3D gravity inversion and thermodynamic modelling reveal properties of shallow silicic magma reservoir beneath Laguna del Maule, Chile

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Abstract

Active, large volume, silicic magma systems are potentially the most hazardous form of volcanism on Earth. Knowledge of the location, size, and physical properties of silicic magma reservoirs, is therefore important for providing context in which to accurately interpret monitoring data and make informed hazard assessments. Accordingly, we present the first geophysical image of the Laguna del Maule volcanic field magmatic system, using a novel 3D inversion of gravity data constrained by thermodynamic modelling. The joint analysis of gravity and thermodynamic data allows for a rich interpretation of the magma system, and highlights the importance of considering the full thermodynamic effects on melt density, when interpreting gravity models of active magmatic systems. We image a 30 km³, low density, volatile rich magma reservoir, at around 2 km depth, containing at least 85 % melt, hosted within a broader 115 km³ body interpreted as wholly or partially crystallised (>70 % crystal) cumulate mush. Our model suggests a magmatic system with shallow, crystal poor magma, overlying deeper, crystal rich magma. Even though a large density contrast (-600 kg/m³) with the surrounding crust exists, the lithostatic load is 50 % greater

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than the magma buoyancy force, suggesting buoyancy alone is insufficient to trigger an eruption. The reservoir is adjacent to the inferred extension of the Troncoso fault and overlies the location of an intruding sill, driving present day deformation. The reservoir is in close proximity to the 2.0 km³ Nieblas (*rln*) eruption at 2-3 ka, which we calculate tapped approximately 7 % of the magma reservoir. However, we suggest that the present day magma system is not large enough to have fed all post-glacial eruptions, and that the location, or size of the system may have migrated or varied over time, with each eruption tapping only a small aliquot of the available magma. The presence of a shallow reservoir of volatile rich, near liquidus magma, in close proximity to regional scale faulting, has important implications for volcano monitoring and hazard mitigation.

Keywords: Bouguer gravity, inversion, MELTS, rhyolite, thermodynamics, volatiles

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1. Introduction

Accumulation of large volumes of silicic magma in the crust is a pre-requisite to Earth's most dangerous style of volcanic eruption (Cashman and Giordano, 2014). Detailed knowledge of the location and physical properties of that magma is therefore important in determining the potential hazard of the magma reservoir, and its likelihood to erupt. Geophysical images of active, high silica magma systems, are usually obtained using InSAR, seismic, or magnetotelluric methods (e.g. Bachmann et al., 2007; Pritchard and Gregg, 2016, and references therein). Surprisingly, gravity images are less common, in spite of the strong density contrast produced by high silica magmas with their surrounding crust (e.g. Masturyono et al., 2001; DeNosaquo et al., 2009; del Potro et al., 2013; Saxby et al., 2016). Many rhyolitic calderas have substantial Bouguer gravity anomalies but these are often caused by caldera infill (Kane et al., 1976; Davy and Caldwell, 1998), rather than the underlying magmatic system.

In this study we present a new, 3D gravity inversion scheme using the open-source Simulation and Parameter Estimation in Geophysics (SimPEG) frame-

work (Cockett et al., 2015) <http://www.simpeg.xyz>, and importantly, consider thermodynamic effects on the magma system density. Thermodynamic modelling using the free MELTS code (Gualda and Ghiorso, 2015) considers crystal, melt and volatile phases as well as pressure and temperature conditions when computing magma system densities. Consideration of the volatile phase is important for determining the total magma system density and thus for the correct interpretation of gravity inversion results, as volatiles control many processes which drive or hinder eruptions (Huppert et al., 1982; Sparks and Huppert, 1984; Wallace et al., 1995; Malfait et al., 2014).

Gravity models of volcanoes are not new, however the inversion codes are often poorly documented, proprietary, or expensive, making them difficult to be widely adopted or benchmarked. While there are popular free codes used in volcanology (e.g. Camacho et al., 2011), a growing number of open-source codes is developing in the broader geophysical community (e.g. Uieda et al., 2013; Cockett et al., 2015; Rucker et al., 2015). These codes are transparent, flexible, and maintained by a cohort of users that contribute bug fixes and improvements as desired. SimPEG is open-source and offers significant benefits, including; 1) a mixed L_p norm inversion on both model values, and the gradient of their distribution, to better simulate geologic models with sharp or gradational boundaries according to the user's preference, 2) the option to use a non-zero starting model to incorporate a priori knowledge, 3) the ability to include a reference model in the objective function that may help reconcile models from complimentary disciplines, 4) the option to designate cells as active or inactive in the inversion. The modular framework also allows inversion of other geophysical datasets. SimPEG is python based, meaning it is free and platform independent, allowing the user to have a seamless work-flow within the python environment, from data processing to inversion, visualisation and interpretation.

Gravity surveys are well suited for rapid reconnaissance of shallow magmatic systems as the data are quick to acquire, and inversion models are fast to run. Using whole rock geochemistry data for input into MELTS, combined gravity and thermodynamic modelling provide insights into the present state of a mag-

48 matic system and the hazard it presents. Importantly, the combined analysis
 49 may offer significantly different interpretations than if the gravity data are con-
 50 sidered in isolation. We apply this method to image a shallow rhyolite magma
 51 reservoir beneath the Laguna del Maule volcanic field (LdMVF) and determine
 52 its key physical properties.

53 The LdMVF is located on the range crest of the southern Andes at 36°S,
 54 bordering Chile and Argentina and is geographically characterised by a lake
 55 enlarged around 19 ka, following the damming of its outlet by the Espejos rhy-
 56 olite lava flow (unit *rle*). The LdMVF comprises the largest concentration of
 57 high silica rhyolite in the Andes with at least 50 post-glacial eruptions since
 58 25 ka; including four between 3.3 and 2.1 ka (units *rsl*, *rcd*, *rln* and western
 59 *rcb*) (Andersen et al., 2016). These eruptions were from 24 vents and produced
 60 15 rhyodacite, and 24 rhyolite lava flows and domes (Singer et al., 2014). The
 61 position of the eruptive units and post-glacial vents is shown in Figure 1. Pre-
 62 vious large volume silicic eruptions include a dacite ignimbrite at 1.5 Ma (*igsp*),
 63 and a rhyodacite tuff associated with the Bobadilla caldera at 0.95 Ma (*igcb*)
 64 (Hildreth et al., 2010). The post-glacial decrease in mafic eruption frequency,
 65 and their spatially peripheral location, is inferred as evidence for a large, low
 66 density, high silica magma system acting as a density barrier for ascent of mafic
 67 magma.

68 Since 2007, the LdMVF has experienced uplift at rates over 20 cm/year (Le
 69 Mével et al., 2015), which Feigl et al. (2014) and Le Mével et al. (2016) interpret
 70 as resulting from inflation of a sill at around 5 km depth beneath the lake caused
 71 by the addition of new magma into the reservoir. By defining the location of
 72 the present day magmatic system and its physical properties, we provide the
 73 context to accurately interpret and understand the ongoing unrest and create a
 74 framework for interpreting monitoring data. From our model we estimate the
 75 melt vs crystal proportions and their distribution, the degree of pressurisation
 76 of the system, and investigate the implications of ratios of magma intruded to
 77 erupted. We consider the location of the magma system in relationship to the
 78 eruptive vents, the local tectonic framework and the ongoing deformation, and

79 throughout we consider the hazard implications of our model.

80 **2. Gravity data collection and reduction**

81 In 2014, 2015, and 2016, we collected 239 terrestrial gravity measurements
82 within a 10 km radius of the lake shore (Figure 1). We took measurements
83 every 500 m around the accessible shoreline, on islands, and in a series of radial
84 profiles extending 3 - 4 km from the lake. Thirty measurements were collected
85 farther from the lake to define the regional gravity field. For precise positioning
86 we used a Leica 530 GPS receiver with 15 minute occupation time per station;
87 with maximum baseline lengths of 10 km to the Observatorio Volcanologico de
88 Los Andes del Sur (OVDAS) CGPS network, we obtain height errors of better
89 than 15 cm (~ 0.045 mGal) at 95 % confidence level. We correct the daily raw
90 gravity data for Earth tide and drift using Gtools (Battaglia et al., 2012), to
91 produce gravity values relative to our local base. We convert our local gravity
92 values to absolute values using two nearby absolute gravity stations (S. Bonvalot
93 pers comm) to a measured accuracy of 0.016 mGal.

94 We correct the gravity data to the ellipsoid height datum as our heights
95 are derived from GNSS measurements referenced to the WGS84 ellipsoid and
96 follow the data reduction scheme outlined in Hinze et al. (2005). See supple-
97 mentary materials for details. The DEM used for terrain corrections does not
98 take into account the bathymetry of the lake, resulting in the terrain correc-
99 tions for stations near the lake being over estimated, due to low density water
100 being corrected as dense rock. To determine the gravity effect of the lake we
101 construct a forward model of the bathymetry using data acquired by Carrevedo
102 et al. (2015). We use an implementation of Talwani’s algorithm (Talwani and
103 Ewing, 1960), to compute the gravity effect of the lake water (density = 1000
104 kg/m³) at each gravity station. The lake correction at each station i , ($g_{i_{LC}}$),
105 which is subtracted from the terrain correction, is given by:

$$g_{i_{LC}} = \left(\frac{\rho_{TC}}{\rho_W} \right) g_{i_{lake}}$$

where ρ_{TC} is the terrain correction density, ρ_W is the water density and g_{lake} is the calculated gravity effect of the lake at each station. A maximum lake correction of 0.32 mGal is calculated at a station on one of the islands.

We use a combination of methods to determine the best Bouguer and terrain correction density with a summary presented in Figure 2 and details in supplementary material. Analysis of our gravity data and topography, using Nettleton (1939) and Parasnis (1966) methods, give densities of 2375 kg/m³ and 2361 ± 61 kg/m³, respectively. The same analysis on a nearby commercial dataset (Energy Development Corporation) gives values of 2450 and 2452 ± 24 kg/m³. Additionally, we convert the OVDAS 1D seismic velocity model to density using the relationship of Brocher (2005). The top 3 km of this model gives densities of 2429 kg/m³, increasing to 2448 kg/m³ at 7 km depth. The average of all the density determinations is 2413 ± 42 kg/m³ which we round to 2400 kg/m³ for use in the gravity reduction. We estimate an RMS error on the Bouguer anomaly of 0.11 mGal.

3. Gravity anomaly

A striking gravity low embedded within an east-west trending gradient is the dominant feature of the Bouguer anomaly map (Figure 3A). The heterogeneous data coverage, results in the gravity anomaly over the lake being interpolated between sparse data points. While this is unavoidable, we highlight that the true shape and amplitude of the anomaly may vary slightly from that presented. Areas shaded in grey are those areas of the main anomaly where the nearest gravity station is more than 1 km away. To isolate the anomaly, we remove a third order polynomial to approximate the regional field, resulting in a residual gravity anomaly (Figure 3B). The third order polynomial was chosen as the lowest order polynomial surface that suitably replicates the regional data. The regional field is oriented parallel to the South American / Nazca tectonic plate boundary and reflects the eastward dipping subducting plate. The residual gravity anomaly shows a -19 mGal, rectangular shaped gravity low, oriented

135 NNE and centred on the south shore of the lake, coincident with the centre of
 136 inflation modelled by Feigl et al. (2014). The gravity low is contained within
 137 the ring of post-glacial rhyolite vents that encircle the lake. Localised gravity
 138 highs up to 9 mGal, occur to the east of the lake.

139 Using Gauss’s theorem, where:

$$\Delta M = \frac{1}{2\pi G} \iint \Delta g(x, y) dx dy$$

140 we calculate the mass deficiency, ΔM , required to produce the main gravity
 141 anomaly, Δg , of 1.1×10^{13} kg, where G is the gravitational constant, contained
 142 within an area (x, y) of 95 km^2 . This mass is independent of the geometry of
 143 the body within which it is contained.

144 **4. Inversion and interpretation**

145 To determine the density distribution within the subsurface, we undertake a
 146 3D inversion using a Gauss-Newton, gradient based approach, implemented in
 147 the open source SimPEG framework (Cockett et al., 2015). SimPEG is partic-
 148 ularly well suited for volcano surveys where inaccessible topography results in
 149 irregular data coverage, as a regular grid of data is not required. We implement
 150 a novel L_p norm, ‘compact’ inversion scheme, with a summary described below
 151 and further details, including a link to the inversion code, in the supplementary
 152 materials.

153 We construct a 3D mesh with 250×250 m horizontal cell size, with vertical
 154 cell size increasing by a factor of 1.2 from 100 m in the top 500 m, to 350 m at
 155 the base of the grid at -5000 m a.s.l., resulting in 190,440 cells. The mesh covers
 156 a subset of the Bouguer gravity anomaly where the station density is highest,
 157 using 192 of the 239 stations (rectangular box in Figure 3B). Note that there
 158 is no interpolation of data in the inversion routine, only the measured gravity
 159 data are used.

160 The two main elements of the inverse problem are the data misfit, and the
 161 regularisation scheme. Data misfit is a metric which measures the difference be-

162 between observed and predicted data, while regularisation is a metric that is con-
 163 structed to evaluate the model’s agreement with assumptions and prior knowl-
 164 edge. We define the data misfit of the model $\phi_d(\mathbf{m})$ as:

$$\phi_d(\mathbf{m}) = \frac{1}{2} \| \mathbf{W}_d (F[m] - \mathbf{d}_{obs}) \|_2^2$$

165 Where $\mathbf{d}_{obs}$ is the observed data, $F[m]$ is a forward model that produces
 166 predicted data, and $\mathbf{W}_d$ is a diagonal matrix with elements equal to $\mathbf{W}_{d_{ii}} =$
 167 $1/\epsilon_i$ where ϵ_i is the standard deviation of the i th data point. We set ϵ to 0.1
 168 mGal for our model.

169 The second element is the regularisation function. Infinitely many models
 170 can fit the dataset, so we wish to find the model that has the desired characteris-
 171 tics, and is compatible with a priori information. A single model can be selected
 172 from the infinite set of models by measuring the length, or norm of each model.
 173 In our workflow, we run a two step inversion process, firstly inverting the data
 174 with a ‘smooth’ constraint imposed with a L_2 norm. Secondly, we compact the
 175 smooth model using a L_p norm, where $0 \leq p \leq 2$. L_2 norm models tend to create
 176 bodies with poorly defined, ‘smeared out’ boundaries (Sun and Li, 2014), how-
 177 ever our geologic conceptions of a magma reservoir as a finite body with defined
 178 edges (e.g. Bachmann and Bergantz, 2008; Cashman and Giordano, 2014, and
 179 examples therein), are better represented mathematically by a L_p norm that
 180 produces a more compact body with sharper edges.

181 To produce a compact model, we inspect the distribution of model values
 182 and gradients of the L_2 norm model, ϵ_p and ϵ_q , to find suitable values (See sup-
 183plementary material Figure 2). The choice of ϵ_p and ϵ_q is arbitrary, depending
 184 on the level of compactness required, and we choose the 95th percentile of the
 185 L_2 model distribution after visual inspection of models with a range of values.
 186 The effect of ϵ_p and ϵ_q is to flatten the model value and gradient histograms
 187 towards having more values equal to ϵ_p and ϵ_q , such that the model becomes
 188 less smooth, and more compact.

189 We solve the inverse problems using a scaled iterative re-weighted least

190 squares optimisation approach and define a composite objective function as:

$$\phi(\mathbf{m}) = \phi_d(\mathbf{m}) + \beta\phi_m(\mathbf{m})$$

191 where β is the tradeoff or Tikhonov parameter. A small β results in a model
 192 that fits the data very well, but may include excessive structure so that $\phi_m(\mathbf{m})$
 193 is large. Conversely if β is large, the optimisation results in a large $\phi_d(\mathbf{m})$. To
 194 select the optimal β , we use an L-curve approach and stop adjusting β when
 195 we have reached the target misfit. The optimisation is nonlinear so a gradient-
 196 based, Gauss-Newton method is used.

197 To run the inversion, we search for a perturbation of the model that reduces
 198 the objective function. The iterative optimisation process continues until the
 199 algorithm converges to a minimum and the misfit tolerance is achieved.

200 4.1. Model resolution

201 We investigate how the lack of gravity stations over the lake might affect the
 202 accuracy of the inversions by creating a checkerboard test, similar to that used in
 203 seismic tomography (Humphreys and Clayton, 1988). We create a density model
 204 consisting of square prisms (3.5 km horizontal sides) of alternating density, +300
 205 and -300 kg/m³. The width of the prisms is chosen to reflect the width of the
 206 gravity anomaly at half its amplitude and also corresponds to the east-west
 207 distance across the lake, through the center of the anomaly, where there are no
 208 gravity stations. The elevation of the top of the prisms is 1000 m a.s.l., and
 209 they are 1 km thick, except for the centre prism which is 2 km thick. The depth
 210 to the top of the prisms from topographic surface, ranges from ~ 1200 m to
 211 1500 m. In this model, we test the sensitivity of the dataset to the lateral, top
 212 and bottom boundaries of bodies of similar dimensions to the residual gravity
 213 anomaly as well as the resolved densities. We calculate the gravity effect of
 214 the checkerboard at each station, and use this as input data into the inversion
 215 routine.

216 The results of the checkerboard (Figure 4) show that the outlines of all the
 217 prisms are resolved even for prisms with few data points. Prisms bordering

218 the lake shore, where there are many stations closer to the top of the prisms,
 219 are better resolved, both laterally and at depth, compared to the outer prisms
 220 with fewer stations at greater depths. Prisms with fewer stations are generally
 221 recovered shallower, and with less density contrast than the true model. As
 222 such, we have confidence that for the size of the observed anomaly, the coverage
 223 of our stations should accurately determine the lateral extent of the anomalous
 224 body, especially in an east-west direction. Lateral boundaries across the lake
 225 in the north-south direction may be less well resolved and the density may be
 226 under estimated in areas of poor station coverage. The 500 m station spacing
 227 places an additional resolution limit on the smallest resolvable feature, in areas
 228 of good station coverage, of approximately 1 km.

229 *4.2. Inversion constraints from thermodynamic models*

230 The data collected in any geophysical survey are finite in number, while
 231 the physical property distribution in the Earth is continuous, resulting in an ill
 232 posed problem with no unique solution. To constrain the inversion, and reduce
 233 the number of possible models, we investigate the likely range of densities of
 234 magma found at LdMVF. As the observed gravity anomaly is an integral of all
 235 magmatic phases, (solids, liquids and exsolved gases), we must include the effect
 236 of H₂O and CO₂ and their behaviours in the system at different temperature and
 237 pressure conditions, when calculating magma system densities. While there is no
 238 open vent degassing or surface fumaroles at the LdMVF, we have measured CO₂
 239 soil gas concentrations of up to 5 % by volume in places around the lake shore,
 240 and in our modelling we test the sensitivity of the melt density to varying CO₂
 241 concentrations as described below. For the volumetrically dominant H₂O vapour
 242 phase we assume a closed system where that phase is retained. We use rhyolite
 243 MELTS for Excel (Gualda and Ghiorso, 2015) to calculate total magma system
 244 densities at equilibrium conditions from whole rock analyses representative of
 245 the most recently erupted rhyolites (*rsl*, *rcd*, *rln*), over a range of temperature
 246 and pressure conditions, representing various proportions of vapour, liquid and
 247 solid magma. We note that evolution and degassing of rhyolite magma may not

248 always strictly occur at equilibrium conditions (e.g. Gonnermann and Manga,
 249 2005) so that the melt may be more or less saturated in volatiles than our
 250 calculations suggest. As such, the calculated densities of the magma system at
 251 equilibrium should be considered as minimum values.

252 The rhyolites contain a narrowly varying range of SiO_2 , between 73.5 to
 253 75 wt % (Andersen et al., 2016), and we use an average composition for our
 254 analysis. We also compute densities from rhyodacites *rdcn* and *rdcd* (average
 255 69.4 wt % SiO_2), and andesites *asd* and *mcp* (63 and 53 wt % SiO_2 respectively)
 256 to test our model sensitivity to different compositions, as the magma reservoir
 257 is unlikely to be solely rhyolite (see Supplementary Figure 3). Common biotite,
 258 sparse amphibole and very rare pyroxene in the silicic units indicates high water
 259 content, so we used 5 wt % H_2O , derived from plagioclase-glass hygrometry
 260 (range 4.75-7.25 wt %, N. Andersen pers comm July 2016), and in the absence
 261 of melt inclusion information, fixed CO_2 at 100 ppm, typical of arc-rhyolites
 262 (Gonnermann and Manga, 2012). As a sensitivity test, we calculate rhyolite
 263 densities using water contents of 3, 4 and 5 wt % and found that a minimum
 264 of 4 wt % H_2O is required to reproduce the observed gravity (Supplementary
 265 Figure 4). The effect of increasing water content is to decrease the total density
 266 of the system for any given pressure or temperature. At the low pressures
 267 present at the depth of our gravity model, CO_2 is already exsolved so varying
 268 its concentration several fold makes only negligible density difference and we do
 269 not consider further the role of CO_2 .

270 Calculated rhyolite magma system densities (Figure 5) vary from 1300 to
 271 2100 kg/m^3 , over the pressure range of 50 to 120 MPa (~ 2 to 5 km depth
 272 for a crustal density of 2400 kg/m^3), and temperature range of 700 to 920 $^\circ\text{C}$
 273 (approximate solidus to above liquidus), giving a range of density contrasts from
 274 the gravity reduction density of -1100 to -300 kg/m^3 . Feigl et al. (2014) model
 275 a sill at ~ 5 km depth (120 MPa), which we use as a constraint for the upper
 276 end of the likely pressure range. The rhyodacite density varies from the rhyolite
 277 density by a maximum of only 80 kg/m^3 making the distinction between the two
 278 compositions not explicitly resolvable in our model. Andesite densities at 1020

279 °C geothermometer temperatures (Andersen et al., 2016) with 3 wt % H₂O at
280 90 MPa, are 2200 to 2300 kg/m³, a contrast of -200 to -100 kg/m³.

281 Figure 5A is annotated with isolines of the percentage rhyolite melt in the
282 system at each pressure modelled. The green shaded area represents the temper-
283 ature range of the Holocene rhyolite and rhyodacite magmas prior to eruption
284 (~790 to 854 °C), calculated from the Fe-Ti geothermometer and provide fur-
285 ther constraints on the range of probable densities.

286 The shape of the temperature vs density curves reflects the importance of
287 including the full thermodynamic, crystallisation and volatile phase effects in
288 the density calculations, not simply calculating the density of the liquid phase
289 (Figure 6). Below the liquidus, the system density decreases, reflecting the
290 start of feldspar crystallisation, exsolving H₂O from the melt. When quartz
291 and K feldspar start to crystallise, more H₂O is exsolved further lowering the
292 total density of the magma system. However, at LdMVF quartz + K feldspar
293 crystallisation is not observed in the rhyolite mineralogy, implying the temper-
294 ature of the system remains above this point (green box in Figure 6). The
295 high temperatures also limit the amount of exsolved H₂O to around 2.5 wt %.
296 The implication of these dynamics and the importance of modelling the vapour
297 phase, with respect to the interpretation of gravity data at active magmatic
298 systems, is explored further in the Discussion section 5.1.

299 To utilise these constraints, we run a suite of inversions where the lower
300 density contrast bound is increased from -1100 to -200 kg/m³ in steps of 100
301 kg/m³. The upper bound is fixed at 300 kg/m³ to account for the positive grav-
302 ity anomalies, assuming that the densities of intermediate and silicic volcanic
303 rocks are unlikely to be more than 2700 kg/m³.

304 4.3. Inversion results

305 First, we consider which of the suite of geophysically feasible models is most
306 viable from a geological perspective. The models that are valid over the greatest
307 ranges of pressure conditions are from -600 to -400 kg/m³ (Figure 5). For
308 example, the -600 kg/m³ model is valid from 80 to 100 MPa, or 3 to 4.2 km

309 depth, while the -400 kg/m^3 model is valid from 100 to 120 MPa, or 4.2 to 5.1
 310 km depth. The central body of the -500 and -600 kg/m^3 density contrast models
 311 have very similar volume ($\sim 30 \text{ km}^3$), indicating that at these values the model is
 312 less sensitive to changes in density and these are our preferred models. While the
 313 more extreme density contrast models, -900 to -1100 kg/m^3 , are geophysically
 314 feasible, they are unable to represent the bulk of the magma reservoir when
 315 thermodynamic data are considered as they are valid only for very low pressures
 316 and result in low melt proportion magmas ($< 50 \%$) inconsistent with geologic
 317 observations of crystal poor lava. In the lowest density contrast models (-200
 318 and -300 kg/m^3), a poorly defined low density body extends to the base of
 319 the model volume. There is no difference between the L_p norm and L_2 norm
 320 models, indicating that at these density contrasts, there is insufficient mass
 321 available to reproduce the gravity anomaly. These two models are not considered
 322 geologically reasonable, and are not discussed further.

323 The preferred inversion model of $+300$ to -600 kg/m^3 (Figure 7) converges
 324 with an RMS of 0.032 mGal, in 175 seconds on an Intel i7, 2.7 GHz processor
 325 with 16 Gb ram. The distribution of observed - calculated values are normally
 326 distributed around zero and show no obvious geographic bias (see Supplemen-
 327 tary Figure 5). The shallowest layers of this model show a ring of higher density
 328 material broadly coincident with vent locations to the west and south of the lake,
 329 while in the north, the Espejos vent, lava flow (*rle*) and pumice deposit is coin-
 330 cident with lower density material. The high - low - high density pattern across
 331 the north shore of the lake is present in all depth slices. The western boundary
 332 of this feature is broadly coincident with the inferred boundary of the Bobadilla
 333 caldera (Hildreth et al., 2010). The inferred caldera boundary is 1 - 2 km to the
 334 west of the gravity boundary, however its location is poorly constrained as it is
 335 buried by more recent volcanism.

336 Depth slices at 1500 and 1100 m a.s.l. show an irregular shaped low density
 337 region (contrast -200 kg/m^3 , outlined with a dashed blue line in Figure 7),
 338 covering most of the lake and shoreline, and extending to the north under the
 339 Espejos rhyolite lava flow (*rle*) and pumice deposits. This is similar to the

340 extent of the shallow conductor mapped with magnetotellurics by Cordell et al.
 341 (2015). At 0 m a.s.l. a low density body (contrast -600 kg/m^3), appears in the
 342 central portion of the lake. At -1100 m a.s.l. (approx 3 km depth), the main
 343 features of the model are a central low density body and a finger of lesser density
 344 contrast material to the north. The central low density body sits within a 1-2
 345 km wide ring of lower density contrast material, shown as the dashed red line
 346 in slice -2150 m , which is broadly coincident with the location of a sill modelled
 347 by Feigl et al. (2014) to explain present day deformation.

348 The main feature of our gravity model is a central low density body of
 349 uniform density throughout its thickness. Gravity is inherently poor at resolving
 350 horizontal layering that a magma system may exhibit, so our density model
 351 reflects the average bulk density of the system. The possible composition of this
 352 body is outlined in the Discussion, section 5.1. The volume of the central body
 353 at each model density contrast is shown in Figure 8 along with the proportion
 354 of melt possible at each density contrast and pressure range. Volumes of the
 355 anomalous body range from 6 to 50 km^3 for density contrasts of -1100 to $-$
 356 400 kg/m^3 , respectively. The overall effect of different density contrast models
 357 is to swell or shrink the central body, whilst retaining its same general shape
 358 (Supplementary Figure 6).

359 At -500 kg/m^3 , the base of the body is around -4000 m a.s.l. At -600 kg/m^3 ,
 360 the central body shrinks with a lower surface at around -3500 m a.s.l. , while
 361 at -700 kg/m^3 and higher contrasts, the base of the body stabilises at around
 362 -3000 m a.s.l. Thus, the thickness of the central body varies between 3 and 4
 363 km over a 200 kg/m^3 density range. A 3 dimensional representation is shown in
 364 Figure 9 for the -600 kg/m^3 density contrast model and shows a 30 km^3 body
 365 hosted within a 115 km^3 body of less density contrast.

366 4.4. Alternate scenario

367 We investigate an alternate scenario of a thick, low density surface layer to
 368 explain the residual Bouguer anomaly. Preliminary results from MT surveys
 369 (Cordell et al., 2015) indicate a low resistivity layer extending from near surface

370 to several hundred metres depth and interpreted as hydrothermal alteration
 371 or conductive lake sediments. This conductive layer covers much of the lake
 372 basin, with a footprint wider than the extent of the main Bouguer anomaly.
 373 Hydrothermal alteration often results in low density (e.g. Miller and Williams-
 374 Jones, 2016) from the chemical alteration of rock to low density clays. Recent
 375 field mapping, (J. Fierstein, pers comm 19th May 2016), identified a probable
 376 vent area under the lake, from an ignimbrite deposit mapped to the east, which
 377 is likely to have produced a thick accumulation of low density pyroclastic ma-
 378 terial within the lake basin. To test this scenario, we create a 500 m thick layer
 379 extending from the surface, with a density contrast of -300 kg/m³. The extent
 380 of the surface layer is the dashed blue outline in Figure 7, slice 0 and 1500 m
 381 a.s.l.. The cells inside the surface layer were fixed in the inversion, with the
 382 remaining model cells allowed to vary, so that the effect of the layer is included
 383 in the inversion but it is not adjusted. The ability to fix cells demonstrates the
 384 flexibility of SimPEG and allows us to easily explore alternate scenarios through
 385 a mix of forward and inverse modelling. The inversion result (Supplementary
 386 Figure 7) still requires considerable material beneath the surface layer to repli-
 387 cate the magnitude of the observed anomaly, and we conclude that the Bouguer
 388 anomaly cannot be solely caused by shallow low density material.

389 **5. Discussion**

390 We use our preferred density model of the LdMVF to consider questions
 391 related to the current state of the magma system, its relationship to eruptive
 392 vents and present day deformation and the potential for future eruptions.

393 *5.1. How much melt is present, and of what composition?*

394 The percentage of melt in a magma reservoir is a key determinant in that
 395 system's ability to erupt. By using a thermodynamic modelling approach, rather
 396 than simply calculating melt densities, we are able to extract proportions of
 397 melt, crystal and volatiles from our model. At crystal fractions greater than

50 to 60 %, a mechanical threshold is reached where crystals start to interlock, eventually forming a rigid sponge, inhibiting the flow of magma and reducing its ability to erupt (Marsh, 1981; Bachmann and Bergantz, 2008). However, the range of 50 - 60 % crystals is also where melt extraction from the crystal mush is most efficient, promoting the growth of eruptible, crystal poor caps to magma reservoirs (Bachmann and Bergantz, 2004; Charlier et al., 2005).

Figures 5 and 8 show that for a given density contrast, a range of scenarios is possible, suggesting a crystallinity zoned reservoir. For example at -500 kg/m³, sub-liquidus magma exists at 90, 100 and 110 MPa ($\sim$ 3.8 to 4.7 km depth). At 90 MPa, 85 % melt is possible, at 100 MPa, 60 % melt is possible, while at 110 MPa, scenarios of 4 or 35 % melt are possible, depending on the temperature.

Compositional gradients and the spatial distribution of crystals and melt within the reservoir are not explicitly resolvable within our model as gravity methods are relatively insensitive to horizontal layering. So, although we have assumed a uniform composition of the magma comprising the reservoir, the crystallinity gradient inferred is likely partially driven by variations in composition. Including more mafic material at depth will allow for higher crystallinities at higher temperature, without saturating in sanidine at the granite minimum (Andersen et al., 2016). These thermodynamic requirements are backed by field evidence of mafic and intermediate enclaves in post-glacial rhyodacite lavas (Hildreth et al., 2010) and by the model of Le Mével et al. (2016) for the injection of basalt into the base of the reservoir.

The calculated densities of andesite produce only a small density contrast (-100 to -200 kg/m³) with the background density (Supplementary Figure 3B). For the entire active reservoir to be made of this composition requires a volume 3 to 6 times greater than what is modelled to reproduce the observed gravity anomaly and such a volume is not feasible. The density contrast of andesite magma is close to that ascribed to the larger rim surrounding the active reservoir (red dashed line Figure 7), and it is possible that this rim is of intermediate composition and high crystal proportion ($\sim$ 70 % crystals at the modelled density contrast). A composite magma system of andesite and rhyolite magma

would have an overall density contrast of -400 kg/m^3 (-600 kg/m^3 rhyolite and -200 kg/m^3 andesite) which our gravity models show requires a volume of 50 km^3 . However, to reproduce the observed gravity (ignoring geometry), a 3:1 ratio of andesite to rhyolite is required to account for the lower density contrast of andesite compared to rhyolite magma. A 50 km^3 reservoir would require $\sim 38 \text{ km}^3$ andesite and 12 km^3 rhyolite. This simplistic analysis ignores the effects of geometry, where the andesite is likely at greater depth than the rhyolite, resulting in a lower contribution of andesite magma to the observed gravity. In order to quantify the proportions of different magma types, an independent estimate of the reservoir volume is required.

The rhyolite melt proportions are calculated using 5 wt % H_2O , which is at the lower end of the range of measured values (N. Andersen, pers comm July 2016). The effect of increasing water content is to increase the melt proportion for any given density contrast. For example, at 6 wt % H_2O at 90 MPa, ~ 95 % melt is calculated at a density contrast of -600 kg/m^3 , compared to ~ 60 % melt at 5 wt % H_2O . Therefore melt proportions may be considered minimums.

We can further constrain the likely range of melt proportion, by considering the temperature of the magma system. Using the Fe-Ti oxide geothermometer, Andersen et al. (2016) calculated pre-eruptive magmatic temperatures of between ~ 790 to 854°C for the Holocene rhyolites and rhyodacites. Models with density contrasts within this temperature range are shown by a star symbol in Figure 8. This temperature range intersects our preferred -500 and -600 kg/m^3 density contrast models at melt fractions of 50 to 85 % (shown by larger star symbols). As the temperatures are derived from erupted lavas, the 85 % melt fraction is likely to represent the shallowest portion of the magma reservoir and is consistent with field observations that the rhyolitic lavas and tephras are crystal-poor (Singer et al., 2014; Hildreth et al., 2010).

A credible configuration of the present day LdMVF magma system from density contrast, temperature and melt percentage data is a crystallinity zoned magma reservoir with a shallow, crystal poor rhyolite magma (>85 % melt), e.g. ‘holding zone’ of Charlier et al. (2005), overlying progressively more crystal rich

460 magma (<50 % melt). This reservoir is surrounded by a cumulate mush e.g.
 461 ‘root zone’ of Charlier et al. (2005), of intermediate composition and densities
 462 close to the host rock. A density zoned magma reservoir is consistent with a
 463 hypothesis that each of the LdMVF eruptions tapped only the top of the magma
 464 system. As melt proportion falls below 50 %, it becomes more difficult to erupt,
 465 favouring the eruption of only the shallowest, crystal poor part of the system.
 466 The 50 % melt fraction is also within the zone of most efficient extraction of melt
 467 from crystal mush, so the magma reservoir may be in a suitable configuration to
 468 repeatedly generate a cap of low crystal content, eruptible rhyolite overlying a
 469 crystal rich mush. Importantly for hazard considerations, Castro and Dingwell
 470 (2009), from the 2008 Chaiten eruption, and Castro et al. (2013) and, Jay et al.
 471 (2014), from the 2011-12 Cordon Caulle eruption, showed that, near liquidus,
 472 volatile rich rhyolite magma, such as is present at LdMVF, is very fluid and
 473 can migrate rapidly from source to eruption site, resulting in very little warning
 474 prior to eruption.

475 5.2. *Is the LdMVF magma system over-pressured?*

476 We investigate whether the large density contrast of the imaged shallow
 477 magma reservoir with the surrounding crust is enough to generate magma over-
 478 pressure that exceeds lithostatic pressure. Gregg et al. (2015) provide a simple
 479 1-D formulation for calculating the over pressure from a density contrast, taking
 480 into account the restoring force imposed by the lithostatic load (P_{litho}) on the
 481 magma pressure head (ΔP_{buoy}). The overpressure OP is defined as:

$$OP = \Delta P_{buoy} + P_{litho}$$

482 OR

$$OP = \Delta \rho g h_{ch} - \rho_r g z$$

483 where $\Delta \rho$ is the density contrast between magma, ρ_m , and host rock, ρ_r , g is
 484 acceleration due to gravity (9.8 m/s²), h_{ch} is the height of the magma reservoir
 485 and z is the depth to top of the reservoir.

486 The assumptions of this formulation are of a spherical magma reservoir at
 487 a depth greater than the reservoir radius. The reservoir imaged at LdMVF
 488 approximates these conditions, as the geometry aspect ratio is $\sim 2:1$ and radius
 489 $\approx$ depth.

490 For the ~ 3 km thick magma reservoir imaged at LdMVF, at 2 km depth,
 491 with a maximum magma/host rock density contrast of -1100 kg/m^3 , $\Delta P_{buoy} =$
 492 32.3 MPa and $P_{litho} = 47.0 \text{ MPa}$, resulting in an overpressure of -14.7 MPa .
 493 The lithostatic load is around 50 % more than the magma pressure head and
 494 no resultant buoyancy force exists, and is likely to require external triggering
 495 to generate an eruption (e.g. Allan et al., 2012). The preferred -500 to -600
 496 kg/m^3 density contrast models, more representative of the whole reservoir, result
 497 in even lower head pressures. However, if the system is decompressed, the
 498 significant volatile phase is likely to result in a hazardous explosive eruption.

499 We note that if we only consider the melt phase density (refer to Figure 6),
 500 then our estimates of buoyancy would be substantially incorrect as the melt
 501 only phase density is much closer to the host rock density, resulting in a much
 502 reduced buoyancy force value. Additionally, melt only density models will not
 503 consider the exsolution of a free vapour phase, which may lead to an incorrect
 504 assessment of the hazard potential and likelihood of an explosive gas driven
 505 eruption.

506 *5.3. Magma reservoir volume vs erupted volume*

507 The volume of magma erupted is often used as a proxy for the volume of
 508 magma intruded into a system, to gain insights into the overall size of the mag-
 509 matic system and what proportion of the total system is eruptible at any given
 510 time. There are relatively few studies where both the intruded and extruded
 511 volumes are accurately known (White et al., 2006), usually because the intruded
 512 volumes are poorly defined either by lack of outcrop or from low resolution geo-
 513 physical imaging. There are two parts of the LdMVF magmatic system whose
 514 volumes we can compare to erupted volumes. A broad cumulate mush (115
 515 km^3) within which a smaller volume (30 km^3) of active magma exists, from

516 which an even smaller volume ($<2 \text{ km}^3$) of magma is periodically erupted.

517 The volume of post-glacial lavas at LdMVF is estimated at about 7 km^3
518 with $30\text{-}40 \text{ km}^3$ of pyroclastic material (Hildreth et al., 2010, J. Fierstein, pers
519 comm 19th May 2016), for a total minimum volume of at least 40 km^3 . Com-
520 paring our modelled magma reservoir volume (including the cumulate mush
521 margin, 115 km^3) to the total post-glacial erupted volume, we calculate an in-
522 truded:extruded (I:E) ratio of 2.9:1, similar to that found at large silicic systems
523 such as Yellowstone (3:1) (White et al., 2006).

524 When comparing the ratios of only the *rl*n eruptives (the largest of the 3
525 most recent rhyolites, $\sim 2 \text{ km}^3$ rhyolite lava and tephra), to our active magma
526 reservoir volume (excluding the margin), using our preferred models with density
527 contrasts of -600 and -500 kg/m^3 , we find an I:E ratio of 15:1. If the volume
528 of the present day magma system has remained approximately constant since
529 the last eruption at $\sim 2\text{-}3 \text{ ka}$ (ongoing intrusions account for relatively minor
530 volumetric change), then a 15:1 I:E ratio suggests that the *rl*n eruption tapped
531 $\sim 7 \%$ of the reservoir.

532 Andersen et al. (2016) propose that chemical and temporal trends of erupted
533 LdMVF rhyolites implies the extraction of chemically distinct melts from a
534 long-lived, compositionally evolving upper crust reservoir. Our I:E ratios are
535 consistent with such a reservoir, and as we discuss in section 5.4 the location of
536 this eruptible magma may be required to migrate within the larger system to
537 feed the mapped eruption vents.

538 The inflating sill modelled by Feigl et al. (2014), and the injection model
539 of Le Mével et al. (2016) proposed to explain the current deformation event
540 likely represent the intrusions required to sustain the reservoir, and imply that
541 episodes similar to the present day intrusion and associated deformation have
542 probably occurred many times. Indeed, the $>60 \text{ m}$ of uplift and tilting of
543 the high-stand paleoshoreline since 9.5 ka , suggests the current intrusion is not
544 unique in the history of the LdMVF.

545 The LdMVF scenario, where small magma volumes are erupted from a long
546 lived system, is similar to the small volume ($<1 \text{ km}^3$) post caldera forming

eruptions at Long Valley. Reid et al. (1997) from analysis of Zr crystals from the 115 ka Deer Mountain rhyolite, found that frequent injections of mafic magma were required to maintain the magma system at high temperatures for long periods of time. A consequence of this is that the Long Valley magma reservoir remained molten for longer, resulting in the accumulation of a much larger volume of magma than that erupted.

In contrast, Taupo volcano is characterised by frequent eruption of compositionally distinct magma batches, held in ephemeral chambers little bigger than the erupted volume (Sutton et al., 2000), implying I:E ratios closer to unity. As an example, Wilson (1993) proposed that the eruption of unit S from Taupo evacuated the majority of the chamber, and that this was representative of the 28 other eruptions from Taupo since the 26.5 ka, 530 km³ dense rock equivalent, Oruanui eruption. Additionally, Sutton et al. (2000) categorised Taupo eruptions into three subgroups and with no overlap in composition or time, suggesting the complete emptying of the reservoir between subgroups. The ability of Taupo eruption to effectively drain a large proportion of the chamber, and rapidly regenerate it, is proposed by Wilson and Charlier (2009) to be related to the active rifting and high heat flow environment of the Taupo Volcanic Zone, that allows magma otherwise bound within the crystal mush to be mobilised and erupted. We discuss the relationship between tectonics and the LdMVF magma system further in section 5.4.

5.4. Relationship to eruption vents, deformation and tectonics

To address the question of Singer et al. (2014) as to whether a crystal poor rhyolitic magma reservoir underlies the entire volcanic field, our model suggests that a body of crystal poor magma currently only exists under the southern portion of the LdMVF. To test the hypothesis that ring faulting at the reservoir margins has produced the surface pattern of vents encircling the lake (e.g. Singer et al., 2014, Figure 6), we compare our imaged magma reservoir to analytic and numerical models of volcanic ring fault growth. Yokoyama (2015) developed an analytic, shear fracture model, for the formation of parasitic vents due to a

dilatation source representing a magma chamber, and found that vents forming along a ring fracture occur at a distance from the source of $r = \pm 0.82D$, where D is the depth to the source. For a magma reservoir at 3 km depth, this would indicate a maximum vent distance of around 2.5 km from the edge of the magma reservoir. Using numerical models, Gudmundsson et al. (1997) found that the maximum tensile stress at the free surface, was a distance of 1 to 2 times the chamber radius, from the centre of the chamber, depending on chamber geometry. The Nieblas (rln) vent is approximately 2.5 km from the southern edge of the imaged magma reservoir, or 2 times the reservoir radius from the centre, and was conceivably fed from the present magma system location. However, the remainder of the post-glacial rhyolite vents are farther away, suggesting the size of the present day reservoir is smaller, or in a different location to that which fed the other eruptions. Migration of the active, eruptible, portion of the larger reservoir system to the edges of the crystalline rim would allow for eruption of more distal vents via a ring fault model. Cole et al. (2010) also found magma migration is required to explain subsidence and eruptive trends within the Okataina Caldera and similar magma migration to distant vents has been suggested for Katmai-Novarupta (Hildreth, 1991) and at Rotorua-Kapenga calderas (Nairn, 2002). The higher density material associated with vents in the north west may represent solidified rhyolite and andesite dykes that fed these vents.

The spatial and depth extent of the magma system imaged at LdMVF in this study, correlates well with the extent of a 5 km deep sill modelled by Feigl et al. (2014) as the cause of the ongoing deformation (Figure 9). The base of our low density body lies at approximately the sill depth, so our model is consistent with a high silica magma reservoir being underplated by an injection of mafic magma, also required by thermodynamics as discussed previously. Our gravity model is not able to resolve the sill, as it is too deep (5 km) and too thin (10 m), and we calculate a gravity effect of ~ 0.03 mGal. We observe that the low aspect ratio (width/height) of the magma reservoir diverges from the typical high aspect ratio commonly portrayed in theoretical models (e.g.

Gregg et al., 2015) and suggests that a wider range of orientations should be considered in these models, especially for shallow systems. We propose that shallow reservoirs are influenced more by the regional horizontal stress regime, than by the depth of burial, resulting in a more vertically oriented system (e.g. Saxby et al., 2016). Indeed, the thin overburden, with frequent injection of heat and volatiles into the base of the system, likely leads to over-pressures being more regularly exceeded. This may help explain differences in eruption trends between LdMVF and deeper sourced systems like Taupo, where the magma storage zone is at a minimum 6-8 km deep (Charlier et al., 2005).

The LdMVF magma system is located adjacent to the Troncoso fault, a 20 - 30 km long linear structure visible from satellite imagery. Although our gravity data are unable to map any displacement across the fault, focal mechanisms from an 8 km deep Mw 6.0 earthquake in June 2012 <http://earthquake.usgs.gov/earthquakes/eventpage/usp000jmf2#executive>, 2 km west of the fault, suggest right lateral motion on a steeply eastward dipping fault plane. If the Troncoso fault was extended north-east into the lake, as unpublished magnetic data collected by the authors suggest, and connected to faults mapped on the north shore of the lake, then it would bound the western edge of the magma reservoir in a releasing bend. Extensional or strike-slip environments are common features of silicic systems elsewhere (e.g. Taupo, Okataina and Long Valley) and appear to play important roles in the formation of large magma reservoirs. Saxby et al. (2016) proposed that the accumulation and ascent of magma at the Ilopango caldera, El Salvador, was similarly controlled by local and regional tectonic stress and that caldera collapse occurred during pull apart basin formation along the main strike slip tectonics of the region. However at the LdMVF, caldera collapse has not occurred since the 950 ka Bobadilla Caldera formation, so it is likely that tectonics, while important for controlling the shape and location of the magma reservoir, is not the only factor determining if caldera collapse occurs.

In addition to placing controls on the location and shape of magmatic reservoirs, faults may place important roles in the localisation of future eruptions.

Strongly magnetic lineaments mapped in beneath the lake, are interpreted as dykes intruding along faults parallel to the dominant tectonic strike. Fault and magma reservoir interaction have been important in the control of other Andean silicic eruptions (e.g. Chaitén; Wicks et al., 2011), and should be considered as part of hazard assessment and monitoring at LdMVF. We suggest that further work is undertaken on the Troncoso and nearby faults, to characterise local and regional stress fields, and assess their influence on the magma reservoir location, geometry and possible future eruption sites.

6. Conclusions

A -19 mGal gravity anomaly at the Laguna del Maule volcanic field (LdMVF) can be explained by a shallow, 30 km³ magmatic reservoir, that contains at least 85 % melt in the shallowest part, reducing to <50 % at the base of the system. This active reservoir sits within a 115 km³ partially to wholly crystallised (>70 % crystal) cumulate mush zone. Though we cannot image it, thermodynamic considerations require a compositional gradient of more mafic material at depth, to maintain inferred crystallinity and temperature conditions. Our data cannot be solely explained by a low density surface layer, although some low density surficial material is modelled, likely representing a shallow hydrothermal system. Nor can the data be explained by a density model which ignores the effects of the free volatile phase on lowering the total magma system density.

Thermodynamic modelling provides additional insights into the magma system over approaches that consider only the melt phase density. Crucially, we have determined the distribution of melt and crystal in the magma system and included the free vapour phase that may drive future explosive eruptions. This approach allows for a more comprehensive hazard determination than density models based only on the liquid phase. Additionally our modelling code offers a new open source approach to geophysical inversion, highly applicable to volcano studies.

The present day crystal poor reservoir does not underlie the extent of the

668 post-glacial LdMVF vents, and stress models of dyke propagation from magma
 669 reservoirs require it to have moved over time to be in closer spatial proximity
 670 to older vents. Alternately, the present day magma system is smaller than that
 671 which formed the early post-glacial eruptions, or lateral migration of magma
 672 from older eruption vents to the present location has occurred. Intruded to
 673 extruded volume ratios suggest that the Las Nieblas (*rln*) eruption tapped $\sim 7\%$
 674 of the magma reservoir volume and that the post-glacial LdMVF is characterised
 675 by eruptions of small aliquots of magma from the top of a larger reservoir, rather
 676 than the evacuation and catastrophic collapse of the entire reservoir.

677 The reservoir lies directly above the modelled source of the ongoing defor-
 678 mation and supports the model of Le Mével et al. (2016) that the deformation
 679 is caused by injection of mafic magma into the base of a reservoir. Even though
 680 a large density contrast with the crust exists, we calculate that the magma
 681 reservoir is not over pressured, and that eruption due to buoyancy effects alone
 682 is unlikely. The western margin of the magma reservoir is in the hanging wall
 683 of the hypothesized Troncoso fault and we suggest that the location and ori-
 684 entation of the reservoir may be influenced by local tectonics which may also
 685 provide the trigger required for the next eruption.

686 The presence of a shallow, large volume of volatile rich magma stored at near
 687 liquidus conditions, adjacent to a regional scale fault has important implications
 688 for the onset of any future eruptive activity. As such, our 3D density model of
 689 the LdMVF magma reservoir provides essential context to interpret monitoring
 690 data, and contributes to greater understanding of the ongoing dynamics of this
 691 unique area of volcanism. Finally, our combined gravity and thermodynamic
 692 analysis, provides a framework for richer interpretation of gravity data from
 693 silicic magma systems worldwide.

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Figure 1: Simplified geology map of the central basin of the LdMVF (after Hildreth et al., 2010). Gravity station locations are shown as black dots. The red arrow indicates the centre of inflation from Feigl et al. (2014). Ages are from $^{40}\text{Ar}/^{39}\text{Ar}$ ratios (Andersen et al., 2016) and red stars show post-glacial vents. The dashed black box shows the outline of the modelled area. The smaller location map shows Laguna del Maule volcanic field as a red ellipse, with other Holocene volcanoes as yellow triangles.

Figure 2: Bulk density (blue line) derived from 1D Observatorio Volcanologico de Los Andes del Sur (OVDAS) seismic velocity model (Vp plot, right) using the relationships of Brocher (2005). The density plot shows the results of the Nettleton and Parasnis density determinations from our gravity data in black, whilst the equivalent density determinations from a neighbouring commercial gravity dataset are shown in green. The correction density chosen for this study (2400 kg/m^3) is in red.

Figure 3: A) Bouguer gravity anomaly. B) Residual gravity anomaly after removal of a third order polynomial surface from the Bouguer data. The black box in B is the area modelled. Red stars are post-glacial eruption vents. The shorelines of Laguna del Maule and Laguna Fea are shown in white. Areas of the main anomaly shaded in grey are poorly constrained by the gravity observations and rely on interpolated data.

Figure 4: Elevation and depth slices from the checkerboard test at 1125 m a.s.l. (top) and -125 m a.s.l. (lower). The checkerboard grid is shown in the 1125 m a.s.l. slice and also in the cross sections. The outline of the lake is shown in grey, while gravity stations are shown as dots and triangles (in cross sections).

Figure 5: A) Plot of magma temperature vs density for a representative rhyolite magma, calculated using MELTS. The plot shows the results from 50 to 120 MPa at 5 wt % H₂O. Annotations include the proportion of liquid melt (dashed lines), gravity model volumes at each density contrast (solid black lines). Densities are the total magma system density, including liquid, crystal and free volatile phases. The green shaded area shows the magmatic temperature range from Fe-Ti oxide geothermometers (Andersen et al., 2016). B) Plot of magma density (contour lines in kg/m³) and melt percent (colour gradient), as a function of pressure and temperature. The green rectangle shows the magmatic temperature range from Fe-Ti oxides, as in A.

Figure 6: Phase diagram showing the evolution of H_2O in the system as a function of temperature (dark blue and grey lines). The yellow and red lines show the corresponding density evolution of the total system (yellow) and liquid phase (red) and shows the large density difference caused by the exsolution of H_2O . The green shaded box highlights the magma temperatures derived by Fe-Ti geothermometry (Andersen et al., 2016). The vertical black lines show the temperatures at which feldspar and then quartz and K feldspar appear. The quartz, K feldspar crystallisation point is below the observed temperature range and explains the lack of quartz and K feldspar in LdMVF rhyolites.

Figure 7: Elevation slices and cross sections from the inversion model with density contrast $+300$ to -600 kg/m^3 . The outline of the lake is shown in grey, while gravity stations are shown as grey dots and triangles (in cross sections). Mapped faults are shown as black lines and post-glacial eruption vents are shown as red stars in the 1500 m and 0 m a.s.l. slices. The dashed blue line in slices 1500 m and 1100 m a.s.l. represents a low density surface layer similar to the conductive surface layer imaged by Cordell et al. (2015). The dashed red line in the -2150 m a.s.l. slice shows the lower density contrast rim of the main low density body, while the green outlined rectangle indicates the projection of the sill modelled by Feigl et al. (2014) at 5 km depth (-3000 m a.s.l.).

Figure 8: Summary of model volumes and proportion of melt possible for each model. The four boxes show the proportion of melt for a pair of pressure ranges, with the lower pressure model shown as blue dots and the higher pressure model as red dots. Star symbols indicate the models that fall within the temperature range of pre-eruptive temperatures determined by Fe-Ti oxide geothermometry by Andersen et al. (2016), (see Figure 5). The volume of the gravity model for each density contrast is shown along the base of the plot. Note that at some density contrasts, a range of pressures at which various proportions of melt exist are possible. Our preferred suite of models is highlighted in green shading and the four larger stars show the models that are most likely when all constraints are considered.

Figure 9: 3D view of the preferred $-600/+300 \text{ kg/m}^3$ density contrast model, illustrating the LdMVF magma system. The isosurface (volume = 30 km^3) shown in purple is -600 kg/m^3 , and sits within a 115 km^3 body of -100 to -200 kg/m^3 density contrast (dashed red line) interpreted to represent a partially ($>70\%$ crystal) to wholly crystallised mush surrounding an active magma reservoir that contains 50 to $>85\%$ melt. The orange plane is the sill modelled by Feigl et al. (2014) to explain the current deformation episode and the grey shaded vertical plane is the Troncoso fault. The outline of the lake is shown superimposed on the topography.

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