Microgravity changes at the Laguna del Maule volcanic field: Magma-induced stress changes facilitate mass addition

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Abstract Time-dependent, or 4-D, microgravity changes observed at the Laguna del Maule volcanic field, Chile, since 2013, indicate significant ($1.5 \times 10^{11}$ kg) ongoing mass injection. Mass injection is focused along the Troncoso fault, and subparallel structures beneath the lake at 1.5–2 km depth, and is best modeled by a vertical rectangular prism source. The low-density change (156 to 307 kg/m$^3$) and limited depth extent suggest a mechanism of hydrothermal fluid intrusion into existing voids, or voids created by the substantial uplift, rather than deeper-sourced dike intrusion of rhyolite or basalt magma. Although the gravity changes are broadly spatially coincident with ongoing surface deformation, existing models that explain the deformation are deeper sourced and cannot explain the gravity changes. To account for this discrepancy and the correspondence in time of the deformation and gravity changes, we explore a coupled magmatetectonic interaction mechanism that allows for shallow mass addition, facilitated by deeper magma injection. Computing the strain, and mean, normal, and Coulomb stress changes along northeast trending faults, caused by the opening of a sill at 5 km depth, shows an increase in strain and mean and normal stresses along these faults, coincident with the areas of mass addition. Seismic swarms in mid-2012 to the west and southwest of the mass intrusion area may be responsible for dynamically increasing permeability on the Troncoso fault, promoting influx of hydrothermal fluids, which in turn causes larger gravity changes in the 2013 to 2014 interval, compared to the subsequent intervals.

Plain Language Summary Movement of liquid below the Earth’s surface occurs in response to a variety of volcanic or tectonic processes and may result in changes that are measurable on the surface of the Earth. Understanding what causes these changes helps inform on the state of a volcanic system and how close it is to eruption. We present results of measurements showing changes in the Earth’s gravity at the Laguna del Maule volcanic field, Chile. These measurements show small increases in gravity around the lake since 2013, coincident to where the ground is being uplifted by deep magma intrusion. Using computer simulations, we find that the location of the gravity changes is at 1.5 to 2 km below the Earth’s surface and is likely caused by the intrusion of water into an area of rock that is fractured by land movement. Flow of water into this fractured rock is facilitated by pressure exerted by deeper intruding magma, opening up the rock above and allowing water to flow into the space created. Our study provides important constraints on processes that are otherwise undetectable and allow us to better understand the dynamics of an active magma system.

1. Introduction

Redistribution of mass in the Earth’s crust occurs in response to a variety of magmatic, volcanic, hydrothermal, or tectonic processes, many of which display complex interactions. Ground deformation and surface gravity changes often occur in response to mass movement and may herald a change in state of a magma system that may or may not lead to eruption [e.g., Bagnardi et al., 2014]. Mass addition can result from magma intrusion or hydrothermal processes such as fluid movement, in response to temperature and pressure gradients or changes in permeability (Manga et al. [2012] and examples therein). Local stress fields can be altered by the intrusion of magma [Troise, 2003; Amelung et al., 2007; Currenti et al., 2008; Jónsson, 2009], in turn clamping or
Figure 1. Simplified geology map of the central basin of the Laguna del Maule Volcanic Field [after Hildreth et al., 2010; Miller et al., 2017]. Microgravity benchmarks are shown as black triangles along with their benchmark number. The absolute gravity stations are shown as inverted green triangles. Station MAUL is off the map as indicated by the arrow. The red arrow and rectangle indicates the center of inflation and sill outline from Feigl et al. [2014]. Ages are from $^{40}$Ar/$^{39}$Ar ratios [Andersen et al., 2017], and red stars show postglacial vents. Faults are shown in black lines with lake faults from Peterson et al. [2016]. The smaller location map shows LdMVF as a red ellipse, with other Holocene volcanoes as yellow triangles.

Unclamping nearby faults [Roman and Heron, 2007]. This, in turn, alters the permeability of faults, promoting or inhibiting fluid migration into them [Hautmann et al., 2010; Strehlow et al., 2015]. Distinguishing between magmatic and hydrothermal causes of volcanic unrest, and understanding the role tectonics plays in modulating them, is important for understanding the potential hazard the unrest represents.

In many cases [e.g., Battaglia et al., 2003; Currenti, 2008; Tizzani et al., 2009; Greco et al., 2016], observed deformation and gravity changes are shown to be caused by the same source. Joint inversion of those data yields the absolute density of the source fluid, allowing magmatic versus hydrothermal processes to be distinguished. However, in some cases, gravity and deformation signals are not produced from the same source [e.g., Johnson et al., 2010] and shallow fluid injection may be related to permeability changes induced by deeper magma intrusion [Strehlow et al., 2015]. Interpreting the two as resulting from a single source may mask other processes at work, resulting in an incomplete understanding of the system dynamics.

The Laguna del Maule volcanic field (LdMVF, Figure 1), is a large silicic, multivent volcanic center surrounding a 23 km by 16.5 km lake basin [Singer et al., 2014]. While eruptive products span the compositional range from basalt to rhyolite, since 25 ka, rhyolite has dominated, with at least 50 silicic (rhyodacite to rhyolite) eruptions nested in a concentric ring around the lake [Hildreth et al., 2010; Andersen et al., 2017]. Since 2007, large-scale deformation at rates of $>20$ cm/yr [Feigl et al., 2014; Le Mével et al., 2015, 2016] have been observed and modeled as an inflating sill at $\sim$5 km depth. Miller et al. [2017] interpreted a 19 mGal Bouguer gravity low, centered over the deformation source, as resulting from a shallow, crystal-poor, volatile-rich, silicic magma system (density 1800 kg/m$^3$) overlying the sill. Seismic reflection and magnetic studies [Peterson et al., 2016] map numerous northeast to southwest trending fault structures in close proximity to the imaged magma system.
In this paper, we present the first time series of time-dependent microgravity measurements [Battaglia et al., 2008] at LdMVF made between 2013 and 2016. We investigate the likely source geometry and illustrate how the source of ongoing deformation interacts with nearby faults, producing stress and strain changes that locally increase permeability, resulting in mass addition that is spatially separate from the deformation source and previously imaged magma system. We show that microgravity measurements add an important constraint on interpreting volcano deformation data, as they are sensitive to mass and density changes and can reveal other active processes that do not cause ground deformation, therefore helping to more completely define the dynamics of active magma and tectonic systems.

2. Gravity Measurements and Processing

In March 2013, we established a network of 35 gravity benchmarks around the lake, with a spacing of 1 to 2 km, decreasing benchmark spacing closer to the center of uplift as initially shown by Fournier et al. [2010] (Figure 1). We repeated the network at 10–14 month intervals in 2014, 2015, and 2016 to determine mass changes over time. To mitigate tares to the gravity meters from boat, foot, and vehicle travel, we measured each benchmark simultaneously with three LaCoste and Romberg meters. Over the course of the study period, five different gravity meters were used: D217, EG26, G19, G127, and G943. Meter D217 was used in every survey, and we calibrate the other gravity meters to D217 to account for small but significant differences between meters. The calibration is updated for each survey to account for long-term drifts in the instruments between surveys (see supporting information Table S1). Using Gtools [Battaglia et al., 2012], we correct daily measurements for Earth tide and ocean loading, along with linear drift corrections, to provide readings relative to benchmark BASE. In 2013, 2014, and 2015 the gravity values were referenced to benchmark GLMREF2, outside the deformation area; however, this benchmark was destroyed prior to the 2016 survey. As such, we re-referenced the 2015 data to benchmark GLM21, which is also outside the deforming area, for the 2015 to 2016 interval; GLM21 is used as reference for the 2016 survey. We did not reference the entire data set to GLM21 as in previous years this station was relatively poorly constrained, with few independent measurements. In 2016 we made seven independent measurements of GLM21 (supporting information Figure S2) to ensure that the error associated with the new reference is tightly constrained. Standard errors on the individual gravity measurements average 14 μGal for the whole dataset. When the full error budget is calculated on the gravity change data (including height change errors), the average standard error across the entire data set is 19 μGal with a maximum of 47 μGal. Further details and data (Table S2) are given in the supporting information.

In 2012 and 2015, absolute gravity stations MAUL and LDMA were established in the Maule valley by Institut de Recherche pour le Développement (IRD) (S. Bonvalot, personal communication, 2015), allowing us to check the absolute calibration of the D217 instrument. Repeat ties between the two absolute stations result in a calibration factor of 1.0005 for D217. However, as the D217 calibration has also likely changed over the course of the study period, we do not recalibrate the 2016 data to absolute, so that the previous years’ data remain consistent and relative to D217. For the 2016 data, we calculate a mean difference between absolute and relative gravity differences for D217, of 10 μGal, over the gravity range of 40 mGal. The residual gravity change (Δg) is the difference between yearly gravity measurements, corrected for benchmark uplift and lake level variations [see Battaglia et al., 2008, and references therein].

2.1. Free-Air Gradient and Uplift Correction

The substantial uplift occurring between survey periods (~0.7 m between 2013 and 2016) must be corrected to determine the residual gravity change due only to mass addition or subtraction. In 2014 and 2015, we measured the free-air gradient at BASE to determine the local gradient for use in correcting the gravity data for the uplift. Calculating a local gradient is important [Rymer, 1994] because if it differs greatly from the global average (~308.6 μGal/m), interpretation of the data may be less meaningful [Battaglia et al., 2008]. In addition, the local gradient may vary significantly across a volcanic area [de Zeeuw-van Dalfsen et al., 2005]; however, logistics prevented us from making gradient measurements at other locations. The local free-air gradient is mostly affected by local Bouguer anomalies, such as the 19 mGal low imaged beneath the lake by Miller et al. [2017]. As most of our microgravity benchmarks are approximately equidistant from this anomaly, local variations in the free-air gradient are expected to be minor. At LdMVF, some benchmarks have experienced over 0.25 m of uplift between survey periods, resulting in a large free-air correction. In 2014 and 2015, we measured a local free-air gradient of −332.5 ± 4 μGal/m and −338.7 ± 0.5 μGal/m, respectively. We use a value of −335 μGal/m for each measurement interval. Compared with the global value, this results in an increased...
correction of 5 μGal/m at benchmarks experiencing the greatest uplift. For example, in the 2015 to 2016 interval, benchmark GLM06 experienced 0.187 m uplift, requiring a free-air correction of 63 μGal.

We apply the free-air gradient to height change derived from a combination of interferometric synthetic aperture radar (InSAR) and Continuous Global Positioning System (CGPS) data (Table S2 in supporting information). For each time interval, the InSAR-derived velocity field from Feigl et al. [2014] and Le Mével et al. [2015] was updated with new data covering the study period and is used in combination with the daily position data from five continuous GPS stations to estimate the height change between surveys at each gravity benchmark. An affine transformation is calculated to adjust the InSAR-derived velocities to the GPS velocities, to resolve discrepancies in the choice of reference pixel, and in the amount of residual postseismic deformation (more details in Feigl et al. [2014]). This adjustment uses a standard weighted least squares algorithm to estimate nine parameters including the partial derivatives of the three components of the vector velocity field with respect to the two horizontal coordinates. We then extract the height change from this model at each benchmark location. We conservatively estimate the height change error between each time interval to be 1 cm, based on the uncertainties in the observed CGPS time series [Le Mével et al., 2015], which produces a gravity error of < 3 μGal.

2.2. Lake Level and Water Table Variations
The LdMVF surrounds a 54 km² lake that is up to 50 m deep. The elevation of the lake surface is artificially controlled and can vary by several meters throughout the year. Daily records of the lake elevation are kept by the dam operators (El Ministerio de Obras Publicas,Direccion de Obras Hidraulicas), and we use these to apply a correction for the changes in mass caused by changes in lake level. See supporting information Figure S1.

Between the 2013 and 2014 surveys, the lake level rose by 1.1 m adding ~ 75 × 10⁶ kg of water. To correct for this extra mass, we calculate the gravitational effect at each benchmark, of a lake-shaped polygon of water 1.1 m thick [Talwani and Ewing, 1960] and subtract it from the 2014 gravity value for each benchmark. The maximum correction applied for the lake change between 2013 and 2014 surveys is +11 μGal at benchmark GLM22. In 2015, the lake level was 0.32 m lower than in 2014. The maximum gravity effect of this change is −1 μGal which we subtract from the 2015 data. In 2016, the lake rose 4.0 m after the 2015 survey causing a maximum gravity increase of +28 μGal at GLM22, which we subtract from the 2016 data.

The level of the groundwater table around the lake is not monitored and is likely to be highly complex, with perched aquifers within individual lava flows, as observed from springs located at different elevations around the lake. We assume that the lake level is the main control on the level of larger aquifers in the basin and that these aquifers are recharged from rainfall and snow melt. Using a Bouguer slab approximation for small local aquifers would overestimate the groundwater contribution (the Bouguer slab correction is approximately 8 μGal per meter of groundwater change in an aquifer with 25% porosity), and in the absence of groundwater data, we do not apply a separate correction for the local water table. Our surveys are undertaken in the Austral summer months, so we assume that local aquifers are in a similar state from year to year.

3. Residual Gravity Change Results
We evaluate the residual gravity changes, Δg, over three time periods between four surveys in March 2013, January 2014, March 2015, and February 2016 (Figures 2a – 2c). Data from some benchmarks were not usable in particular years, so not all time intervals contain the same benchmarks. Figure 2d shows the vertical component of the displacement for continuous GPS station MAU2, as well as the number of earthquakes per day in the LdMVF region from the Observatorio Volcanologico de los Andes del Sur (OVDAS) catalogue. Supporting information Figure S2 shows a summary of benchmark occupations per year. Positive gravity changes, indicating mass addition, occur between each gravity survey.

3.1. From 2013 to 2014
Residual gravity changes in the 10 months between the 2013 and 2014 surveys, recorded at 26 benchmarks, show a maximum of 124 ± 12 μGal, at benchmark GLM06 (Figure 2a). The gravity increase is distributed in a circular pattern around the maximum, with a slight elongation to the southwest. Δg is zero around the north shore of the lake; however, the anomaly is open to the south where values of 10–20 μGal are recorded at the distal benchmarks.

Using Gauss’s theorem, where

\[ \Delta M = \frac{1}{2\pi G} \int \int \Delta g(x, y) \, dx \, dy \]
we calculate the excess mass, $\Delta M$, required to produce the gravity change anomaly, $\Delta g$, of $9.2 \times 10^{10}$ kg, where $G$ is the gravitational constant, contained within an area $(x, y)$ of 122 km$^2$. This mass is a minimum as the anomaly is not complete and is independent of the geometry of the body within which it is contained.

To determine the statistical significance of $\Delta g$, we run a nonparametric Wilcoxon signed-rank test [Helsel and Hirsch, 1992]. For an alpha = 0.05 significance level (or at 95% confidence level), if the $p$ value is < 0.05, then the null hypothesis that there is no gravity change difference between the years can be rejected. The test shows $p = 0.004$ for the 2013 and 2014 surveys, and therefore $\Delta g$ for this period is statistically significant.

3.2. From 2014 to 2015

Residual gravity changes in the 14 months between the 2014 and 2015 surveys at 28 benchmarks show a double-peaked spatial distribution (Figure 2b). The larger, northeast to southwest oriented anomaly to the south reaches an amplitude of $60 \pm 15 \mu$Gal at benchmark GLM27, while to the north, a smaller northwest to southeast anomaly has a maximum amplitude of $32 \pm 12 \mu$Gal at benchmark GLM34. The larger northeast to southwest anomaly is coincident with the Troncoso fault (Figure 1). A Wilcoxon signed-rank test on the whole data set gives $p = 0.012$. Computing the test separately on the benchmarks comprising the two individual anomalies shows, for the larger anomaly, that the 2015 data are significantly different from the 2014 data.
(p = 0.004). However, for the smaller anomaly, p = 0.06 and using the criteria of alpha = 0.05, we accept the null hypothesis of no difference between the 2014 and 2015 data at benchmarks that produce the small anomaly and thus do not further consider this anomaly. A negative anomaly associated with benchmarks GLM16, GLM17, and GLM01 in the northwest of the area is thought to be related to ongoing road works and tunneling activities as part of a hydroelectricity construction project. For the larger anomaly, we calculate a minimum excess mass of $3.96 \times 10^{10}$ kg over an area of 97 km$^2$.

### 3.3. From 2015 to 2016

Residual gravity changes in the 11 months between the 2015 and 2016 surveys at 30 benchmarks show an elliptical northeast to southwest oriented anomaly with a peak amplitude of $68 \pm 16 \mu$Gal at benchmark GLM07 (Figure 2c). This anomaly is closed around the zero contour except for a small region west of the lake which closes to the 10 $\mu$Gal contour. The positive anomaly is surrounded by a ring of small negative gravity change with a minimum of $-37 \pm 26 \mu$Gal at benchmark GLM25. A Wilcoxon signed-rank test gives $p = 0.017$ (alpha = 0.05), indicating a significant gravity difference between the 2015 and 2016 results. We calculate an excess mass of $3.96 \times 10^{10}$ kg over an area of 88 km$^2$.

### 4. Residual Gravity Change Modeling

To begin, we assess whether the model of Feigl et al. [2014] to explain the deformation changes can replicate the observed gravity changes, when the volume change geometry is filled with mass.

Feigl et al. [2014] propose a rectangular, sill-like body, with an annual tensile opening rate of $\sim 1$ m/yr at a depth of $\sim 5$ km (or $-3$ km above sea level, asl), to explain the deformation trends observed in an InSAR time series from 2007 to 2014. To test whether the sill model can explain our gravity data, we construct a polygon with the same dimensions as the inflating sill and density 2700 kg/m$^3$, i.e., that of basalt magma [Le Mével et al., 2016]. The maximum gravity effect calculated is $\sim 3 \mu$Gal, which is both below our surveys detection limit and much lower than our observed data. In order to approximate the observed $\Delta g$, we increase the sill density while keeping the sill thickness constant and then increase the sill thickness while keeping the density constant. A 1 m thick sill requires an unrealistic density of around 15,000 kg/m$^3$, while keeping the sill at density 2700 kg/m$^3$ requires a sill opening of around 40 m. Therefore, the intrusion of mafic magma at 5 km depth is not able to explain our gravity change results. The observed deformation is completely accounted for by the sill model [Feigl et al., 2014]; consequently, the gravity source is spatially independent of the deformation source, and the cause of the gravity changes produce no additional deformation.

The implication of separate deformation and gravity sources is that we are only able to calculate the mass change from the gravity data. To calculate the true density of the gravity change source requires the deformation and gravity bodies to be colocated, so that volume change is independently calculated from the deformation and mass change is calculated from the gravity, from which we can derive the true density of the source. As we have no independent constraint on volume changes associated with the gravity change, the gravity change only reflects the mass change of the crust between each survey. Additionally, our gravity-only models do not include the effects of internal boundary displacements caused by inflating sources [Currenti et al., 2007].

#### 4.1. Inversion Method

We use a genetic algorithm (GA) [e.g., Tiampo et al., 2000; Carbone et al., 2008] to invert for source parameters of several geometric shapes: sphere, prolate spheroid, oblate spheroid, triaxial spheroid, and vertical rectangular prism Clark et al. [1986]. Genetic algorithms are a class of Monte Carlo algorithm [Sambridge and Mosegaard, 2002] and are effective at searching model parameter space for optimal solutions, especially where the objective function may have several local minima [Tiampo et al., 2000]. Local minima often occur in potential field data where nonunique solutions are inevitable. The GA randomly generates an initial population of 100 sets of model parameters within a predefined range. This range is defined to limit the search to geologically realistic domains. We iteratively evolve this population by mimicking genetic evolutionary processes for 150 generations, keeping only the best fit models until a single model remains. To test the sensitivity of the algorithm to different starting models, we repeat the entire procedure 10,000 times for each geometry. In this way we generate populations of 10,000 final models, where each individual model has been selected from 100 individuals chosen from a randomized starting model. This approach allows us to test the stability and sensitivity of the models over a wide range of randomized starting parameters and determine which geometry best explains the observed data.
4.2. Source Models

We explore a range of analytic models to determine likely source geometries and quantify the mass addition causing the observed gravity changes. Given the large survey area with relatively few survey points, especially over the lake, using analytical solutions for simple geometries is easily justified to approximate the gravity sources. For each measurement interval, we model the following geometries, with the number of fitted parameters indicated in parentheses: sphere (5), prolate spheroid (8), triaxial spheroid (9), oblate triaxial spheroid (9), and prism (8). The sphere geometry is defined by its centroid coordinates (Xc, Yc, depth to center), axis radius and density change. The prolate spheroid is defined by its centroid coordinates (Xc, Yc, depth to center), major axis radius, ratio of length of minor axis to major axis, strike angle, and density change. The triaxial spheroid is the same as the prolate spheroid except that both minor axes are free to adjust. The oblate triaxial spheroid restricts one minor axis ratio so that an oblate geometry is produced. The prism geometry (Figure 3) is defined by its centroid coordinates (Xc, Yc, depth to top), length, width, strike, thickness, and density change, with dip fixed at 90°. The location of the source centroid is confined to be within the gravity anomaly area, and the depth is constrained from 2000 m asl to −5000 m asl.

To determine the best fit source geometry, after removing nonphysical models from the population, we calculate the reduced chi-square ($\chi^2_{\text{red}}$) statistic on each of the 10,000 models for each geometry, defined as

$$\chi^2_{\text{red}} = \frac{1}{\nu} \sum_{k=1}^{n} \frac{(O_k - E_k)^2}{\sigma^2_k}$$

where $O_k$ are the observed data, $E_k$ are the calculated data, $\sigma_k$ is the standard deviation of the observation, $\nu$ is the number of degrees of freedom given by $N - n$ where $N$ is the number of observations and $n$ is the number of fitted parameters, which varies between model geometries. We use an $F$ test [Wackerly et al., 2007] on the $\chi^2_{\text{red}}$ values to determine which model geometry population best fits the data. Figure 4 shows box plots of $\chi^2_{\text{red}}$ calculated for each model type, grouped by observation interval. We additionally calculate the RMS fit ($\mu$Gal) for each model.

To assess the variability of each parameter, we calculate the kernel density estimate (KDE) with a Gaussian kernel operator [Silverman, 1986], where the bandwidth for the KDE is chosen using Scott’s method [Scott, 1992]. We refer to the peak of the KDE as the mode from here on.

4.3. The 2013 to 2014 Source Model Results

The simple sphere has the lowest $\chi^2_{\text{red}}$ of the spheroid-like models, indicating that more complex spheroids offer no improvement to determining the geometry. An $F$ test comparing the sphere and prism shows that the prism models are a better fit at $p < 0.01$ ($F = 725$). The mode RMS fit of the prism model population is 10 $\mu$Gal. The prism models are most sensitive to the thickness and mass change parameters and less sensitive to the length, depth, and width parameters. The length and depth are well constrained by the benchmark distribution, while the width (the bottom of the prism) is less well constrained, being further from the surface measurements. Table 1 shows a summary of the mode prism model parameters.

A summary of the prism model parameters and locations is shown in Figures 5a and 5c–5h with the fit of the models to the observed data shown in Figure 5b. The KDE plots show a bimodal distribution of parameters and $\chi^2_{\text{red}}$, indicating two possible source models. The lowest $\chi^2_{\text{red}}$ population models (labeled I) are shallower, with lower density change (307 kg/m$^3$) and more northeast to southwest orientation (43° strike) than the
Figure 4. Box plots of $\chi^2_{\text{red}}$ of the fit of the model to the data, grouped by each model geometry. In each geometry segment the three observation intervals are shown, as labeled for the sphere model. The box shows the quartiles of the data set, while the whiskers extend to show the rest of the distribution. Individual dots are outliers calculated using a method that is a function of the interquartile range. In most cases there is very tight clustering of $\chi^2_{\text{red}}$ values, making the box and whiskers less obvious.

Higher $\chi^2_{\text{red}}$ model population (labeled II) which are deeper and oriented more east-west (65° strike). The lower $\chi^2_{\text{red}}$ prism models have a mode elevation of 543 m asl, and nearly all are oriented subparallel and coincident with faults imaged by seismic reflection [Peterson et al., 2016]. The mode mass increase of the lower $\chi^2_{\text{red}}$ prism model is $1.27 \times 10^{11}$ kg (cf. $9.2 \times 10^{10}$ kg from Gaussian integration), with mode width of 170 m and mode thickness of 145 m. The higher $\chi^2_{\text{red}}$ model group has a mode mass increase of $1.58 \times 10^{11}$ kg and mode thickness of 43 m. The two model populations reflect the common problem of equivalence in potential field data, where more than one model will fit any data set; however, we prefer the population with the lowest $\chi^2_{\text{red}}$.

### 4.4. The 2014 to 2015 Source Model Results

For the 2014 to 2015 time interval we model the subset of stations that cause the larger anomaly to the south (Figure 2b), as the smaller anomaly to the north is shown to be not statistically significant. We remove the

| Table 1. Summary of Model Parameters for Each Observation Interval From Peak of KDE Distribution |
|-----------------------------------------------|---|---|---|---|
| **Prism KDE Peak Values** | **2013 to 2014** | **2014 to 2015** | **2015 to 2016** |
| | **Model I** | **Model II** | **Model I** | **Model II** | **Model I** | **Model II** |
| Xc (m) | 363,544 | 364,286 | 361,192 | 363,615 |
| Yc (m) | 6,007,447 | 6,007,370 | 6,005,950 | 6,008,391 |
| Elevation (m asl) | 543 | 250 | 875 | 619 |
| Length (km) | 6.2 | 6.2 | 6.4 | 5.9 |
| Width (m) | 170 | 170 | 110 | 110 |
| Thickness (m) | 145 | 43 | 66 | 31 |
| Strike (deg) | 43 | 65 | 40 | 76 |
| Density change (kg/m$^3$) | 307 | 884 | 156 | 279 |
| Mass change (kg) | $1.27 \times 10^{11}$ | $1.58 \times 10^{11}$ | $3.85 \times 10^{10}$ | $5.08 \times 10^{10}$ |
| Flux, Q (m$^3$/s) | 4.9 | 6.14 | 1.1 | 1.7 |
| Hydraulic conductivity, $K$ (m/s) | $3.33 \times 10^{-6}$ | $1.23 \times 10^{-6}$ | $4.96 \times 10^{-7}$ | $4.05 \times 10^{-7}$ |
| Transmissivity, $T$ (m$^2$/s) | $5.65 \times 10^{-4}$ | $2.09 \times 10^{-4}$ | $5.45 \times 10^{-5}$ | $4.45 \times 10^{-5}$ |
| Permeability, $k$ (m$^2$) | $3.3 \times 10^{-11}$ | $1.2 \times 10^{-11}$ | $4.9 \times 10^{-12}$ | $4 \times 10^{-12}$ |

*Lake surface elevation is 2160 m asl. Model I is the lower $\chi^2_{\text{red}}$ models, and model II is the higher $\chi^2_{\text{red}}$ models for the 2013 to 2014 interval.*
Figure 5. (a) Contour plot of $\Delta g$ for 2013 to 2014 interval in microgal. Maximum gravity change is $124 \pm 12 \mu$Gal. Overlain in green lines are the top surfaces of the vertical prism models, while blue dots show the sphere centroid locations. Each symbol represents one run of the genetic algorithm. Red lines are mapped faults onshore, and in the lake bed as mapped from seismic reflection [Peterson et al., 2016], and the dotted box is the outline of the limits of the source location in the inversion. (b) Observed (black dots with error bars), and calculated $\Delta g$ for each of the 10,000 vertical prism models. (c–h) KDE plots of vertical prism geometry parameters. The annotated value is the peak (mode) of the KDE distribution. The lower $\chi^2_{\text{red}}$ models are labeled as I and the higher $\chi^2_{\text{red}}$ models as II.

stations associated with the smaller anomaly so that they do not bias the inversion of the main anomaly. An $F$ test shows that the prism model population is a better fit than the sphere at $p < 0.01$ ($F = 1214$) and has an RMS error of 6 $\mu$Gal. A summary of the prism parameters is shown in Figures 6a and 6c–6h with the fit of the models to the observed data shown in Figure 6b. All solutions show a unimodal distribution and close association with the Troncoso fault with a mode strike of 40°. The mode elevation of the prism solutions is 875 m asl with a mode thickness of 66 m, mode width of 110 m, and mode mass increase of $3.85 \times 10^{10}$ kg (cf. $3.96 \times 10^{10}$ kg from Gaussian integration).

4.5. The 2015 to 2016 Source Model Results
A summary of prism model parameters and locations is shown in Figures 7a and 7c–7h with the fit of the models to the observed data shown in Figure 7b. The model solutions show a unimodal parameter distribution, and an $F$ test shows that the prism model population is a significantly better fit than the sphere at $p < 0.01$ ($F = 346$), with an RMS error of 20 $\mu$Gal. The larger RMS results from some benchmarks with negative $\Delta g$, which are not explicitly fitted in the inversion. The prism solutions are centered over the peninsula, oriented ESE to WSW (mode strike 76°). The strike is more east-west oriented than the previous intervals but still subparallel to faults mapped in the lake bed. The mode elevation of the prisms is 619 m asl with a mode thickness of 31 m, mode width of 110 m, and mode mass increase of $5.08 \times 10^{10}$ kg (cf. $3.96 \times 10^{10}$ kg from Gaussian integration).

5. Discussion
Our preferred vertical prism gravity models are spatially distinct from the horizontal sill deformation models of Feigl et al. [2014]. The gravity sources are spatially coincident with the Troncoso fault and have limited depth extent (width parameter), suggesting that they are confined to a thin geological layer intersected by the fault. The sill deformation model does not reproduce the observed gravity change, and conversely, there appears to
Figure 6. (a) Contour plot of $\Delta g$ for 2014 to 2015 interval in microgal. Maximum gravity change is $60 \pm 15 \mu\text{Gal}$. Overlaid in green lines are the top surfaces of the vertical prism models, while blue dots show the sphere centroid locations. Each symbol represents one run of the genetic algorithm. Red lines are mapped faults onshore, and in the lake bed as mapped from seismic reflection [Peterson et al., 2016]. The grey rectangle is the deformation source of Feigl et al. [2014], and the dotted box is the outline of the limits of the source location in the inversion. (b) Observed (black dots with error bars), and calculated $\Delta g$ for each of the 10,000 vertical prism models. (c–h) KDE plots of vertical prism geometry parameters. The annotated value is the peak (mode) of the KDE distribution.

5.1. Nature of Intruding Fluid and Mass Balance

Our models show large, shallow mass increases between each observation interval, in good agreement with independently calculated mass addition from Gaussian integration of the observed gravity anomaly. The mass addition is spatially correlated with the Troncoso fault and is above the magma source imaged by Miller et al. [2017], suggesting that either hydrothermal fluids or magmatic dike intrusion along the fault is responsible.

We calculate the density change from the mass change and the volume of the source model. The density changes per year are 156 to 307 kg/m³, suggesting that low-density hydrothermal fluid rather than denser magmatic fluids are involved. A hydrothermal brine at 20°C with 10,000 ppm total dissolved solids, and density 1005 kg/m³, filling existing and newly created void space of 30%, would result in a density change of 302 kg/m³. In 2014, measurements of CO₂ soil gas concentrations using a Vaisala GM70 probe revealed CO₂ in the range of 0.2 to 7% around the lake shore [Miller et al., 2014], suggesting that hydrothermal fluids may be gas rich. In addition, the fluid may lie above a high-temperature magma system imaged with Bouguer gravity [Miller et al., 2017]. Hydrothermal brine at 200°C, containing 5% CO₂, would have a density of 865 kg/m³, which filling 30% pore space would result in a density change of 259 kg/m³. A density change of 156 kg/m³ would similarly require filling a pore space of 18% with high-temperature hydrothermal fluid.

An alternate hypothesis is that rhyolite magma leaked from the reservoir and intruded along the fault zone. A rhyolite magma with a density of 1800 kg/m³ intruding into the fault zone as a dike would account for the
observed gravity changes if it filled ~9–17% of existing pore space. However, viscous rhyolite magma is unlikely to passively fill pore space and dike intrusion would cause significant additional deformation (see supporting information S4) and seismicity, as well as having a greater depth extent in the model geometry, which appears to be confined to a thin geological layer. Models of magma chamber overpressure from gravity and deformation measurements by Le Mèvel et al. [2016] and Miller et al. [2017] suggest that current pressures are not sufficient to generate dyking in the roof of the reservoir. As dike-induced deformation or seismicity are not observed, we do not consider magma to be a likely fluid causing the gravity changes.

In a study of ignimbrite and rhyolite lavas from the Chon-Aike Province in Patagonia Sruoga et al. [2004] reported porosities of 2 to 38%. Smyth and Sharp [2006] reported porosities up to 50% from the Topopah Spring tuff at Yucca Mountain. Nonwelded ignimbrites and autobrecciated rhyolites had the highest porosities, while welded ignimbrites and massive rhyolites had the lowest. Our estimates of 18–30% porosity filling required to explain the gravity anomalies are well within these reported ranges, and the higher porosity values support our model of hydrothermal, rather than magmatic, fluids as the mass source. We also assume 100% saturation of the pore space; however, incomplete saturation of the pore space would require a large porosity, which would be less likely given that our estimated 30% porosity is generally at the higher end of measured values.

The issue of mass balance arises when we consider where the intruding fluid originates. Lateral transport of fluid into the fault zone would create mass depletion where the fluid is sourced, causing a negative gravity change equal to the positive gravity change imaged along the fault zone. In the 2015 to 2016 interval (Figure 7c), there is a weak gravity low (minimum $-37 \pm 26 \mu $Gal) surrounding the gravity high, that may be the signal of a deeper source region. As there is no negative gravity change of equal amplitude to the positive gravity change (68 ± 16 μGal), it is likely that the fluids are sourced from depth, below the fault zone, where the
effect of fluid withdrawal from a deep source region is less sensitive to our surface gravity measurements. This would be consistent with a model of hot, buoyant hydrothermal fluids rising into the fault zone. Our observed positive gravity anomaly is likely a composite signal dominated by the shallow mass addition signal but also containing a smaller deeper-sourced mass withdrawal component.

5.2. Stress Change Mechanism for Mass Emplacement

To account for mass injection without additional surface deformation requires a mechanism that allows mass to move into existing or newly created space. Such space may represent empty pore space, pore space created by the inflating sill, or increased permeability from fracturing around fault zones [e.g., Sibson, 1994; Curewitz and Karson, 1997]. Manga et al. [2012] reviewed mechanisms of increasing permeability caused by oscillations in stress, such as those created by earthquakes, and found that minimum strain amplitudes of $10^{-6}$ can result in changed water levels in wells, or increased production from petroleum reservoirs. Enhanced permeability was found to exist for periods of months to years following the stress change event. Mean strain rates measured at LdMVF are $3 \times 10^{-6}$/yr for the horizontal velocities and $20 \times 10^{-6}$ for the vertical velocities [Feigl et al., 2014].

The Troncoso fault bounds the western margin of the geophysically imaged LdMVF magmatic system [Miller et al., 2017] and overlies the sill modeled by Feigl et al. [2014] to account for the deformation. The close proximity of the microgravity model solutions to the Troncoso fault, other faults imaged beneath the lake [Peterson et al., 2016], and LdMVF magma system suggests a causative relationship between the fault geometry, deformation source, and gravity changes.

We test the hypothesis that sill opening at depth changes the stress field around these faults, increasing permeability and lowering the pore pressure within them, allowing fluids to migrate into the faulted areas. An increase in normal stress unclamps and increases volume around the fault zone, producing a region of low pore pressure within the fault [e.g., Vigny, 2002], promoting migration of fluids into the low-pressure area. Increases in mean stress promote fluid flow toward normal faults [Sibson, 1994], while strong permeability changes may also be induced by mean effective stress changes, namely, the mean stress minus pore pressure, through exponential or power law functions [Hummel and Shapiro, 2012]. We explore these scenarios by calculating the strain, and mean, normal, and Coulomb failure stresses (CFS) [Lin, 2004; Toda et al., 2005], caused by sill opening.

The displacements are calculated in an elastic half-space, and the Coulomb failure stress (CFS) is given by

$$\Delta \sigma_f = \Delta \tau_s + \mu' \Delta \sigma_n$$

where $\Delta \sigma_f$ is the change in failure stress on the receiver fault, caused by slip on the source fault (in this case the inflating sill), when $\Delta \tau_s$ is the change in shear stress, $\Delta \sigma_n$ is the change in normal stress, and $\mu'$ is the effective coefficient of friction on the fault, set at 0.4. Poisson’s ratio is set to 0.25 and Young’s modulus to 80 GPa [Toda et al., 2005].

For volume creation around the fault zone, we are mostly interested in the change in normal stress, where a positive change in normal stress represents fault unclamping. An increase in CFS, caused by increased normal stress, also results in volume creation and fault unclamping.

Change in mean stress, $\Delta \overline{\sigma}$, is calculated as

$$\Delta \overline{\sigma} = \frac{\Delta \sigma_{xx} + \Delta \sigma_{yy} + \Delta \sigma_{zz}}{3}$$

The model of Miller et al. [2017] of a high-melt percentage magma reservoir suggests that an isotropic elastic half-space, assumed by the Coulomb code, may not be realistic. However, the half-space model will be appropriate for first-order determinations of stress conditions.

Our model consists of a source sill at 5 km depth, with a strike of N20°E, dipping 18° to the east and opening at 1 m/yr in accordance with the model of Feigl et al. [2014]. We model the stress changes of this sill opening on the Troncoso fault, striking N40°E; a representative lake fault determined from seismic reflection [Peterson et al., 2016], striking N50°E; and a fault mapped on the north shore of the lake, striking N61°E [Hildreth et al., 2010]. Each fault is divided into segments approximately 500 m long (along strike) and 500 m deep (down dip), to determine how stress varies along the fault.
The exact dip and sense of movement of these faults is unknown, although seismic reflection, field observations, and focal mechanisms suggest that normal faults dipping to the southeast are dominant. Earthquakes close to the Troncoso fault also indicate normal and dextral strike slip movement; however, it is uncertain if the strike slip occurs on the Troncoso or other faults (e.g., Laguna Fea fault).

First, we calculate the normal, Coulomb, and mean stress change associated with sill opening on faults oriented 40°, i.e., parallel to the Troncoso fault (Figures 8a–8c), dipping 70°E with rakes of −90, i.e., normal faulting. This fault orientation results in increases in normal stress (i.e., fault unclamping) around the traces of the faults. Second, we calculate the mean stress on the three fault planes (Figures 8d–8f) to determine how mean stress varies along strike and with depth. The gravity change source region, covering all time intervals (blue shaded), occurs in areas of normal and mean stress increase and neutral to positive Coulomb stress changes. In these areas, fault unclamping and volume increase occurs, resulting in reduced pore pressure, allowing influx of new fluids causing the observed gravity changes.

In addition to the stress changes, we calculate the dilatational (volumetric) strain caused by the sill opening (Figure 9). Volumetric strain, $\varepsilon_{\text{vol}}$, is defined as $\varepsilon_{xx} + \varepsilon_{yy} + \varepsilon_{zz}$, and positive strain represents a volume increase. Strain from the sill opening is positive to the west of the Troncoso fault and above ~2 km depth. Maximum strain of around $2.5 \times 10^{-5}$ is calculated. Manga et al. [2012] suggest that strain changes as small as $10^{-6}$ are enough to induce significant permeability changes.

The density change, $\Delta \rho$, resulting from the increased volume, is given by $\Delta \rho/\rho = -E_{\text{vol}}$. From the Bouguer gravity model of Miller et al. [2017], the background density in the fault area, $\rho$, is around 2000 kg/m$^3$, which results in a strain-induced density change of $-0.04$ kg/m$^3$. This value is negligible and can be ignored as the
Figure 9. (a–c) Map views of volumetric strain at 1, 2, and 3 km depth. Lake outline shown in blue lines, simplified faults in black lines, and deformation model of Feigl et al. [2014] in grey rectangle. Thick blue dashed line in Figure 9a shows the cross-section location. (d) East-west cross section of volumetric strain. Positive strain is volume increase. The oval and rectangular blue shaded area represents the region of mass increase from microgravity models over all time intervals.

The lateral movement of the gravity source center from year to year may reflect changing permeability and pore pressure conditions along the fault. For example, it is possible that as one part of the fault is infiltrated gravity change is dominated by mass addition. If the strain is applied over the volume of the gravity source (e.g., in 2013 to 2014, prism volume is $1.53 \times 10^8$ m$^3$), a volume increase due to strain of $3.82 \times 10^3$ m$^3$ is created. Calculating in this way, between 2014 and 2015, $1.16 \times 10^7$ m$^3$ of new volume is created and between 2015 and 2016, $5.03 \times 10^2$ m$^3$ of new volume is created. The prism gravity source volume is much larger than that produced by volumetric strain, suggesting that mostly existing empty pore space is filled to produce the gravity increase. Permeability within this pore space is enhanced by the applied stress field, and dynamic permeability may be further enhanced by shaking from regular earthquake swarms to the south on the Troncoso fault and Laguna Fea faults.

That existing unfilled permeability is filled with fluid, along with a small component of new permeability caused by sill inflation, rather than shallow fluid intrusion directly creating large amounts of new permeability, may account for the lack of shallow deformation and local seismicity often associated with intrusion of fluid along fault zones. More sophisticated poroelastic finite element models [e.g., Strehlow et al., 2015] are required to fully assess the effects of fluid transport into the fault zone and to fully test the hypothesis of no shallow, fluid-induced deformation.

The lateral movement of the gravity source center from year to year may reflect changing permeability and pore pressure conditions along the fault. For example, it is possible that as one part of the fault is infiltrated...
by fluid, it repressurizes and fluid intrusion moves to a different part of the fault where low pore pressure and open permeability exists.

The rate of mass addition dropped markedly between the first and second observation intervals. In June 2012, OVDAS recorded a volcanic tectonic earthquake swarm, to the southwest of the mass addition area with over 550 events in 10 days and magnitudes up to 3.9 (Figure 2d). We propose that this swarm temporarily increased permeability around the fault, allowing greater influx of fluid into the fault zone in the year after the swarm, compared to subsequent years. Examples in Manga et al. [2012] indicate that dynamic permeability changes can persist for 1–2 years following perturbation by transient seismic waves. In subsequent years, the effect of the earthquake swarm on permeability decreased, and subsequent gravity changes only reflect permeability caused by the ongoing sill opening. Indeed, the mass addition in 2014 to 2015 and 2015 to 2016 intervals is similar, suggesting a constant rate of fluid intrusion, related to the constant rate of sill opening. We propose that the ongoing sill opening creates a pore pressure differential across the faults and will continue to allow fluids to be intruded along them for as long as the sill opening occurs.

5.3. Fault Zone Hydraulic Conductivity and Permeability Estimates

From our gravity change models we know the amount of fluid injected into the fault zone. In addition, pressure changes from our stress model allow us to estimate hydraulic parameters of the fault zone. From Darcy’s law we calculate the hydraulic conductivity ($K$, in m/s) of the fault zone for each interval:

$$K = \frac{-Q}{A \left( \frac{\delta l}{\delta h} \right)}$$

where $Q$ is the mass flux (m$^3$/s), derived from the gravity models, $A$ (m$^2$) is the cross-sectional area, $\delta l$ is the flow path length (m), and $\delta h$ is the head change (m).

$A$ is determined from the cross-sectional area of the prism model (see Figure 3), $\delta l$ is taken as half the prism thickness, assuming that fluid infiltrates evenly from both sides. For $\delta h$, we use the normal stress change (10 bar) from the fault opening (Figure 8) converted to head (~101 m). The pressure change across the subvertical fault means that we are calculating the horizontal conductivity $K_h$ rather than the vertical (i.e., gravity driven) conductivity. We use a fluid of density 1000 kg/m$^3$ for these calculations.

For the 2013 to 2014 interval, a mass change of $1.27 \times 10^{11}$ kg in 298 days results in a flux ($Q$) of $4.9 m^3/s$. Between 2014 and 2015 surveys, $Q = 1.1 m^3/s$ from a mass change of $3.85 \times 10^{10} kg$ in 413 days, while between 2015 and 2016, $Q = 1.7 m^3/s$ from a mass increase of $5.08 \times 10^{10} kg$ in 340 days. These rates are the same order of magnitude as the magma injection rate ($1.2 m^3/s$) calculated by Le Mével et al. [2016] to explain the observed deformation, suggesting a temporal link between deep magma injection and shallow hydrothermal fluid movement.

Solving for $K_n$, $K_h = 3.33 \times 10^{-6}$, $4.96 \times 10^{-7}$, and $4.05 \times 10^{-7}$ m/s for the 2013 to 2014, 2014 to 2015, and 2015 to 2016 intervals, respectively.

We also calculate the transmissivity, $T$ (m$^2$/s), defined as $T = K \times d$, where $d$ is the aquifer thickness. We assume that the depth extent of the model (width parameter, Figure 3) is the aquifer thickness. $T$ varies from $5.65 \times 10^{-4}$ m$^2$/s in the first time interval to $4.45 \times 10^{-5}$ m$^2$/s in the second and third time intervals.

Hydraulic conductivity decreases with time and may be related to decaying magma injection opening rates [Le Mével et al., 2016], resulting in lower normal stress and reduced fault opening, causing less permeability change and inhibiting the flow of fluid. In addition, dynamic permeability changes from seismic swarms may still be decreasing as colloidal particles displaced by shaking settle back into interstitial spaces between grains, gradually reducing permeability [Manga et al., 2012].

Hydraulic conductivities of $10^{-7}$ m/s and transmissivities of $10^{-5}$ m$^2$/s are within the ranges of those reported in a compilation of rhyolite and ignimbrite aquifer properties from the Okataina Volcanic Centre [Tschritter and White, 2014] and for Yucca Mountain and other tuffs in central America [Smyth and Sharp, 2006]. They found that groundwater flow in rhyolites and welded ignimbrites is typically fracture dominated, while unconsolidated pyroclastic materials have high primary permeability. The stratigraphy at the depth of the mass addition at LdMVF is unknown but is likely to consist of alternating layers of lavas and pyroclastic material, based on observations in the canyons of the Rio Maule west of the lake. Hence, the conductivities and transmissivities we calculate are typical of volcanic material observed at similar volcanic centers.
The relationship between intrinsic permeability \( (k, \text{ in m}^2) \) and hydraulic conductivity \( (K) \) was defined by Hubbert [1956] as

\[ K = k \rho_w g \frac{\mu}{\sqrt{g}} \]

where \( \rho_w \) is the density of water \((1000 \text{ kg/m}^3)\), \( g \) is gravitational acceleration \((9.8 \text{ m/s}^2)\), and \( \mu \) is dynamic viscosity of water \((1 \times 10^{-3} \text{ kg/m/s at } 20^\circ\text{C})\). Solving for \( k \), we obtain intrinsic permeabilities of \(3.3 \times 10^{-11} \text{ m}^2\) for the 2013 to 2014 interval, \(4.9 \times 10^{-12} \text{ m}^2\) between 2014 and 2015, and \(4 \times 10^{-12} \text{ m}^2\) between 2015 and 2016 for our range of \( K \).

These permeabilities are similar to the ranges found by Smyth and Sharp [2006] for the Paintbrush Group tuffs, containing welded and unwelded units, at Yucca Mountain and by Heap et al. [2014] for tuffs from Campi Flegrei.

6. Conclusions

Positive microgravity changes, observed between January 2013 and March 2016 at Laguna del Maule volcanic field, Chile, reveal an interaction between magma intrusion, local faults, and the hydrothermal system, not discernible from deformation measurements. The location of the mass addition is coincident with the Troncoso fault and other faults mapped in the lake bed [Peterson et al., 2016]. Best fit analytic solution source models to explain positive gravity changes are vertical rectangular prisms located at approximately 1.5 km depth, up to 6.4 km long, 145 m thick, extending 170 m deep. The limited depth extent implies that the mass addition is confined to a single geological layer and is not associated with deep-rooted dike injection.

To understand the mechanism of mass emplacement, we model the normal, mean, and Coulomb stress change on faults, resulting from the opening of a sill, modeled previously by Feigl et al. [2014] at 5 km depth. Sill opening causes increases in normal and mean stresses and dilatational volumetric strain, at the depth of the gravity change models, on NE trending faults that overlie the sill. Positive volumetric strain decreases pore fluid pressure around the fault, allowing migration of new fluid into existing empty pore space and newly created voids without creating additional shallow sourced deformation. Assuming a constant mass flux rate within each time interval, we calculate hydraulic conductivity values of \(3 \times 10^{-6}\) to \(5 \times 10^{-9} \text{ m/s}\) and transmissivity of \(5 \times 10^{-8} \text{ m}^2/\text{s}\) to \(5 \times 10^{-5} \text{ m}^2/\text{s}\). Permeabilities derived from the hydraulic conductivity are on the order of \(3 \times 10^{-11}\) to \(5 \times 10^{-12} \text{ m}^2\), comparable to values measured in similar volcanic regions. Seismic swarms in 2013, to the southwest of the gravity change area, maybe have caused dynamic permeability changes around the Troncoso fault, producing increased mass addition and enhanced hydraulic conductivity and permeability in the 2013 to 2014 observation interval compared to subsequent intervals.

The localization of hydrothermal fluids along faults, and the close proximity of these faults to an active magma system, suggests that these faults may be the foci of future eruptions, and future work should be undertaken to better characterize their geometries, history, and evolution.

References


