap resampling procedure where we generate and invert 1000 perturbed datasets with the distribution of model parameters providing an estimate of their uncertainty. We then calculate the vertical displacement