Volcanic Architecture and Unrest Processes: Insights from Static and Time-varying Potential Field Models

by

Craig Miller

M.Sc., University of Auckland, 1996
B.Sc., University of Auckland, 1994

Thesis Submitted in Partial Fulfillment of the Requirements for the Degree of Doctor of Philosophy in the Department of Earth Sciences Faculty of Science

© Craig Miller 2017
SIMON FRASER UNIVERSITY
Fall 2017

Copyright in this work rests with the author. Please ensure that any reproduction or re-use is done in accordance with the relevant national copyright legislation.
Approval

Name: Craig Miller

Degree: Doctor of Philosophy (Earth Sciences)

Title: Volcanic Architecture and Unrest Processes: Insights from Static and Time-varying Potential Field Models

Examinining Committee: Chair: Doug Stead
Professor

Glyn Williams-Jones
Senior Supervisor
Professor

Gwenn Flowers
Supervisor
Professor

Jeff Witter
Supervisor
Innovate Geothermal Ltd.
Adjunct, Earth Sciences

Diana Allen
Internal Examiner
Earth Sciences

Shanaka de Silva
External Examiner
Professor
Earth, Ocean and Atmospheric Sciences
Oregon State University

Date Defended: 27 November 2017
Abstract

Knowledge of a volcano’s architecture or internal anatomy, provides critical context for correctly interpreting signals of volcanic unrest. In this thesis, I use measurements and models of Earth’s gravity and magnetic fields, applied at two contrasting volcanoes, Laguna del Maule volcanic field, Chile, and Mt Tongariro, New Zealand, to model their architecture and time-varying processes occurring within. I reveal relationships between magma bodies, hydrothermal systems, basement and fault structures, and provide a quantitative basis for improving understanding of the causes of volcanic unrest indicators. Gravity inversion, constrained by thermodynamic modelling at Laguna del Maule volcanic field, images a shallow, volatile rich, silicic magma body, above a previously modelled inflating sill, bound to the west by the regional scale Troncoso fault. Magnetic models show NE trending, remanently magnetised features, parallel to the Troncoso fault, interpreted as dykes intruding along faults. Further evidence of magma and fault interaction, from time-varying gravity changes, shows the inflating sill produces stress changes on the Troncoso fault, allowing shallow hydrothermal system fluids to migrate into it, resulting in mass addition and positive gravity changes through time. Fluid flow into the fault zone may be further modulated by shaking from nearby earthquake swarms. At Mt Tongariro, geologically-constrained gravity and magnetic models map large faults cutting the basement beneath the volcano, and delineate an extensive hydrothermal system. The hydrothermal system is bound laterally by the basement faults, while the basement itself acts as a low permeability barrier. The 2012 eruption at Upper Te Maari crater depressurised the hydrothermal system, promoting subsidence from the evacuation of pore space. Time-varying gravity models show shallow mass addition above the subsidence source, derived from a combination of pore fluid migration, condensation, cooling, and meteoric input, indicating the system is still repressurising. I show that the illumination of volcano architecture provides a rich, quantitative environment, to better interpret volcanic unrest. I combine traditional potential field geophysical methods and ground deformation data, with state of the art modelling techniques, and create a powerful and effective toolbox for the 21st century volcanologist.

Keywords: geophysics; gravity; hydrothermal system; inverse modelling; magma; volcano
Dedication

For my teachers, I still have so much to learn.
Acknowledgements

I would like to thank Dr Glyn Williams-Jones for offering the Laguna del Maule project, his supervision and support of this thesis, and for providing the funding and resources required. Thanks to co-supervisors Dr Jeff Witter and Dr Gwenn Flowers for their support of this project. My sincere thanks to my unofficial supervisor, Dr Gilda Currenti, for teaching me many of the analytic and finite element modelling and inversion codes used here. Special thanks to Dr Brad Singer, who conceived the Laguna del Maule project, as well as to Dr Basil Tikoff for scientific and logistical support of the gravity meters. Thanks to all the Laguna del Maule team, Dr Hélène Le Mével, Dr Nathan Andersen, Dana Peterson, Crystal Wespestad, Dr Cliff Thurber, Dr Kurt Feigl, Neal Lord, Peter Sobol, Dr Martyn Unsworth, Darcy Cordell and Dr Katie Keranen. It has been a fantastic experience working with a large diverse team on such an interesting topic.

Thank you to the many funding agencies that made this work possible. In New Zealand, thanks to EQC for the PhD scholarship, and to GNS Science Core Funding for supporting my time and field costs. In Canada, thanks to Mitacs Accelerate PhD scholarship, and to Mira Geoscience for its contribution to this. I am thankful for support in my final year through Graduate Fellowships from the Department of Earth Sciences at Simon Fraser University. National Science Foundation Integrated Earth Systems grant EAR-1411779 and EAR-1322595 provided funding for the Chile fieldwork.

Geophysical studies of volcanoes are a not a solitary pursuit. Data collected in this thesis and the results and interpretations derived from them could not have occurred without the numerous field assistants that spared their time and energy to help. In New Zealand I would particularly like to thank Nellie Olsen for her tireless and repeated assistance, often at short notice. Special thanks to Natalia Deligne for field assistance, and for being a lasting source of scientific and personal style inspiration. Thank you to Matt Stott, Mel Climo and family for last minute dinners and friendship. In Chile I would like to thank the many Chilean students from University of Chile and University of Concepcion for their enthusiastic help, both in the field and with negotiating aduanaas, carabineros and other officedom. Thanks to Dr Daniel Diaz and Dr Andreas Tassara from these institutions for their assistance. Muchas gracias to Don Luis Torres, Alcalde de Mar of Laguna del Maule for outstanding hospitality.
Thanks to the many OVDAS staff, especially Maria Loreto Cordova, Carlos Cardona, Dr Alvaro Amigo and Rayen Gho for their field support, data requests and local knowledge. You do a fantastic job and I hope the results of this thesis are of benefit to Sernageomin and Chile.

Thanks to my many GNS colleagues, especially to Dr Gill Jolly for letting me take the time off to undertake this study. It was not an easy decision for me to undertake this project as a mature student, but it is a better form of mid-life crisis than many. Thanks to Dr Ian Hamling and Dr Sigrun Hreinsdóttir for assistance with the Tongariro deformation data, on top of your normal workload.

Thank you to all the staff at Mira Geoscience, Vancouver, especially Thomas Campagne for guiding me through the intricacies of GoCAD and VPMG. A big thanks to the SimPEG team at UBC, especially Dominique Fournier for generously helping get me up to speed with the gravity and magnetic inversion codes. Thank you to my fellow SFU colleagues, the SFU Earth Science department support and technical staff, and my lab mates, especially Dr Jeff Zurek, for keeping it real. Finally, thank you to my parents for their quiet support over the years.

I hope those who read this thesis enjoy it as much as I have enjoyed creating it.
Table of Contents

Approval ii
Abstract iii
Dedication iv
Acknowledgements v
Table of Contents vii
List of Tables xii
List of Figures xiv

1 Introduction 1
1.1 Motivation for this Thesis ........................................ 1
1.2 Field Sites ....................................................... 3
  1.2.1 Laguna del Maule ........................................ 3
  1.2.2 Mt Tongariro .............................................. 5
1.3 Overarching Research Theme and Objectives .................... 7
1.4 Methods .......................................................... 8
  1.4.1 Gravity ..................................................... 8
  1.4.2 Magnetics .................................................. 10
  1.4.3 Software and modelling codes .............................. 11
1.5 Thesis Roadmap .................................................. 12
1.6 References ...................................................... 13

Preamble 21

2 3D Gravity Inversion and Thermodynamic Modelling Reveal Properties
  of Shallow Silicic Magma Reservoir Beneath Laguna del Maule, Chile 22
  2.1 Introduction .................................................... 23
  2.2 Gravity Data Collection and Reduction ........................ 26
  2.3 Gravity Anomaly ............................................... 28
2.4 Inversion and Interpretation ........................................... 28
  2.4.1 Model resolution .................................................. 31
  2.4.2 Inversion constraints from thermodynamic models .......... 31
  2.4.3 Inversion results ................................................ 35
  2.4.4 Alternate scenario .............................................. 38

2.5 Discussion .............................................................. 40
  2.5.1 How much melt is present, and of what composition? ....... 40
  2.5.2 Is the LdMVF magma system over-pressured? ............... 42
  2.5.3 Magma reservoir volume vs erupted volume ................. 43
  2.5.4 Relationship to eruption vents, deformation and tectonics .. 45

2.6 Conclusions ............................................................ 47

2.7 Acknowledgements .................................................... 48

2.8 References ............................................................. 49

Preamble 55

3 Microgravity Changes at the Laguna del Maule Volcanic Field: Magma
Induced Stress Changes Facilitate Mass Addition 56
  3.1 Introduction .......................................................... 57
  3.2 Gravity Measurements and Processing ............................ 59
  3.2.1 Free air gradient and uplift correction .................... 60
  3.2.2 Lake level and water table variations ..................... 61
  3.3 Residual Gravity Change Results .................................. 61
  3.3.1 2013 to 2014 .................................................. 63
  3.3.2 2014 to 2015 .................................................. 63
  3.3.3 2015 to 2016 .................................................. 64
  3.4 Residual Gravity Change Modelling ............................... 64
  3.4.1 Inversion method .............................................. 65
  3.4.2 Source models ............................................... 65
  3.4.3 2013 to 2014 source model results ......................... 68
  3.4.4 2014 to 2015 source model results ......................... 69
  3.4.5 2015 to 2016 source model results ......................... 70
  3.5 Discussion .......................................................... 72
  3.5.1 Nature of intruding fluid and mass balance ............... 72
  3.5.2 Stress change mechanism for mass emplacement .......... 73
  3.5.3 Fault zone hydraulic conductivity and permeability estimates . 78
  3.6 Conclusions ........................................................ 80
  3.7 Acknowledgments ................................................... 80
  3.8 References ........................................................ 81
## 4 Internal Structure and Volcanic Hazard Potential of Mt Tongariro, New Zealand, from 3D Gravity and Magnetic Models

4.1 Introduction .............................................. 88
4.2 Geologic Setting and Existing Geophysical Data ......................... 89
  4.2.1 Previous geophysical studies ................................ 91
  4.2.2 Physical property measurements ............................. 91
4.3 Geophysical Data Acquisition and Processing ............................ 92
  4.3.1 Gravity survey design ................................... 92
  4.3.2 Gravity datasets ........................................ 94
  4.3.3 Gravity data reduction and errors .......................... 94
  4.3.4 Complete Bouguer anomaly ................................. 95
  4.3.5 Aeromagnetic data acquisition and processing ............... 97
  4.3.6 Residual total magnetic intensity anomaly ................... 98
  4.3.7 Spectral analysis ........................................ 99
4.4 Geological Modelling and Geophysical Inversion .......................... 101
  4.4.1 Model initialisation ...................................... 102
4.5 Results .................................................. 103
  4.5.1 3D geologically constrained density inversion ............... 103
  4.5.2 3D geologically constrained susceptibility inversion .......... 105
  4.5.3 Limitations and sensitivity of geophysical models .......... 107
4.6 Discussion ............................................... 107
  4.6.1 Basement structure ...................................... 107
  4.6.2 Volcanic edifice structure ................................ 109
  4.6.3 Hydrothermal system .................................... 111
  4.6.4 Implications for volcanic hazards inferred from geophysical models ........ 114
4.7 Conclusions ............................................. 116
4.8 Acknowledgements .......................................... 117
4.9 References ............................................... 118

## 5 Mass Transfer Processes in a Post Eruption Hydrothermal System: Parameterisation of Microgravity Changes at Te Maari Craters, New Zealand

5.1 Introduction ............................................. 128
5.2 Data ..................................................... 129
  5.2.1 Gravity measurements .................................... 129
  5.2.2 Deformation measurements and models ....................... 133
List of Tables

Table 1.1  Summary of thesis research objectives and science questions addressing the research theme. .......................................................... 9

Table 3.1  Summary of model parameters for each observation interval from peak of KDE distribution. * Lake surface elevation is 2160 m a.s.l. Model I are the lower $\chi^2_{red}$ models, and model II are the higher $\chi^2_{red}$ models for the 2013 to 2014 interval. ....................................................... 68

Table 4.1  Summary of physical properties for rock types within the TgVM. All Volcanic includes all Andesite Lava and Pyroclastic samples. Number of samples of each rock type is given by $n$. ................................................. 93

Table 4.2  P wave velocites from Jolly et al. (2014) converted to density using the relationships in Brocher (2005). ........................................... 114

Table 5.1  Microgravity change data from 2014 to 2017. Microgravity change values (Deltag) are in $\mu$Gal, as are the standard error (STDER) values. TGM10 was not able to be surveyed after 2015. Height change values (DeltaH) are in m and are computed at the benchmark locations using the deformation model of Hamling et al. (2016). Height change errors are nominally 0.01 m. .......................................................... 132

Table 5.2  Source parameters for the deformation data, from A), the finite element model of Hamling et al. (2016), updated for the gravity time periods and B), the best fit deformation model from the NSGA-II joint inversion. Xc is the centroid Easting, Yc is the centroid Northing. The reference elevation for depth conversion is 1490 m a.s.l. ............... 135

Table 5.3  Summary of the sill parameters for A) the gravity only model, and B) the best gravity model from the NSGA-II joint inversion. In A) All values are the peak KDE or ‘mode’. * indicates a double peaked distribution. Xc is the centroid Easting, Yc is the centroid Northing, UM is the thickness and density product. The reference elevation for depth conversion is 1490 m a.s.l. .................................................. 142

Table 5.4  Summary of physical properties used in finite element model. ........... 151
Table 5.5  Summary of processes occurring in hydrothermal system and if they are likely to result in gravity increases or decreases and inflation or deflation. ................................................................. 158

Table 6.1  Summary of hypotheses developed and tested in this thesis, with brief outcome description. ................................................................. 167
# List of Figures

<table>
<thead>
<tr>
<th>Figure</th>
<th>Description</th>
<th>Page</th>
</tr>
</thead>
<tbody>
<tr>
<td>1.1</td>
<td>Location of Laguna del Maule volcanic field, on the range crest of the Andes just west of the border between Chile and Argentina. LdM is shown as a yellow triangle. Other Holocene volcanoes are shown in pink triangles.</td>
<td>3</td>
</tr>
<tr>
<td>1.2</td>
<td>Ground displacement time series from comparing LdMVF to other well known volcanic deformation episodes, illustrating the remarkable longevity and amplitude of the uplift. Measurements are maximum point measurements from individual benchmarks. Modified from Le Mével et al. (2015).</td>
<td>5</td>
</tr>
<tr>
<td>1.3</td>
<td>Map of New Zealand’s Taupo Volcanic Zone (orange outline), showing Mt Tongariro at the southern end as a yellow triangle. The central TVZ is dominated by rhyolitic volcanism at Taupo, Maroa, Reporoa, Rotorua and Okataina, while the north and south are predominantly andesitic. Black lines are active faults showing the TVZ rift and the North Island axial fault belt.</td>
<td>6</td>
</tr>
<tr>
<td>1.4</td>
<td>Mechanism of LaCoste and Romberg type gravity meter used in this research (Marson, 2012).</td>
<td>10</td>
</tr>
<tr>
<td>2.1</td>
<td>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 $^{40}$Ar/$^{39}$Ar ratios (Andersen et al., 2017) 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.</td>
<td>25</td>
</tr>
</tbody>
</table>
Figure 2.2  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 2.3  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 2.4  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 2.5  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% H$_2$O. 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., 2017). B) Plot of magma density (contour lines in kg/m$^3$) 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 2.6  Phase diagram showing the evolution of H$_2$O 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$_2$O. The green shaded box highlights the magma temperatures derived by Fe-Ti geothermometry (Andersen et al., 2017). 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 2.7  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 2.8  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 range of pre-eruptive temperatures determined by Fe-Ti oxide geothermometry by Andersen et al. (2017), (see Figure 2.5). 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 two larger stars show the models that are most likely when all constraints are considered.
Figure 2.9  3D view of the preferred -600/+300 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.

Figure 3.1  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 centre of inflation and sill outline from Feigl et al. (2014). Ages are from $^{40}\text{Ar}/^{39}\text{Ar}$ ratios (Andersen et al., 2017) and red stars show post-glacial 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.

Figure 3.2  Residual gravity changes ($\Delta g$) between A) 2013 and 2014 (maximum 124 ± 12 µGal), B) 2014 and 2015 (maximum 60 ± 15 µGal), C) 2015 and 2016 (maximum 68 ± 16 µGal), overlain on ASTER GDEM-2 30 m digital elevation model. Contours in µGal with the same colour scale for all intervals. D) Earthquake hypocenters from OVDAS catalogue from April 2011 to November 2015. Laguna del Maule is outlined in blue. Inset) Time series showing vertical displacement of CGPS station MAU2 (established in 2012), close to maximum uplift, as well as number of earthquakes per day from OVDAS catalogue.

Figure 3.3  Geometry and parameters of the vertical prism model. The prism is vertically oriented (dip = 90°), Xc and Yc are the centroid coordinates. Depth is given as an elevation referenced to the mean altitude of the gravity benchmarks (approximately 2200 m a.s.l.). Strike is calculated in degrees east of north. $K_{hx}$, $K_{hy}$ and $K_{hz}$ are hydraulic conductivity directions.
Figure 3.4  Box plots of $\chi^2_{red}$ 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 labelled for the sphere model. The box shows the quartiles of the dataset, 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 inter-quartile range. In most cases there is very tight clustering of $\chi^2_{red}$ values, making the box and whiskers less obvious.

Figure 3.5  A) Contour plot of $\Delta g$ for 2013 to 2014 interval in $\mu$Gal. 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). 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. The lower $\chi^2_{red}$ models are labelled as I, and the higher $\chi^2_{red}$ models as II.

Figure 3.6  A) Contour plot of $\Delta g$ for 2014 to 2015 interval in $\mu$Gal. Maximum gravity change is $60 \pm 15 \mu$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). 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.
Figure 3.7  A) Contour plot of $\Delta g$ for 2015 to 2016 interval in $\mu$Gal. Maximum gravity change is $68 \pm 16 \mu$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). 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.

Figure 3.8  Plan sections of (A) normal, (B) Coulomb, and (C) mean stress at 1.5 km depth, from sill opening on receiver faults oriented at 40°, i.e., parallel to the Troncoso fault, dipping 70°E with a rake of -90°. Lake outline is shown in blue lines, simplified fault traces in black lines, and the deformation source of Feigl et al. (2014) as grey rectangle. Positive stress change produces fault unclamping. D-F) Mean stress change on the receiver fault plane resulting from the sill opening. These faults dip at 70° to the SE with a rake of -90°, i.e., normal faults. The oval and rectangular blue shaded area represents the region of mass increase from microgravity models over all time intervals.

Figure 3.9  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 A) 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.

Figure 4.1  Simplified geological map of the TgVM. The inset map shows the North Island of New Zealand with the TVZ and model area outlined in black. The dashed red line is the Pacific/Australian plate boundary. Coordinates are easting and northing in m using the NZTM projection.
Figure 4.2  Complete Bouguer anomaly data for A) Regional area around Mt Tongariro, contour interval 5mGal. The detailed 28 x 18km model area is shown in the black dashed rectangle. Black dots are GNS Science stations, blue dots are Cassidy stations, red dots are stations collected in this study. B) The residual CBA in the model area after removal of a 3rd order polynomial, contour interval 2mGal. Vent locations are shown in white triangles and stations as for part A. C) The residual CBA low pass filtered to 10,000m wavelength, contour interval 2mGal. Shown in all figures are the active faults (white lines). Coordinates are easting and northing in m using the NZTM projection.

Figure 4.3  A) Total magnetic intensity anomaly, contour interval 50nT. Black dots show flight lines. B) Reduced to Pole (RTP) map of the TMI anomaly, contour interval 50nT. Vent locations are overlain as white triangles, active faults overlain as white lines. Coordinates are easting and northing in m using the NZTM projection.

Figure 4.4  A) Radially averaged power spectrum of complete Bouguer anomaly data. Both corrected (dots) and non-corrected data (triangles) are shown. Top of source elevation estimates are based on the corrected data, while wavelength filter characteristics are based on the non-corrected data. Vertical lines highlight the wavelength segments tested in filtering. B) Radially averaged power spectrum of TMI data with top of source elevation estimates. Annotated line segments represent elevations of source layers.

Figure 4.5  Depth slices from the conventional (left column B - E) and stochastic (right column G - J) geologically constrained gravity inversions at 1350, 1150, 750, 550m a.s.l. Residual anomaly maps are shown in A and F. Overlain in black lines are active faults and vent locations as white triangles. Coordinates are easting and northing in m using the NZTM projection.

Figure 4.6  Depth slices from the conventional geologically constrained TMI inversion at 1350, 1150, 750, 550m a.s.l. (B - E). Residual anomaly map shown in A. Overlain in black lines are active faults and vent locations as white triangles. Coordinates are easting and northing in m using the NZTM projection.

Figure 4.7  Parallel view from south showing the top of greywacke basement surface colour coded by density. The topography has been raised above the basement surface for clarity. No vertical exaggeration.
Figure 4.8  A) Iso-surfaces of low density (2250 kg/m$^3$) in cyan and high density (2400 kg/m$^3$) in red. B) Iso-surface of low magnetic susceptibility (0.025 SI) in yellow and high magnetic susceptibility (0.09 SI) in purple. Perspective view looking from the south-east. No vertical exaggeration.

Figure 4.9  The TgVM hydrothermal system as shown by iso-surfaces of low magnetic susceptibility (0.025 SI) in yellow, and low density (2250 kg/m$^3$) in cyan. The area outlined in red represents the hydrothermal system as shown in Figure 4.10. Perspective view looking from the east south-east. No vertical exaggeration.

Figure 4.10  Slope angle (calculated on 15m DEM) and extent of demagnetised hydrothermal system (outlined in red). Areas of greatest risk of collapse are steep slopes within the hydrothermal system outline. Coordinates are easting and northing in m using the NZTM projection.

Figure 5.1  Location of microgravity network overlain on 0.5m resolution DEM of the pre 2012 eruption landscape. Red triangles are the gravity benchmarks. Red lines in the smaller figure indicate active faults, and white lines are roads. The reference station, TGKB, is shown as a red triangle in the insert figure. The purple shaded area is the vent and fissure formed during the August 2012 eruption.

Figure 5.2  Gravity changes (shaded contour lines) and line of sight (LOS) displacements (coloured dots) from Hamling et al. (2016) for the intervals, 2014 to 2015, 2015 to 2016 and January 2016 to December 2016. Positive LOS indicates subsidence. Black triangles are gravity benchmarks, purple shaded area is vent region. The red dot is the location of the point source mass (see Figure 5.3) and the red rectangle is the outline of the sill dislocation model from Hamling et al. (2016).

Figure 5.3  Curve fit for point source solution for all intervals. Dots are data points with error bars; curved lines are the fit of the data. For 2015–2014 $R^2 = 0.62$, in 2016–2015 $R^2 = 0.71$, and in 2017–2016 $R^2 = 0.85$.

Figure 5.4  Box and whiskers plots showing the $\chi^2_{red}$ for the range of gravity only model geometries tested. In each geometry segment the three observation intervals are shown. The box shows the quartiles of the data set, while the whiskers extend to show the rest of the distribution. Dots are outlier values.
Figure 5.5 Distribution of model depths for each time interval. The sphere and sill model depth distributions are shown as green and blue KDE plots from the gravity only inversions, while the best fit deformation solution is shown as a black dot. 141

Figure 5.6 Summary of the sill and sphere gravity model locations (A, B, C) and observed vs calculated data (D, E, F) for each gravity observation interval. The blue rectangles are the sill outlines, green dots are the sphere centroids. Gravity change contours are shown in black lines, with benchmarks as yellow triangles. The purple outline is the vent area. The black dashed square in A–C is the extent of the position ranges for the inversion. The observed data in plots D–F are shown with standard error bars. The calculated data are for the sill model. 143

Figure 5.7 Parameters of the Okada and Okubo models for joint gravity and deformation inversion. In our model the sill is horizontal and there are no strike or dip slip components. UM (density and thickness product) is required to be approximately 100 times greater than U3 (tensile displacement) to fit the observed gravity data (the diagram is not to scale). Note that U3 in our models is negative, representing a closure of the sill, not opening. Hydraulic conductivity directions are shown as $K_x, K_y, K_z$. 145

Figure 5.8 Summary of the joint gravity and deformation NSGA-II inversion models. A, B, C show the locations of the joint sill models for the 2015–2014, 2016–2015 and 2017–2016 time intervals, respectively. Grey rectangles represent the optimal Pareto set of models with the best gravity model in blue and best deformation in red. Observed gravity change contours are shown in black lines, with coloured dots as the observed line of sight (LOS) displacement. Positive LOS indicates subsidence. Yellow triangles are gravity benchmarks. D, E and F show the Pareto front of the RMS gravity vs RMS deformation solutions. Light blue dots represent gravity solutions within the 95% confidence limit of the best gravity solution, and light red dots are the equivalent for the deformation solutions. The best individual gravity and deformation solutions are shown in larger dark blue and dark red dots, respectively. The widespread distribution of the Pareto front indicates separate gravity and deformation sources. 147
Figure 5.9  Schematic cross section showing locations of gravity and deformation sources in relation to the hydrothermal system. The cross section shows a region of higher permeability rock (orange shading) above the gravity source allowing greater influx of meteoric fluids. Red shading above the dyke shows an area of higher temperature, where the liquid phase has been pushed out immediately after the eruption. Over time this region contracts (red arrows) as the dyke cools, allowing the liquid condensate to return, adding mass to the system (black arrows). Blue arrows on the gravity and subsidence sources show the direction of movement of these sources over time. The depth of the dyke intrusion is inferred from seismicity (Hurst et al., 2014) and the buoyancy boundary created by the large density contrast between greywacke basement and overlying volcanics. Greywacke basement and extent of altered vs unaltered volcanic rock from Miller and Williams-Jones (2016).

Figure 5.10  Conceptual evolution of the gravity and deformation sources after the August 2012 eruption. Open blue ellipses represent evacuated pores undergoing compaction, while narrower ellipses are already compacted pores. Light blue coloured circles are pores filled by precipitation infiltration, while darker blue circles are pore fluid expelled from deeper pores or condensation. The solid or dashed thick pink line represents the intact or broken condensate seal, respectively.

Figure 5.11  Temperature of fumarole F3 (dots) with 1D conductive cooling curve (line). Shaded boxes are the 3 gravity change intervals. The timescale is days after the August 6th 2012 eruption. Mass flux rates from the gravity data are annotated in each survey interval.
Chapter 1

Introduction

1.1 Motivation for this Thesis

Volcanic eruptions are the spectacular culmination of the movement of magma from its source region, through Earth’s crust, and then to the surface. Volcanic edifices built by eruptions may conceal complex internal architecture, including hydrothermal systems, and reservoirs and conduits that store and transport magma, leading to fundamental questions about their composition, geometry, and location. Ascending magma generates stresses, strains, and other time-varying geophysical and geochemical signals that are detectable on Earth’s surface, and are collectively termed ‘unrest’ (Phillipson et al., 2013). Whether an ascending magma reaches the surface depends on its internal properties, such as chemical composition, gas content, temperature and crystallinity, which affect magma density, compressibility and rheology, as well as the external effects of the crust density, strength and stress state. The combination of these internal and external factors ultimately determine whether an ascending magma will erupt or stall (National Academies of Sciences, Engineering, and Medicine, 2017). I propose therefore, that knowledge of a volcano’s internal structure, architecture, or anatomy, is critical, as it provides the necessary context for correct interpretation of unrest. If the underlying causes of unrest are understood, then eruption forecasting becomes possible, and the harmful effects of an eruption (e.g. Wilson et al., 2014) can be mitigated.

While this proposal may sound simplistic, it is worth considering that of the 1449 Holocene volcanoes listed in the Smithsonian Global Volcanism database, I estimate fewer than half of these volcanoes are monitored and less than half of those have some sort of multiparameter geophysical model. For example, in the United States fewer than half of the 169 potentially active volcanoes have even one seismometer to detect small earthquakes that accompany underground magma movement (National Academies of Sciences, Engineering, and Medicine, 2017). Of these monitored volcanoes, only a small percentage have some sort of geophysical model, specific to that volcano on which to guide interpretation of monitoring observations. In order to progress the use of physics-based models to understand the causes,
and forecast the outcomes of volcanic unrest (e.g. Anderson and Segall, 2013), it is critical that detailed physical property distributions of volcanoes, i.e. the volcano’s anatomy, be determined.

The importance of understanding a volcano’s internal structure is highlighted by silicic volcanism. Caldera forming eruptions require reservoirs of eruptible magma at shallow depth (Smith and Bailey, 1968; Cashman and Giordano, 2014). Whether fresh magma intrusion into these reservoirs results in eruption or not, depends on the rate of magma supply, existing reservoir size and composition, and the thermo-mechanical properties of the surrounding crust (Caricchi et al., 2014; de Silva and Gregg, 2014). If the influx of buoyant new magma greatly exceeds the long-term supply rate, then the critical magma chamber overpressure required to induce dyking in the magma chamber wall can be reached, resulting in eruption (Jellinek and DePaolo, 2003). While in other cases, external triggering, such as shaking from earthquakes on nearby faults, is required for eruptions (Gregg et al., 2015). Hence, for large silicic systems, knowing the physical properties of a magma reservoir, as well as the properties of the surrounding crust, is critical in determining magma pressures due to buoyancy. The geometry of the reservoir is important in determining how stresses will propagate from it as it inflates (e.g. Gudmundsson et al., 1997), and if pressures resulting from the intrusion exceed the strength of the overlying crust (e.g. Browning et al., 2015). Finally, knowing where the magma system is located in relation to tectonic structures (e.g. Saxby et al., 2016) allows us to determine whether external triggering is likely to be a factor in initiating eruption.

Intruding magma can perturb shallow hydrothermal systems, causing detectable mass changes and pressurisation in those systems (e.g. Jousset et al., 2000; Fournier and Chardot, 2012). Hydrothermal systems can be important in controlling the initial style of eruption, e.g. phreatic vs phreatomagmatic (Jolly et al., 2014), so distinguishing hydrothermal from magmatic unrest has important implications for hazard forecasting.

A volcano’s internal structure can be derived from geophysical potential field measurements (e.g. gravity and magnetic surveys) that map the distribution of rock density and magnetic minerals (e.g. Gudmundsson and Milsom, 1997; Cassidy et al., 2009; del Potro et al., 2013; Espindola et al., 2016). Spatial variations in these properties may relate to magma reservoirs and conduits, zones of hydrothermal alteration, tectonic faults and changes in the underlying basement (e.g. Finn et al., 2007). Time-varying gravity measurements (Carbone et al., 2017) can monitor mass and density changes caused by magma intrusion or hydrothermal fluid movement. Combining space and time varying potential field measurements with deformation data, together with analytic and numerical modelling, creates a powerful tool kit for understanding volcanic unrest and the context within which it occurs.

In this thesis, I present a series of studies illustrating novel approaches to modelling volcanic architecture and time-varying processes, using potential field and ground deformation data. Studies from two contrasting volcanoes are presented: Laguna del Maule volcanic field,
a highly productive rhyolite system in Chile, and Mt Tongariro, an active andesite system in New Zealand that experienced two eruptions in 2012. In each case, models of volcanic architecture are developed, followed by analysis of time varying processes occurring within them.

1.2 Field Sites

1.2.1 Laguna del Maule

Laguna del Maule volcanic field (LdMVF) is a large silicic, multi-vent volcanic field on the Chile/Argentina border, described by Hildreth et al. (2010) as ‘the superlative site of post-glacial silicic volcanism in the Andean Southern Volcanic Zone (SVZ)’ (Figure 1.1).

![Figure 1.1: Location of Laguna del Maule volcanic field, on the range crest of the Andes just west of the border between Chile and Argentina. LdM is shown as a yellow triangle. Other Holocene volcanoes are shown in pink triangles.](image-url)
LdMVF is located in a back-arc setting, 300 km east of the Nazca-South American plate subduction trench axis, 200 km from the coast, 130–150 km above the Benioff zone and 40–50 km above the continental Moho (Hildreth et al., 2010, and references therein). The field has been active since 1.5 Ma, producing at least 130 independent vents, including the 80 km³ Bobadilla Caldera-forming eruption at 950 ka. Since the last glaciation (<25 ka), at least 50 eruptions have occurred from 24 vents surrounding the 23 by 16.5 km lake basin (Singer et al., 2014). The most recent rhyolite lava is dated at 2.2 ka (Andersen et al., 2017). Eruptive products span the compositional range from basalt to rhyolite. However, since the last glaciation, all known eruptions have been rhyodacite to rhyolite and nested in a concentric ring around the lake, leading to the hypothesis that a large silicic magma body exists under the volcanic field (Singer et al., 2014).

Interest in the LdMVF increased with the detection of large ground displacement signals that began sometime between 2004 and 2007 (Fournier et al., 2010). Since then, deformation has continued at rates of up to 250 mm/year (Feigl et al., 2014; Le Mével et al., 2015) (Figure 1.2), covering an area of around 15 x 15 km, making LdMVF one of Earth’s fastest deforming volcano not actually in eruption. In comparison, other calderas showing uplift at much lower rates include Yellowstone, USA, at 70 mm/year between 2004 and 2006 (Chang et al., 2010); Long Valley, USA, at 40 mm/year from 1980 to 2000 (Newman et al., 2006) and Santorini, Greece, at 80 mm/year since 2011 (Parks et al., 2012). The 1.5 m of uplift at Campi Flegrei, Italy, between 1982 and 1984 (Amoruso et al., 2008), the 3.5 m of uplift at Rabaul, Papua New Guinea, (McKee et al., 1984) between 1971 and 1984, and the more than 4 m of uplift at Iwo Jima between 1977 to 2003, are the only comparable modern episodes of widespread uplift at a silicic volcanic centre.

In addition to the large-scale ground displacement, swarms of volcano seismicity have been recorded around the margins of the volcanic field, and CO₂ gas concentrations up to 7% by volume have been measured around the lake shore (Miller et al., 2014). Preliminary magnetotelluric (Cordell et al., 2015) and seismic (Wespestad et al., 2016) results map conductive and low velocity bodies beneath the volcanic field, suggesting regions of magma or other magmatic fluids.

In addition to the present day uplift a high stand shoreline created ∼10 ka has been tilted around 60 m to the north. This tilting may represent multiple episodes of intrusion, that importantly, can thermally condition the surrounding crust changing the conditions where storage of magma is preferred over eruption (Jellinek and DePaolo, 2003), depending on the size of the reservoir and the overpressure it produces. Therefore defining the volume and composition of the magma reservoir is of first order importance for long term eruption forecasting and hazard assessment.

These observations may represent the active assembling of a magma body capable of producing a caldera forming eruption, possibly the only such place on Earth where this is currently occurring. Official concern about the state of the volcano was sufficiently high
that in March 2013, Chile’s national geological agency, Sernageomin, raised the volcano alert level to Yellow for a period of 6 months, indicating that eruption was possible within weeks to months. However, as of 2017 an eruption has yet to occur, making the extended period of unrest remarkable. As such, LdMVF is a superb natural laboratory for the study of active volcanic processes that may produce a large-scale, potentially very hazardous rhyolitic eruption.

### 1.2.2 Mt Tongariro

Tongariro Volcanic Massif (TgVM) is a compound stratovolcano at the southern end of the Taupo Volcanic Zone (TVZ) (Figure 1.3). It includes >17 overlapping vents in an area of 5 x 13 km and has been active for the past 275 ka. Mt Ngauruhoe is named separately, but is geologically part of Mt Tongariro, and forms its highest peak. The last major period of cone building was around 10 ka when six eruptions of the Pahoka-Mangamate sequence occurred over a 200–400 year period from a 15 km long zone of multiple vents within the
TgVM. Nairn (2000) considered that this eruptive sequence was triggered by a regional rifting episode occurring within the southern TVZ at about 10 ka. Ngauruhoe cone was built within the last 7 ka (Moebis et al., 2011).

Figure 1.3: Map of New Zealand’s Taupo Volcanic Zone (orange outline), showing Mt Tongariro at the southern end as a yellow triangle. The central TVZ is dominated by rhyolitic volcanism at Taupo, Maroa, Reporoa, Rotorua and Okataina, while the north and south are predominantly andesitic. Black lines are active faults showing the TVZ rift and the North Island axial fault belt.

Vents for the post-glacial eruptions lie on a 25 km long NNE trending alignment within a major graben defined by regional normal faults and filled with Quaternary volcanics. Tertiary marine sediments overlie Mesozoic basement greywacke to the west and south of the graben. Nairn (2000) suggested the vents overlie a major basement fracture axial to the graben, indicating that structure may play an important role in the localisation of activity.
Most TgVM eruptive products are plagioclase-dominant pyroxene andesites and dacites, with strongly porphyritic textures, indicating their derivation from magmas that ascended slowly and stagnated at shallow depths (Nakagawa et al., 1998). A magnetotelluric study (Hill et al., 2015) imaged a low resistivity zone at 4–12 km depth, offset to the east of the Tongariro vent system, interpreted to represent a shallow crustal magma accumulation zone containing 18–45% melt fraction.

In the 19th and 20th centuries, the TgVM erupted from many vents including Ngauruhoe (last eruption 1975), Red Crater (1934), Ketetahi (1927) and Te Maari (1897 or 1928) (Scott and Potter, 2014). These historic eruptions produced lava flows, pyroclastic flows, ballistic fields up to 3 km from the vent, and ash dispersal.

On 6 August 2012, an eruption occurred from Upper Te Maari Crater, following approximately 3 weeks of precursory seismicity (Hurst et al., 2014) and changes to gas chemistry (Christenson et al., 2013), however no precursor deformation was observed (Fournier and Jolly, 2014). A second eruption followed without warning on 21 November 2012 (Scott and Potter, 2014). It is proposed that the eruptions occurred as a result of dyke intrusion into the hydrothermal system, even though no juvenile material erupted (Pardo et al., 2014).

### 1.3 Overarching Research Theme and Objectives

To fully benefit from the increasing use of probabilistic multi-physical, physics-based models of volcanic unrest (e.g. Anderson and Segall, 2013; Anderson and Poland, 2016), requires increasing detailed physics-based models of volcano anatomy. Such physics-based anatomy models will allow the proper construction of boundary conditions and physical property contrasts required for input into the physics-based unrest models. Hence, the overarching research theme of this thesis is that knowledge of volcanic architecture improves understanding of volcanic unrest. Under this theme, there are four main volcano specific research objectives, from which several science questions are posed (Table 1.1). The research objectives of this thesis are:

- To assess the current configuration of the magma reservoir at the LdMVF, and determine what hazard it represents after 10 years of unrest.
- To characterise the causes of unrest at LdMVF in relation to the magma, hydrothermal and tectonic systems, and propose a model that unifies the geophysical observations.
- To define the TgVM volcanic and basement structure, and hydrothermal system extent.
- To understand the response of the TgVM hydrothermal system to eruptions in 2012.
Each objective is addressed as a separate chapter, and the specific hypotheses developed and tested to address the science questions, are outlined in the preamble to each chapter. The investigation techniques are discussed in the following sections.

1.4 Methods

Here, I briefly describe the main geophysical methods and software used for data acquisition, processing and modelling. Many textbooks (e.g. Hinze et al., 2013; Long and Kaufmann, 2013) are devoted to the intricacies of gravity and magnetic surveying. The specifics of each survey are described in the following chapters, with detailed information on data collection and processing in the appendices.

1.4.1 Gravity

A gravity meter measures the force exerted (by gravity) on a mass, suspended by a zero-length spring (Figure 1.4). From Hooke’s Law:

\[ F = -Kx \]  

(1.1)

where \( F \) is the force, \( K \) is a constant related to the properties of the spring, and \( x \) is the length of the spring. Hooke’s Law implies the spring should be of zero length when no force is applied, which is not physically possible as the spring has a finite thickness. It is however possible to construct a ‘zero-length’ spring that obeys Hooke’s Law over the range of gravity variation found on Earth. The zero-length spring which counteracts the force of gravity on a test mass in the gravity meter, is constructed so that if the mass were removed, the force would be zero and the spring would have zero length (Hinze et al., 2013). The force is proportional to the strength of gravity, which is locally influenced by the density of underlying rocks. The zero-length spring concept is implemented in gravity meters with the necessary damping mechanisms and electronics required to make stable measurements (Figure 1.4, Marson (2012)). The common unit of gravity is the Gal and gravity variations are expressed in micro or milliGal \( (10^{-8}\text{ or } 10^{-5}\text{ms}^{-2}, \text{respectively}) \). Each spring-based gravity meter has a unique spring constant, \( K \), requiring measurements from different meters to be calibrated before being combined.
## RESEARCH THEME

Knowledge of volcanic architecture improves understanding of volcanic unrest.

<table>
<thead>
<tr>
<th>THESIS OBJECTIVES</th>
<th>SCIENCE QUESTIONS</th>
<th>HOW IT WILL BE ADDRESSED</th>
<th>RELEVANT CHAPTER</th>
</tr>
</thead>
<tbody>
<tr>
<td>Understand the current configuration of the magma reservoir at the LdMVF, and assess the hazard it represents.</td>
<td>What is the location, geometry and composition of current magmatic system?</td>
<td>Gravity measurements and thermodynamic modelling</td>
<td>2</td>
</tr>
<tr>
<td></td>
<td>What is the relationship to tectonic structures and eruption vents?</td>
<td>Gravity measurements and stress modelling</td>
<td>2 and 3</td>
</tr>
<tr>
<td></td>
<td>Is the system over-pressurised?</td>
<td>Interpretation of gravity models</td>
<td>2</td>
</tr>
<tr>
<td>Characterise the causes of magmatic and/or hydrothermal unrest at LdMVF.</td>
<td>Where are the active processes occurring in relation to magmatic, hydrothermal and tectonic systems?</td>
<td>Time-varying gravity measurements and analytic modelling</td>
<td>3</td>
</tr>
<tr>
<td></td>
<td>What is the interplay between these systems?</td>
<td>Numerical modelling</td>
<td>3</td>
</tr>
<tr>
<td>Define Mt Tongariro volcano and basement structure, and hydrothermal system extent.</td>
<td>What is the shape of the basement under the volcano?</td>
<td>Gravity and magnetic measurements and modelling</td>
<td>4</td>
</tr>
<tr>
<td></td>
<td>Does the basement influence the location of eruption vents?</td>
<td>Geologically constrained modelling</td>
<td>4</td>
</tr>
<tr>
<td></td>
<td>What is the extent of the hydrothermal system?</td>
<td>Geologically constrained modelling</td>
<td>4</td>
</tr>
<tr>
<td></td>
<td>What are the hazard implications from the derived models?</td>
<td>Interpretation of models</td>
<td>4</td>
</tr>
<tr>
<td>Understand the response of the Tongariro hydrothermal system to eruptions in 2012.</td>
<td>What mass and phase changes have occurred in the hydrothermal system post eruption?</td>
<td>Gravity change measurements with numerical modelling</td>
<td>5</td>
</tr>
<tr>
<td></td>
<td>Is the hydrothermal system depressurising or re-pressurising?</td>
<td>Gravity and deformation measurements</td>
<td>5</td>
</tr>
</tbody>
</table>

Table 1.1: Summary of thesis research objectives and science questions addressing the research theme.
I used two styles of gravity measurements in this research, both of which measure relative gravity changes, rather than absolute gravity. Spatial gravity variation mapping, termed Bouguer gravity surveying or mapping after its founder Pierre Bouguer, uses spatially distributed gravity measurements to map the density of underlying rocks (e.g. Hautmann et al., 2013). Microgravity, 4D, or time-varying gravity, measures changes in gravity with time, at benchmarks around the feature of interest (e.g. Carbone et al., 2017). Both styles of measurement use similar field techniques and instruments.

1.4.2 Magnetics

Measurements of Earth’s magnetic field consist of components due to the large-scale, slowly varying dynamo within Earth’s core, a smaller amplitude, more rapidly varying diurnal variation, and a component due to the distribution of magnetic minerals (i.e. magnetite) in the crust. In this thesis I am interested in anomalies due to the latter, which consist of two parts remanent ($J_r$) and induced ($J_i$) magnetisation. The measured magnetisation $J$ is the sum of $J_r + J_i$.

Induced magnetisation ($J_i$) is acquired when a material is exposed to a magnetic field, $H$. These are related through the magnetic susceptibility, $k$, a dimensionless number describing the ability of a rock to take on magnetisation, where $J_i = kH$. In volcanic rocks, $k$ ranges from $10^{-5}$ to $10^{-1}$. The inducing magnetic field is a vector with amplitude ($\sim$20,000 to
70,000 nT), inclination and declination, describing its magnitude and orientation. When the inducing field is removed the magnetisation associated with it disappears. Remanent magnetisation is the component of magnetisation permanently acquired when volcanic rocks cool below the Curie point (~560 °C) and may be different than the present day field depending on the age of the rock.

Magnetometers used in this research are the proton-precession type and operate by applying a high amplitude external magnetic field to a proton rich liquid (i.e. kerosene). The protons align with the applied field and when it is switched off, they realign (precess) themselves with the ambient field. Proton precession generates a weak AC magnetic field that is measured by the instrument, and whose strength is proportional to the ambient field. Variations of the proton precession magnetometer utilise the nuclear Overhauser effect (Kaiser, 1963). By adding nitroxide free radicals to the measurement fluid and stimulating it with a low power radio frequency field, the electron spin of the free radicals is aligned, which then couples to the protons via the Overhauser effect. Overhauser instruments require less power and can sample at faster rates as the electron–proton coupling can happen as measurements are taken.

In this research, magnetic field measurements were made by towing a magnetometer behind a light aircraft (Tongariro) and boat (Laguna del Maule). Samples were taken at 1 or 2 Hz along predefined lines (5–100 m depending on speed), with 400–500 meters between lines.

1.4.3 Software and modelling codes

I use a range of software to process, model and interpret the data collected. Different modelling codes are used in each chapter because each application has different complexities, requirements, data coverage and constraining information. Full descriptions of the modelling procedures are given in each chapter or the relevant appendices.

Matlab scripts (Battaglia et al., 2012) are used to initially process all raw gravity data and apply standard corrections for instrument drift, Earth and ocean tides (Battaglia et al., 2008). Python scripts written specifically for this research work are used to apply the standard corrections to Bouguer gravity data (Hinze et al., 2005), and to format the results for input into various modelling codes. Additional Python scripts I wrote further process microgravity data and produce final gravity change results. I wrote Python scripts to filter and apply diurnal corrections to the magnetic data. Nearly all the figures are created with Python.

Python scripts implementing the SimPEG inversion code (Cockett et al., 2015) are used for mathematically constrained Bouguer gravity models presented in Chapter 2. Interpretation of these geologically unconstrained inverse models is aided by thermodynamic models made using the rhyolite-MELTS for Excel software (Gualda and Ghiorso, 2015).
Matlab codes implementing a genetic algorithm (Carbone et al., 2008; Tiampo et al., 2000) are used to model analytic gravity sources of time varying gravity (Chapters 3 and 5). Analytic solutions are appropriate for gravity change models as there are fewer gravity measurements compared to Bouguer gravity surveys, so a full 3D mesh based model is not appropriate. Stress and strain modelling that explains a gravity change mechanism in Chapter 3 uses the Coulomb Matlab code (Lin, 2004; Toda et al., 2005).

Geologically constrained gravity and magnetic modelling (Chapter 4) uses the software package GOCAD® Mining Suite to construct the starting 3D geological model from published geologic maps which is directly coupled to the VPmg (Vertical Prism Magnetics Gravity) inversion routines (Fullagar and Pears, 2007; Fullagar et al., 2008). The geologic model is then adjusted to fit the observed geophysical data.

Comsol Multiphysics® is used for the finite-element forward model (Hickey and Gottsmann, 2014), simulating gravity change and ground deformation (Currenti et al., 2007) in Chapter 5, and incorporates physical property distributions determined from the previous models.

1.5 Thesis Roadmap

Chapters 2 to 5 of this thesis can be considered as independent studies addressing each of the research objectives. For convenience, the chapters are grouped by geographic location, and then by volcanic architecture, followed by active processes. Chapters 2 to 5 were published or submitted to journals prior to compilation of this thesis, and the style of writing (i.e. use of 'we') indicates that there is more than one author. The preamble to each chapter lists the contributions of each author; however all the research described in this thesis has been conducted by me (Craig Miller) unless otherwise stated. Preparing the chapters as a series of papers results in some duplication within the thesis, particularly in the introduction of each chapter, but this allows the chapters to be read somewhat independently. Each of the published chapters is formatted to meet the requirements of this thesis, but internet links to the journal formatted publication are given in a preamble at the start of each chapter. The preamble also introduces the chapter, poses volcano specific hypotheses to be tested, and places it in the context of the research objectives and science questions.

Chapter 2 presents a geophysical model of the magma reservoir at the Laguna del Maule volcanic field derived from measurements of Bouguer gravity, with interpretation aided by thermodynamic modelling of erupted lavas. This chapter is published in the peer-reviewed journal, Earth and Planetary Science Letters (Miller et al., 2017a).

Chapter 3 outlines a model derived from time-varying gravity measurements at Laguna del Maule. It illustrates how intrusion into the magma reservoir described in Chapter 2 locally alters the stress and strain fields at nearby faults, facilitating the movement of
hydrothermal fluids into faults above the reservoir. This chapter is published in the peer-reviewed Journal of Geophysical Research - Solid Earth (Miller et al., 2017b).

Chapter 4 describes a three-dimensional, geologically constrained model of gravity and magnetic data collected at Mt Tongariro. It maps the hydrothermal system, and provides a new model of the basement structure beneath the volcano, and an updated estimate of the volume of erupted material. This chapter is published in the peer-reviewed Journal of Volcanology and Geothermal Research (Miller and Williams-Jones, 2016).

Chapter 5 presents a combined microgravity and deformation model of changes to the hydrothermal system at Mt Tongariro, following the 2012 Te Maari eruptions, in the context of the model presented in chapter 4. This chapter has been submitted to the peer-reviewed Journal of Volcanology and Geothermal Research.

Chapter 6 summarises and extends the main findings, resolves the posed volcano specific hypotheses, and suggests routes for future research.

Appendices A and B give supplementary material for Chapters 2 and 3. Appendix C gives details of a magnetic survey at Laguna del Maule, which is in preparation for publication as part of a paper written by colleagues at Cornell University. Appendix D provides supplementary material for Chapter 4.

1.6 References


Browning, J., K. Drymoni, and A. Gudmundsson 2015, Forecasting magma-chamber rupture at Santorini volcano, Greece, Scientific Reports, 5, 15,785, doi:10.1038/srep15785.


Preamble

In the following chapter, I present a model of the present-day magma reservoir at LdMVF as inferred from a Bouguer gravity survey. I present the Bouguer anomaly data, a 3D inversion of the data, and thermodynamic models to aid interpretation of the gravity inversion model. By creating a detailed density model of the subsurface, this chapter addresses the first research objective in Table 1.1 to ‘Understand the current configuration of the magma reservoir at LdMVF’, and answers the research questions related to the magma reservoir location, geometry and composition, its relationship to tectonic structures and eruption vents, and the degree of pressurisation of the system. The following hypotheses are posed:

- There is a silicic magma body that underlies the LdMVF.
- An inflating sill at 5km depth is the only active magma body at LdMVF.
- The present-day magma body under LdMVF is not large enough to have been the source of all Holocene eruptions at LdMVF.
- The present-day magma body under LdMVF is not overpressurised.

Chapter 2 provides the context needed to interpret active processes discussed in Chapter three. This chapter is published in Earth and Planetary Science Letters by Craig Miller, Glyn Williams-Jones, Dominique Fournier and Jeff Witter. The journal formatted article is available at http://doi.org/10.1016/j.epsl.2016.11.007. The author contributions are as follows. CM designed the gravity survey, collected and processed the data, ran the gravity inversions and MELTS models, interpreted the data and models, and wrote the manuscript. GWJ participated in data collection. DF wrote the inversion code and assisted CM with its implementation. JW provided guidance on the interpretation of the MELTS results. Additional material is presented in Appendix A. All authors reviewed the manuscript prior to submission to the journal.
Chapter 2

3D Gravity Inversion and Thermodynamic Modelling Reveal Properties of Shallow Silicic Magma Reservoir Beneath Laguna del Maule, Chile

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 km$^3$, low density, volatile rich magma reservoir, at around 2 km depth, containing at least 85% melt, hosted within a broader 115 km$^3$ 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/m$^3$) with the surrounding crust exists, the lithostatic load is 50% greater 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 km$^3$ Nieblas (unit $rln$) 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.

2.1 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) framework (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 are 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 magmatic system and the hazard it presents. Importantly, the combined analysis may offer significantly different interpretations than if the gravity data are considered in isolation. We apply this method to image a shallow rhyolite magma reservoir beneath the Laguna del Maule volcanic field (LdMVF) and determine its key physical properties.

The LdMVF is located on the range crest of the southern Andes at 36°S, bordering Chile and Argentina and is geographically characterised by a lake enlarged around 19 ka, following the damming of its outlet by the Espejos rhyolite lava flow (unit rle). The LdMVF comprises the largest concentration of high silica rhyolite in the Andes with at least 50 post-glacial eruptions since 25 ka; including four between 3.3 and 2.1 ka (units rsl, rcd, rln and western rcb) (Andersen et al., 2017). These eruptions were from 24 vents and produced 15 rhyodacite, and 24 rhyolite lava flows and domes (Singer et al., 2014). The positions of the eruptive units and post-glacial vents are shown in Figure 2.1. Previous large volume silicic eruptions include a dacite ignimbrite at 1.5 Ma (igsp), and a rhyodacite tuff associated with the Bobadilla caldera at 0.95 Ma (igcb) (Hildreth et al., 2010). The post-glacial decrease in mafic eruption frequencies, and their spatially peripheral location, is inferred as evidence for a large, low density, high silica magma system acting as a density barrier for ascent of mafic magma.

Since 2007, the LdMVF has experienced uplift at rates over 20 cm/year (Le Mével et al., 2015), which Feigl et al. (2014) and Le Mével et al. (2016) interpret as resulting from inflation of a sill at around 5 km depth beneath the lake caused by the addition of new magma into the reservoir. By defining the location of the present day magmatic system and its physical properties, we provide the context to accurately interpret and understand the ongoing unrest and create a framework for interpreting monitoring data. From our model we estimate the melt vs crystal proportions and their distribution, the degree of pressurisation of the system, and investigate the implications of ratios of magma intruded to erupted. We consider the location of the magma system in relationship to the eruptive vents, the local tectonic framework and the ongoing deformation, and throughout we consider the hazard implications of our model.
Figure 2.1: 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., 2017) 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.
2.2 Gravity Data Collection and Reduction

In 2014, 2015, and 2016, we collected 239 terrestrial gravity measurements within a 10 km radius of the lake shore (Figure 2.1). We took measurements every 500 m around the accessible shoreline, on islands, and in a series of radial profiles extending 3–4 km from the lake. Thirty measurements were collected farther from the lake to define the regional gravity field. For precise positioning we used a Leica 530 GPS receiver with 15 minute occupation time per station; with maximum baseline lengths of 10 km to the Observatorio Volcanologico de Los Andes del Sur (OVDAS) CGPS network, we obtain height errors of better than 15 cm (∼0.045 mGal) at 95% confidence level. We correct the daily raw gravity data for Earth tide and drift using Gtools (Battaglia et al., 2012) to produce gravity values relative to our local base. We convert our local gravity values to absolute values using two nearby absolute gravity stations (S. Bonvalot pers. comm.) to a measured accuracy of 0.016 mGal.

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

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

(2.1)

where $\rho_{TC}$ is the terrain correction density, $\rho_W$ 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 2.2 and details in Appendix A. 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.
Figure 2.2: 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.
2.3 Gravity Anomaly

A striking gravity low embedded within an east-west trending gradient is the dominant feature of the Bouguer anomaly map (Figure 2.3A). 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 2.3B). 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 NNE and centred on the south shore of the lake, coincident with the centre of inflation modelled by Feigl et al. (2014). The gravity low is contained within the ring of post-glacial rhyolite vents that encircle the lake. Localised gravity highs up to 9 mGal, occur to the east of the lake.

Using Gauss’s theorem, where:

$$\Delta M = \frac{1}{2\pi G} \int \int \Delta g(x, y) \, dx \, dy$$  \hspace{1cm} (2.2)$$

we calculate the mass deficiency, $\Delta M$, required to produce the main gravity anomaly, $\Delta g$, of $1.1 \times 10^{13}$ kg, where $G$ is the gravitational constant, contained within an area $(x, y)$ of 95 km$^2$. This mass is independent of the geometry of the body within which it is contained.

2.4 Inversion and Interpretation

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

We construct a 3D mesh with 250 x 250 m horizontal cell size, with vertical cell size increasing by a factor of 1.2 from 100 m in the top 500 m, to 350 m at the base of the grid at -5000 m a.s.l., resulting in 190,440 cells. The mesh covers a subset of the Bouguer gravity anomaly where the station density is highest, using 192 of the 239 stations (rectangular box in Figure 2.3B). Note that there is no interpolation of data in the inversion routine, only the measured gravity data are used.
Figure 2.3: 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.
The two main elements of the inverse problem are the data misfit, and the regularisation scheme. Data misfit is a metric which measures the difference between observed and predicted data, while regularisation is a metric that is constructed to evaluate the model’s agreement with assumptions and prior knowledge. We define the data misfit of the model $\phi_d (m)$ as:

$$
\phi_d (m) = \frac{1}{2} \| W_d (F[m] - d_{\text{obs}}) \|_2^2
$$

(2.3)

where $d_{\text{obs}}$ is the observed data, $F[m]$ is a forward model that produces predicted data, and $W_d$ is a diagonal matrix with elements equal to $W_{d_{ii}} = 1/\epsilon_i$ where $\epsilon_i$ is the standard deviation of the $i$th data point. We set $\epsilon$ to 0.1 mGal for our model.

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

To produce a compact model, we inspect the distribution of model values and gradients of the $L_2$ norm model, $\epsilon_p$ and $\epsilon_q$, to find suitable values (See Figure A.2). The choice of $\epsilon_p$ and $\epsilon_q$ is arbitrary, depending on the level of compactness required, and we choose the 95th percentile of the $L_2$ model distribution after visual inspection of models with a range of values. The effect of $\epsilon_p$ and $\epsilon_q$ is to flatten the model value and gradient histograms towards having more values equal to $\epsilon_p$ and $\epsilon_q$, such that the model becomes less smooth, and more compact.

We solve the inverse problems using a scaled iterative re-weighted least squares optimisation approach and define a composite objective function as:

$$
\phi(m) = \phi_d (m) + \beta \phi_m (m)
$$

(2.4)

where $\beta$ is the tradeoff or Tikhonov parameter. A small $\beta$ results in a model that fits the data very well, but may include excessive structure so that $\phi_m (m)$ is large. Conversely if $\beta$ is large, the optimisation results in a large $\phi_d (m)$. To select the optimal $\beta$, we use an L-curve approach and stop adjusting $\beta$ when we have reached the target misfit. The optimisation is nonlinear so a gradient-based, Gauss-Newton method is used.
To run the inversion, we search for a perturbation of the model that reduces the objective function. The iterative optimisation process continues until the algorithm converges to a minimum and the misfit tolerance is achieved.

2.4.1 Model resolution

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

The results of the checkerboard (Figure 2.4) show that the outlines of all the prisms are resolved even for prisms with few data points. Prisms bordering the lake shore, where there are many stations closer to the top of the prisms, are better resolved, both laterally and at depth, compared to the outer prisms with fewer stations at greater depths. Prisms with fewer stations are generally modelled shallower, and with less density contrast than the true model. As such, we have confidence that for the size of the observed anomaly, the coverage of our stations should accurately determine the lateral extent of the anomalous body, especially in an east-west direction. Lateral boundaries across the lake in the north-south direction may be less well resolved and the density may be under estimated in areas of poor station coverage. The 500 m station spacing places an additional resolution limit on the smallest resolvable feature, in areas of good station coverage, of approximately 1 km.

2.4.2 Inversion constraints from thermodynamic models

The data collected in any geophysical survey are finite in number, while the physical property distribution in the Earth is continuous, resulting in an ill-posed problem with no unique solution. To constrain the inversion, and reduce the number of possible models, we investigate the likely range of densities of magma found at LdMVF. As the observed gravity anomaly is an integral of all magmatic phases (solids, liquids and exsolved gases), we must include the effect of H₂O and CO₂ and their behaviours in the system at different temperature and pressure conditions, when calculating magma system densities. See Appendix A for a discussion on the possible role of brines in the density model. While there is no
Figure 2.4: 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).
open vent degassing or surface fumaroles at the LdMVF, we have measured CO\textsubscript{2} soil gas concentrations of up to 7\% by volume in places around the lake shore, and in our modelling we test the sensitivity of the melt density to varying CO\textsubscript{2} concentrations as described below. For the volumetrically-dominant \textit{H}_{2}\textit{O} vapour phase we assume a closed system where that phase is retained. We use rhyolite MELTS for Excel (Gualda and Ghiorso, 2015) to calculate total magma system densities at equilibrium conditions from whole rock analyses representative of the most recently erupted rhyolites (\textit{rsl, rcd, rln}), over a range of temperature and pressure conditions, representing various proportions of vapour, liquid and solid magma. We note that evolution and degassing of rhyolite magma may not always strictly occur at equilibrium conditions (e.g. Gonnermann and Manga, 2005) so that the melt may be more or less saturated in volatiles than our calculations suggest. As such, the calculated densities of the magma system at equilibrium should be considered as minimum values.

The rhyolites contain a narrowly varying range of SiO\textsubscript{2}, between 73.5 to 75 wt\% (Andersen et al., 2017), and we use an average composition for our analysis. We also compute densities from rhyodacites \textit{rdcn} and \textit{rdcd} (average 69.4 wt\% SiO\textsubscript{2}), and andesites \textit{asd} and \textit{mcp} (63 and 53 wt\% SiO\textsubscript{2} respectively) to test our model sensitivity to different compositions, as the magma reservoir is unlikely to be solely rhyolite (see Figure A.3). Common biotite, sparse amphibole and very rare pyroxene in the silicic units indicates high water content, so we used 5 wt\% \textit{H}_{2}\textit{O}, derived from plagioclase-glass hygrometry (range 4.75-7.25 wt\%, N. Andersen pers. comm. July 2016), and in the absence of melt inclusion information, fixed \textit{CO}_{2} at 100 ppm, typical of arc-rhyolites (Gonnermann and Manga, 2012). As a sensitivity test, we calculate rhyolite densities using water contents of 3, 4 and 5 wt\% and found that a minimum of 4 wt\% \textit{H}_{2}\textit{O} is required to reproduce the observed gravity (Figure A.4). The effect of increasing water content is to decrease the total density of the system for any given pressure or temperature. At the low pressures present at the depth of our gravity model, \textit{CO}_{2} is already exsolved, so varying its concentration several fold makes only negligible density difference and we do not consider further the role of \textit{CO}_{2}.

Calculated rhyolite magma system densities (Figure 2.5) vary from 1300 to 2100 kg/m\textsuperscript{3}, over the pressure range of 50 to 120 MPa (\sim 2 to 5 km depth for a crustal density of 2400 kg/m\textsuperscript{3}), and temperature range of 700 to 920 °C (approximate solidus to above liquidus), giving a range of density contrasts from the gravity reduction density of -1100 to -300 kg/m\textsuperscript{3}. Feigl et al. (2014) model a sill at \sim 5 km depth (120 MPa), which we use as a constraint for the upper end of the likely pressure range. The rhyodacite density varies from the rhyolite density by a maximum of only 80 kg/m\textsuperscript{3} making the distinction between the two compositions not explicitly resolvable in our model. Andesite densities at 1020 °C geothermometer temperatures (Andersen et al., 2017) with 3 wt\% \textit{H}_{2}\textit{O} at 90 MPa, are 2200 to 2300 kg/m\textsuperscript{3}, a contrast of -200 to -100 kg/m\textsuperscript{3}.

Figure 2.5A is annotated with isolines of the percentage rhyolite melt in the system at each pressure modelled. The green shaded area represents the temperature range of the
Figure 2.5: 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., 2017). 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.
Holocene rhyolite and rhyodacite magmas prior to eruption (∼790 to 854 °C), calculated from the Fe-Ti geothermometer and provide further constraints on the range of probable densities.

The shape of the temperature vs density curves reflects the importance of including the full thermodynamic, crystallisation and volatile phase effects in the density calculations, not simply calculating the density of the liquid phase (Figure 2.6). Below the liquidus, the system density decreases, reflecting the start of feldspar crystallisation, exsolving H₂O from the melt. When quartz and K feldspar start to crystallise, H₂O is exsolved further lowering the total density of the magma system. However, at LdMVF quartz + K feldspar crystallisation is not observed in the rhyolite mineralogy, implying the temperature of the system remains above this point (green box in Figure 2.6). The high temperatures also limit the amount of exsolved H₂O to around 2.5 wt%. The implication of these dynamics and the importance of modelling the vapour phase, with respect to the interpretation of gravity data at active magmatic systems, is explored further in the Discussion section 2.5.1.

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

2.4.3 Inversion results

First, we consider which of the suite of geophysically feasible models is most viable from a geological perspective. The models that are valid over the greatest ranges of pressure conditions are from -600 to -400 kg/m³ (Figure 2.5). For example, the -600 kg/m³ model is valid from 80 to 100 MPa, or 3 to 4.2 km depth, while the -400 kg/m³ model is valid from 100 to 120 MPa, or 4.2 to 5.1 km depth. The central body of the -500 and -600 kg/m³ density contrast models have very similar volumes (∼30 km³), indicating that at these values the model is less sensitive to changes in density and these are our preferred models. While the more extreme density contrast models, -900 to -1100 kg/m³, are geophysically feasible, they are unable to represent the bulk of the magma reservoir when thermodynamic data are considered as they are valid only for very low pressures and result in low melt proportion magmas (<50%) inconsistent with geologic observations of crystal poor lava. In the lowest density contrast models (-200 and -300 kg/m³), a poorly defined low density body extends to the base of the model volume. There is no difference between the Lₚ norm and L₂ norm models, indicating that at these density contrasts, there is insufficient mass available to reproduce the gravity anomaly. These two models are not considered geologically reasonable, and are not discussed further.

The preferred inversion model of +300 to -600 kg/m³ (Figure 2.7) converges with an RMS of 0.032 mGal, in 175 seconds on an Intel i7, 2.7 GHz processor with 16 Gb ram. The distribution of observed minus calculated values are normally distributed around zero.
Figure 2.6: Phase diagram showing the evolution of H$_2$O 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$_2$O. The green shaded box highlights the magma temperatures derived by Fe-Ti geothermometry (Andersen et al., 2017). 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 2.7: 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.).
and show no obvious geographic bias (see Figure A.5). The shallowest layers of this model (1500 and 1000 m depth slices in Figure 2.7) show a ring of higher density material broadly coincident with vent locations to the west and south of the lake, while in the north, the Espejos vent, lava flow (rd) and pumice deposit is coincident with lower density material. The high - low - high density pattern across the north shore of the lake is present in all depth slices. The western boundary of this feature is broadly coincident with the inferred boundary of the Bobadilla caldera (Hildreth et al., 2010). The inferred caldera boundary is 1–2 km to the west of the gravity boundary, however its location is poorly constrained as it is buried by more recent volcanism.

Depth slices at 1500 and 1100 m a.s.l. show an irregular shaped low density region (contrast -200 kg/m$^3$, outlined with a dashed blue line in Figure 2.7), covering most of the lake and shoreline, and extending to the north under the Espejos rhyolite lava flow (rd) and pumice deposits. This is similar to the extent of the shallow conductor mapped with magnetotellurics by Cordell et al. (2015). At 0 m a.s.l. a low density body (contrast -600 kg/m$^3$) appears in the central portion of the lake. At -1100 m a.s.l. (approximately 3 km depth), the main features of the model are a central low density body and a finger of lesser density contrast material to the north. The central low density body sits within a 1–2 km wide ring of lower density contrast material, shown as the dashed red line in slice -2150 m, which is broadly coincident with the location of a sill modelled by Feigl et al. (2014) to explain present day deformation.

The main feature of our gravity model is a central low density body of uniform density throughout its thickness. Gravity is inherently poor at resolving horizontal layering that a magma system may exhibit, so our density model reflects the average bulk density of the system. The possible composition of this body is outlined in the Discussion, section 2.5.1. The volume of the central body at each model density contrast is shown in Figure 2.8 along with the proportion of melt possible at each density contrast and pressure range. Volumes of the anomalous body range from 6 to 50 km$^3$ for density contrasts of -1100 to -400 kg/m$^3$, respectively. The overall effect of different density contrast models is to swell or shrink the central body, whilst retaining its same general shape (Figure A.6).

At -500 kg/m$^3$, the base of the body is around -4000 m a.s.l. At -600 kg/m$^3$, the central body shrinks with a lower surface at around -3500 m a.s.l., while at -700 kg/m$^3$ and higher contrasts, the base of the body stabilises at around -3000 m a.s.l. Thus, the thickness of the central body varies between 3 and 4 km over a 200 kg/m$^3$ density range. A 3 dimensional representation is shown in Figure 2.9 for the -600 kg/m$^3$ density contrast model and shows a 30 km$^3$ body hosted within a 115 km$^3$ body of less density contrast.

### 2.4.4 Alternate scenario

We investigate an alternate scenario of a thick, low density surface layer to explain the residual Bouguer anomaly. Preliminary results from MT surveys (Cordell et al., 2015) in-
Figure 2.8: 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 range of pre-eruptive temperatures determined by Fe-Ti oxide geothermometry by Andersen et al. (2017), (see Figure 2.5). 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 two larger stars show the models that are most likely when all constraints are considered.
dicate a low resistivity layer extending from near surface to several hundred metres depth and interpreted as hydrothermal alteration or conductive lake sediments. This conductive layer covers much of the lake basin, with a footprint wider than the extent of the main Bouguer anomaly. Hydrothermal alteration often results in low density (e.g. Miller and Williams-Jones, 2016) from the chemical alteration of rock to low density clays. Recent field mapping, (J. Fierstein, pers. comm. 19th May 2016), identified a probable vent area under the lake, from an ignimbrite deposit mapped to the east, which is likely to have produced a thick accumulation of low density pyroclastic material within the lake basin. To test this scenario, we create a 500 m thick layer extending from the surface, with a density contrast of -300 kg/m³. The extent of the surface layer is the dashed blue outline in Figure 2.7, slices 0 and 1500 m a.s.l.. The cells inside the surface layer were fixed in the inversion, with the remaining model cells allowed to vary, so that the effect of the layer is included in the inversion but it is not adjusted. The ability to fix cells demonstrates the flexibility of SimPEG and allows us to easily explore alternate scenarios through a mix of forward and inverse modelling. The inversion result (Figure A.7) still requires considerable material beneath the surface layer to replicate the magnitude of the observed anomaly, and we conclude that the Bouguer anomaly cannot be solely caused by shallow low density material.

2.5 Discussion

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

2.5.1 How much melt is present, and of what composition?

The percentage of melt in a magma reservoir is a key determinant in that system’s ability to erupt. By using a thermodynamic modelling approach, rather than simply calculating melt densities, we are able to extract proportions of 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 2.5 and 2.8 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 (≈ 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., 2017). 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/m$^3$) with the background density (Figure A.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 given the well defined spatial extent of the gravity anomaly. The density contrast of andesite magma is close to that ascribed to the larger rim surrounding the active reservoir (red dashed line Figure 2.7), and it is possible that this rim is of intermediate composition and high crystal proportion (∼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 ∼38 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 as density and volume scale together to produce the observed gravity signal.

The rhyolite melt proportions are calculated using 5 wt% H$_2$O, 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$_2$O at 90 MPa, ∼95% melt is calculated at a density contrast of -600 kg/m$^3$, compared to ∼60% melt at 5 wt% H$_2$O. 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. (2017) 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 2.8. 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 magma (<50% melt). This reservoir is surrounded by a cumulate mush e.g. ‘root zone’ of Charlier et al. (2005), of intermediate composition and densities close to the host rock. A density zoned magma reservoir is consistent with a hypothesis that each of the LdMVF eruptions tapped only the top of the magma system. As melt proportion falls below 50%, it becomes more difficult to erupt, favouring the eruption of only the shallowest, crystal poor part of the system. The 50% melt fraction is also within the zone of most efficient extraction of melt from crystal mush, so the magma reservoir may be in a suitable configuration to repeatedly generate a cap of low crystal content, eruptible rhyolite overlying a crystal rich mush. Importantly for hazard considerations, Castro and Dingwell (2009), from the 2008 Chaiten eruption, and Castro et al. (2013) and Jay et al. (2014), from the 2011–12 Cordon Caulle eruption, showed that, near liquidus, volatile rich rhyolite magma, such as is present at LdMVF, is very fluid and can migrate rapidly from source to eruption site, resulting in very little warning prior to eruption.

### 2.5.2 Is the LdMVF magma system over-pressured?

We investigate whether the large density contrast of the imaged shallow magma reservoir with the surrounding crust is enough to generate magma overpressure that exceeds lithostatic pressure. Gregg et al. (2015) provide a simple elastic 1-D formulation for calculating the over pressure from a density contrast, taking into account the restoring force imposed by the lithostatic load ($P_{\text{litho}}$) on the magma pressure head ($\Delta P_{\text{buoy}}$). Note that this formulation ignores likely important effects of temperature and host rock viscosity (Jellinek and DePaolo, 2003), but serves as a useful first order approximation in the absence of other detailed information about the rheological state of the crust. The overpressure $OP$ is defined as:

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

(2.5)

or

$$OP = \Delta \rho gh_{\text{ch}} - \rho_r gz$$

(2.6)
where $\Delta \rho$ is the density contrast between magma, $\rho_m$, and host rock, $\rho_r$, $g$ is acceleration due to gravity (9.8 m/s$^2$), $h_{ch}$ is the height of the magma reservoir and $z$ is the depth to top of the reservoir.

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

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

The northward tilting of the high-stand shoreline by up to 60 m in the last 10 ka indicates significant long term shallow intrusion into the crust. The large volume of intruded material required to generate this uplift, will have modified the thermal properties of the crust, increasing the local geotherm and lowering the wallrock viscosity. Jellinek and DePaolo (2003) showed that as wall rock temperature increases (hence wall-rock viscosity decreases) small volume magma chambers (10–100 km$^3$) are more likely to remain in storage than erupt.

We note that if we only consider the melt phase density (refer to Figure 2.6), then our estimates of buoyancy would be substantially incorrect as the melt only phase density is much closer to the host rock density, resulting in a much reduced buoyancy force value. Additionally, melt only density models will not consider the exsolution of a free vapour phase, which may lead to an incorrect assessment of the hazard potential and likelihood of an explosive gas driven eruption.

2.5.3 Magma reservoir volume vs erupted volume

The volume of magma erupted is often used as a proxy for the volume of magma intruded into a system, to gain insights into the overall size of the magmatic system and what proportion of the total system is eruptible at any given time. This ratio can vary between periods of quiescence and activity, with often higher intruded than erupted volumes during quiescent times. There are relatively few studies where both the intruded and extruded volumes are accurately known (White et al., 2006), usually because the intruded volumes are poorly defined either by lack of outcrop or from low resolution geophysical imaging. There are two parts of the LdMVF magmatic system whose volumes we can compare to erupted volumes. A broad cumulate mush (115 km$^3$) within which a smaller volume (30 km$^3$) of
active magma exists, from which an even smaller volume (<2 km$^3$) of magma is periodically erupted.

The volume of post-glacial lavas at LdMVF is estimated at about 7 km$^3$ with 30–40 km$^3$ of pyroclastic material (Hildreth et al., 2010, J. Fierstein, pers. comm. 19th May 2016), for a total minimum volume of at least 40 km$^3$. Comparing our modelled magma reservoir volume (including the cumulative mush margin, 115 km$^3$) to the total post-glacial erupted volume, we calculate an intruded:extruded (I:E) ratio of 2.9:1, similar to that found at large silicic systems such as Yellowstone (3:1) (White et al., 2006).

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

Andersen et al. (2017) propose that chemical and temporal trends of erupted LdMVF rhyolites imply the extraction of chemically distinct melts from a long-lived, compositionally evolving upper crust reservoir. Our I:E ratios are consistent with such a reservoir, and as we discuss in section 2.5.4 the location of this eruptible magma may be required to migrate within the larger system to feed the mapped eruption vents.

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

The LdMVF scenario, where small magma volumes are erupted from a long lived system, is similar to the small volume (<1 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 km$^3$ dense rock equivalent, Oruanui eruption. Additionally, Sutton et al. (2000) cate-
gorised 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 2.5.4.

2.5.4 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 3km depth, this would indicate a maximum vent distance of around 2.5km 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 2.5km 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 5km deep sill modelled by Feigl et al. (2014) as the cause of the ongoing deformation (Figure 2.9). 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
Figure 2.9: 3D view of the preferred -600/+300 kg/m³ density contrast model, illustrating the LdMVF magma system. The isosurface (volume = 30 km³) shown in purple is -600 kg/m³, and sits within a 115 km³ body of -100 to -200 kg/m³ 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.

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 un-
able 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), 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 be 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.

2.6 Conclusions

A -19 mGal gravity anomaly at the Laguna del Maule volcanic field (LdMVF) can be explained by a shallow, 30 km$^3$ 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 km$^3$ 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 post-glacial LdMVF vents, and stress models of dyke propagation from magma reservoirs require it to have moved over time to be in closer spatial proximity to older vents. Alternately, the present-day magma system is smaller than that which formed the early post-glacial eruptions, or lateral migration of magma from older eruption vents to the present location has occurred. Intruded to extruded volume ratios suggest that the Las Nieblas (rln) eruption tapped ∼7% of the magma reservoir volume and that the post-glacial LdMVF is characterised by eruptions of small aliquots of magma from the top of a larger reservoir, rather than the evacuation and catastrophic collapse of the entire reservoir.

The reservoir lies directly above the modelled source of the ongoing deformation and supports the model of Le Mével et al. (2016) that the deformation is caused by injection of mafic magma into the base of a reservoir. Even though a large density contrast with the crust exists, we calculate that the magma reservoir is not over pressured, and that eruption due to buoyancy effects alone is unlikely. In addition, the intrusion of large volumes of magma in the last 10 ka as evidenced from the tilted shoreline, are likely to have thermally conditioned the surrounding crust into a state where small volumes (10–100 km$^3$) of magma remain stored rather than erupted. The western margin of the magma reservoir is in the hanging wall of the hypothesized Troncoso fault and we suggest that the location and orientation of the reservoir may be influenced by local tectonics which may also provide the trigger required for the next eruption.

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

2.7 Acknowledgements

We thank Basil Tikoff and University of Wisconsin-Madison for loan of the gravity meters and logistical support. We acknowledge the many field assistants involved with data collection, especially Daniel Diaz and students from Universidad de Chile-Santiago, Daniel Cabrera, Valentina Reyes, Ariel Figueroa, Daniella Calle, Gustavo Peréz, as well as Alex
We also thank Hélène Le Mével for discussions on the gravity model. We thank Diego Lillo and Sergio Morales from OVDAS, Sernageomin, for field assistance, Loreto Cordova for the CGPS base station data and Carlos Cardona for the 1D velocity model. We acknowledge Energy Development Corporation, Chile, for sharing their gravity dataset. Thanks to Andreas Tassara for bathymetry data and to Sylvain Bonvalot and Bureau Gravimetric International for establishing the absolute gravity stations. Thanks to Nathan Andersen and Brad Singer for whole rock analyses, discussions on the MELTS results and their implications, as well as to Natalia Deligne for discussions on presentation of those results. We wish to thank Don Luis Torres Jara, the Alcade de Mar, for his hospitality, nautical assistance and trucha ceviche. Field work was funded by NSF Integrated Earth Systems grant EAR-1411779 and EAR-1322595. C.M is supported by GNS Science Core Funding, EQC New Zealand, Mitacs Accelerate Canada and Mira Geoscience. We gratefully acknowledge two anonymous reviewers and the editorial advice of Tamsin Mather that improved this manuscript.

2.8 References


Preamble

In the preceding chapter I introduced a model of the magma reservoir at LdMVF, derived from Bouguer gravity data and thermodynamic modelling. In the following chapter, I discuss active processes that occur within the volcano architecture presented in Chapter 2, as detected by time-varying gravity measurements. This chapter therefore addresses the second thesis objective, ‘Understand the causes of magmatic and/or hydrothermal unrest’ (Table 1.1), and answers the science questions related to where the active processes are occurring in relation to the magma reservoir, hydrothermal system, and local fault systems. I propose a mechanism that links active processes in the magma system to changes detected in the hydrothermal system. The changes detected by microgravity are not discernible from the deformation results, and this highlights the benefit of gravity measurements to complement deformation monitoring, to help reveal the full dynamics of magma, tectonic and hydrothermal system interactions. In this chapter the following hypotheses are posed:

- Microgravity changes through time at LdMVF can be explained by the deformation changes observed by InSAR.

- Microgravity changes through time at LdMVF can be explained by mass changes within the shallow hydrothermal system.

- Sill opening at depth changes the stress field around regional scale faults at LdMVF, allowing fluids to migrate in.

Chapter 3 is published in the Journal of Geophysical Research (Solid Earth), by Craig Miller, Hélène Le Mével, Gilda Currenti, Glyn Williams-Jones and Basil Tikoff. The journal formatted article is available at http://doi.org/10.1002/2017jb014048. The author contributions are as follows. GWJ, HLM, and BT set up the microgravity network and participated in data collection in following years. CM participated in data collection, processed, modelled, interpreted the data, and wrote the manuscript. HLM provided height change data from InSAR measurements and deformation models listed in Appendix B. GC provided the inversion code and assisted with its implementation. All authors commented on the manuscript prior to submission.
Chapter 3

Microgravity Changes at the Laguna del Maule Volcanic Field: Magma Induced Stress Changes Facilitate Mass Addition

Abstract

Time-dependent, or 4D, 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 modelled 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 dyke 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 can not explain the gravity changes.

To account for this discrepancy and the correspondence in time of the deformation and gravity changes, we explore a coupled magma-tectonic interaction mechanism that allows for shallow mass addition, facilitated by deeper magma injection. Computing the strain and mean, normal and Coulomb stress changes, on northeast trending faults, caused by the opening of a sill at 5 km depth, shows an increase in strain, 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.
3.1 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 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 vs 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 3.1) is a large silicic, multi-vent volcanic centre 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/year (Feigl et al., 2014; Le Mével et al., 2015, 2016) have been observed and modelled as an inflating sill at ~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
Figure 3.1: 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 centre 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 post-glacial 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.
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.

3.2 Gravity Measurements and Processing

In March 2013, we established a network of 35 gravity benchmarks around the lake (Figure 3.1), 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). 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 3 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 (Table B.1). 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 dataset to GLM21 as in previous years this station was relatively poorly constrained, with few independent measurements. In 2016 we made 7 independent measurements of GLM21 (Figure B.2) to ensure that the error associated with the new reference is tightly constrained. Standard errors on the individual gravity measurements average 14 \( \mu \)Gal 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 dataset is 19 \( \mu \)Gal with a maximum of 47 \( \mu \)Gal. Further details and data (Table B.2) are given in the Appendix B.

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 pers. comm.), allowing us to check the absolute calibration of the D217 instrument. Repeat ties between the two absolute stations results 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 remains consistent and relative to D217. For the 2016 data, we calculate a mean difference between absolute and relative gravity differences for D217, of 10 $\mu$Gal, over the gravity range of 40 mGal.

The residual gravity change ($\Delta g$) is the difference between yearly gravity measurements, corrected for benchmark uplift and lake level variations, (see Battaglia et al., 2008, and references therein).

3.2.1 Free air gradient and uplift correction

The substantial uplift occurring between survey periods ($\sim$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 $\mu$Gal/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 Dalsen 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 $\mu$Gal/m and -338.7 ± 0.5 $\mu$Gal/m, respectively. We use a value of -335 $\mu$Gal/m for each measurement interval. Compared with the global value, this results in an increased correction of 5 $\mu$Gal/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 $\mu$Gal.

We apply the free air gradient to height change derived from a combination of InSAR and CGPS data (Table B.2). 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 post-seismic deformation (more details in Feigl et al., 2014). This adjustment uses a standard weighted least-squares algorithm to estimate 9 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 timeseries (Le Mével et al., 2015), which produces a gravity error of <3µGal.

### 3.2.2 Lake level and water table variations

The LdMVF surrounds a 54 km$^2$ lake that is up to 50 m deep. The elevation of the lake surface is artificially controlled, and can vary by several metres throughout the year. Daily records of the lake elevation are kept by the dam operators (El Ministerio de Obras Publicas, Direcccion de Obras Hidraulicas), and we use these to apply a correction for the changes in mass caused by changes in lake level. See Figure B.1.

Between the 2013 and 2014 surveys, the lake level rose by 1.1 m adding \( \sim 75 \times 10^6 \) 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 local aquifers are in a similar state from year to year.

### 3.3 Residual Gravity Change Results

We evaluate the residual gravity changes, \( \Delta g \), over 3 time periods between 4 surveys in March 2013, January 2014, March 2015 and February 2016 (Figures 3.2A, B, C). Data from some benchmarks were not usable in particular years, so not all time intervals contain the same benchmarks. Figure 3.2D 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. Figure B.2 shows a summary of benchmark occupations per year. Positive gravity changes, indicating mass addition, occur between each gravity survey.
Figure 3.2: Residual gravity changes ($\Delta g$) between A) 2013 and 2014 (maximum $124 \pm 12 \mu$Gal), B) 2014 and 2015 (maximum $60 \pm 15 \mu$Gal), C) 2015 and 2016 (maximum $68 \pm 16 \mu$Gal), overlain on ASTER GDEM-2 30 m digital elevation model. Contours in $\mu$Gal with the same colour scale for all intervals. D) Earthquake hypocenters from OVDAS catalogue from April 2011 to November 2015. Laguna del Maule is outlined in blue. Inset) Time series showing vertical displacement of CGPS station MAU2 (established in 2012), close to maximum uplift, as well as number of earthquakes per day from OVDAS catalogue.
3.3.1 2013 to 2014

Residual gravity changes in the 10 months between the 2013 and 2014 surveys, recorded at 26 benchmarks, show a maximum of $124 \pm 12 \mu \text{Gal}$, at benchmark GLM06 (Figure 3.2A). The gravity increase is distributed in a circular pattern around the maximum, with a slight elongation to the southwest. $\Delta g$ is zero around the north shore of the lake; however, the anomaly is open to the south where values of 10–20 $\mu \text{Gal}$ are recorded at the distal benchmarks.

Using Gauss’s theorem, where:

$$\Delta M = \frac{1}{2\pi G} \int \int \Delta g (x, y) \, dx \, dy \quad (3.1)$$

we calculate the excess mass, $\Delta M$, required to produce the gravity change anomaly, $\Delta g$, of $9.2 \times 10^{10} \text{kg}$, 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 non-parametric 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.

3.3.2 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 3.2B). The larger, northeast to southwest oriented anomaly to the south, reaches an amplitude of $60 \pm 15 \mu \text{Gal}$ at benchmark GLM27, while to the north, a smaller northwest to southeast anomaly has a maximum amplitude of $32 \pm 12 \mu \text{Gal}$ at benchmark GLM34. The larger northeast to southwest anomaly is coincident with the Troncoso fault (Figure 3.1). 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, 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 tunnelling activities as part of a hydroelectricity construction project. For the larger anomaly, we calculate a minimum excess mass of $3.96 \times 10^{10} \text{kg}$ over an area of 97 km$^2$. 
3.3.3 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 $68 \pm 16\,\mu\text{Gal}$ at benchmark GLM07 (Figure 3.2C). 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 $-37 \pm 26\,\mu\text{Gal}$ 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 $3.96 \times 10^{10}\,\text{kg}$ over an area of $88\,\text{km}^2$.

3.4 Residual Gravity Change Modelling

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 $\sim 1\,\text{m/year}$ at a depth of $\sim 5\,\text{km}$ (or -3 km above sea level, a.s.l.), 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 $2700\,\text{kg/m}^3$, i.e., that of basalt magma (Le Mével et al., 2016). The maximum gravity effect calculated is $\sim 3\,\mu\text{Gal}$, 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 whilst keeping the density constant. A 1 m thick sill requires an unrealistic density of around $15,000\,\text{kg/m}^3$, while keeping the sill at density $2700\,\text{kg/m}^3$ 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 co-located, 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).
3.4.1 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 spheriod, 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 non-unique 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 randomised starting model. This approach allows us to test the stability and sensitivity of the models over a wide range of randomised starting parameters and determine which geometry best explains the observed data.

3.4.2 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), prism (8). The sphere geometry is defined by its centroid coordinates (Xc, Yc, depth to centre), axis radius and density change. The prolate spheroid is defined by its centroid coordinates (Xc, Yc, depth to centre), 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 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 3.3) is defined by its centroid coordinates (Xc, Yc, depth to centre), length, width, strike, thickness and density change, with dip fixed at 90 degrees. The location of the source centroid is confined to be within the gravity anomaly area and the depth is constrained from 2000 m a.s.l. to -5000 m a.s.l. .

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

\[ \chi_{red}^2 = \frac{1}{v} \sum_{k=1}^{n} \frac{(O_k - E_k)^2}{\sigma_k^2} \]  

(3.2)

where \( O_k \) are the observed data, \( E_k \) are the calculated data, \( \sigma_k \) is the standard deviation of the observation, \( v \) 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 a F-test (Wackerly et al., 2007) on the \( \chi_{red}^2 \) values to determine which model geometry population best fits the data. Figure 3.4 shows box plots of \( \chi_{red}^2 \) calculated for each model type, grouped by observation interval. We additionally calculate the RMS fit (\( \mu \)Gal) for each model.
Figure 3.4: Box plots of $\chi^2_{red}$ 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 labelled for the sphere model. The box shows the quartiles of the dataset, 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 inter-quartile range. In most cases there is very tight clustering of $\chi^2_{red}$ values, making the box and whiskers less obvious.
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.

### 3.4.3 2013 to 2014 source model results

<table>
<thead>
<tr>
<th></th>
<th></th>
<th></th>
<th></th>
<th></th>
</tr>
</thead>
<tbody>
<tr>
<td>Xc (m)</td>
<td>363,544</td>
<td>364,286</td>
<td>361,192</td>
<td>363,615</td>
</tr>
<tr>
<td>Yc (m)</td>
<td>600,747</td>
<td>600,737</td>
<td>600,595</td>
<td>600,8391</td>
</tr>
<tr>
<td>Elevation m a.s.l *</td>
<td>543</td>
<td>250</td>
<td>875</td>
<td>619</td>
</tr>
<tr>
<td>Length (km)</td>
<td>6.2</td>
<td>6.2</td>
<td>6.4</td>
<td>5.9</td>
</tr>
<tr>
<td>Width (m)</td>
<td>170</td>
<td>170</td>
<td>110</td>
<td>110</td>
</tr>
<tr>
<td>Thickness (m)</td>
<td>145</td>
<td>43</td>
<td>66</td>
<td>31</td>
</tr>
<tr>
<td>Strike (deg)</td>
<td>43</td>
<td>65</td>
<td>40</td>
<td>76</td>
</tr>
<tr>
<td>Density change (kg/m^3)</td>
<td>307</td>
<td>884</td>
<td>156</td>
<td>279</td>
</tr>
<tr>
<td>Mass change (kg)</td>
<td>1.27 x 10^{11}</td>
<td>1.58 x 10^{11}</td>
<td>3.85 x 10^{10}</td>
<td>5.08 x 10^{10}</td>
</tr>
<tr>
<td>Volumetric flow rate, Q (m^3/s)</td>
<td>4.9</td>
<td>6.14</td>
<td>1.1</td>
<td>1.7</td>
</tr>
<tr>
<td>Hydraulic conductivity, K_hx (m/s)</td>
<td>6.65 x 10^{-6}</td>
<td>2.45 x 10^{-6}</td>
<td>9.92 x 10^{-7}</td>
<td>8.10 x 10^{-7}</td>
</tr>
<tr>
<td>Permeability, k (m^2)</td>
<td>6.54 x 10^{-11}</td>
<td>2.41 x 10^{-11}</td>
<td>9.75 x 10^{-12}</td>
<td>7.96 x 10^{-12}</td>
</tr>
</tbody>
</table>

Table 3.1: Summary of model parameters for each observation interval from peak of KDE distribution. * Lake surface elevation is 2160 m a.s.l. Model I are the lower $\chi^2_{\text{red}}$ models, and model II are the higher $\chi^2_{\text{red}}$ models for the 2013 to 2014 interval.

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. A F-test comparing the sphere and prism shows 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 3.1 shows a summary of the mode prism model parameters.

A summary of the prism model parameters and locations is shown in Figures 3.5A and C-H with the fit of the models to the observed data shown in Figure 3.5B. 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 (labelled I) are shallower, with lower density change (307 kg/m^3) and more northeast to southwest orientation (43° strike) than the higher $\chi^2_{\text{red}}$ model population (labelled 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 a.s.l. 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 x 10^{11} kg (c.f. 9.2 x 10^{10} kg from Gaussian integration), with mode width 170 m and mode thickness is 145 m.
Figure 3.5: A) Contour plot of $\Delta g$ for 2013 to 2014 interval in $\mu$Gal. 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). 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. The lower $\chi^2_{\text{red}}$ models are labelled as I, and the higher $\chi^2_{\text{red}}$ models as II.

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 dataset; however, we prefer the population with the lowest $\chi^2_{\text{red}}$.

3.4.4 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 3.2B), as the smaller anomaly to the north is shown to be not statistically significant. We remove the stations associated with the smaller anomaly so that they do not bias the inversion of the main anomaly. A F-test shows the prism model
population is a better fit than the sphere at \( p < 0.01 \) (\( F = 1214 \)) and has an RMS error of \( 6 \mu \text{Gal} \). A summary of the prism parameters is shown in Figures 3.6A and C-H with the fit of the models to the observed data shown in Figure 3.6B. All solutions show a uni-modal distribution and close association with the Troncoso fault with a mode strike of 40°. The mode elevation of the prism solutions is 875 m a.s.l. with a mode thickness of 66 m, mode width of 110 m, and mode mass increase of \( 3.85 \times 10^{10} \) kg (c.f. \( 3.96 \times 10^{10} \) kg from Gaussian integration).

---

Figure 3.6: A) Contour plot of \( \Delta g \) for 2014 to 2015 interval in \( \mu \text{Gal} \). Maximum gravity change is \( 60 \pm 15 \mu \text{Gal} \). 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). 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.

### 3.4.5 2015 to 2016 source model results

A summary of prism model parameters and locations is shown in Figures 3.7A and C-H with the fit of the models to the observed data shown in Figure 3.7B. The model solutions show
a uni-modal parameter distribution and a F-test shows 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 centred over the peninsula, oriented ESE to WSW (mode strike $76^\circ$). The strike is more east-west oriented than the previous intervals, but still subparallel to faults mapped in the lakebed. The mode elevation of the prisms is $619$ m a.s.l with a mode thickness of $31$ m, mode width of $110$ m, and mode mass increase of $5.08 \times 10^{10}$ kg (c.f. $3.96 \times 10^{10}$ kg from Gaussian integration).

Figure 3.7: A) Contour plot of $\Delta g$ for 2015 to 2016 interval in $\mu$Gal. Maximum gravity change is $68 \pm 16 \mu$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) Observes (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.
3.5 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 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 be no shallow sourced deformation signal, within the large deformation field accounted for by the sill model of Feigl et al. (2014). Below, we consider the nature of the intruding fluid and estimate the hydraulic properties of the fault zone. We propose a model that allows for mass increase without additional deformation, and show how the deformation and gravity change models are indirectly coupled by induced stress changes.

3.5.1 Nature of intruding fluid and mass balance

Our models show large, shallow mass increases between each observation interval, in good agreement with the 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 dyke 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/m$^3$ suggesting low density hydrothermal fluid rather than denser magmatic fluids are involved. A hydrothermal brine at 20°C with 10000 ppm total dissolved solids, and density 1005 kg/m$^3$, filling existing and newly created void space of 30%, would result in a density change of 302 kg/m$^3$. In 2014, measurements of CO$_2$ soil gas concentrations using a Vaisala GM70 probe, revealed CO$_2$ 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% CO$_2$, would have a density of 865 kg/m$^3$, which filling 30% pore space would result in a density change of 259 kg/m$^3$. A density change of 156 kg/m$^3$ 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/m$^3$ intruding into the fault zone as a dyke 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 dyke intrusion would cause significant additional deformation (Appendix B) 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 Mével 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 dyke-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. Non-welded ignimbrites and auto-breciated 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 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 3.7C), there is a weak gravity low (minimum -37 ± 26 µGal) 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 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.

3.5.2 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 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 $3 \times 10^{-6}$/year 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 modelled 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 a sill intruding 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 towards 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 halfspace and the Coulomb failure stress 
(CFS) is given by:

\[ \Delta \sigma_f = \Delta \tau_s + \mu' \Delta \sigma_n \]  (3.3)

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 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, \( \Delta \sigma \) is calculated as :

\[ \Delta \sigma = \frac{\Delta \sigma_{xx} + \Delta \sigma_{yy} + \Delta \sigma_{zz}}{3} \]  (3.4)

The model of Miller et al. (2017) of a high melt percentage magma reservoir suggests an isotropic elastic halfspace, assumed by the Coulomb code, may not be realistic. However, the halfspace 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 N20E, dipping 18° to the east, and opening at 1 m/year in accordance with the model of Feigl et al. (2014). We model the stress changes of this sill opening on the Troncoso fault, striking N40E, a representative lake fault determined from seismic reflection (Peterson et al., 2016), striking N50E and a fault mapped on the north shore of the lake, striking N61E (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.
Figure 3.8: Plan sections of (A) normal, (B) Coulomb, and (C) mean stress at 1.5 km depth, from sill opening on receiver faults oriented at 40°, i.e., parallel to the Troncoso fault, dipping 70°E with a rake of -90°. Lake outline is shown in blue lines, simplified fault traces in black lines, and the deformation source of Feigl et al. (2014) as grey rectangle.

Positive stress change produces fault unclamping. D-F) Mean stress change on the receiver fault plane resulting from the sill opening. These faults dip at 70° to the SE with a rake of -90°, i.e., normal faults. The oval and rectangular blue shaded area represents the region of mass increase from microgravity models over all time intervals.
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 3.8A - C), 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. Secondly, we calculate the mean stress on the three fault planes, (Figures 3.8D - F) 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 3.9). 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 x 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/m³ which results in a strain-induced density change of -0.04 kg/m³. This value is negligible and can be ignored as the 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 x 10^8 m³), a volume increase due to strain, of 3.82 x 10^3 m³ is created. Calculating in this way, between 2014 and 2015, 1.16 x 10^3 m³ of new volume is created and between 2015 and 2016, 5.03 x 10^2 m³ of new volume is created. The prism gravity source volume is much larger than that produced by volumetric strain, suggesting 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 poro-elastic finite element models (e.g. Strehlow et al., 2015) are required to fully assess the
Figure 3.9: 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 A) 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.
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 centre 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 re-pressurises 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 3.2D). 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 the on-going 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.

3.5.3 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:

\[
K = \frac{-Q}{A} \left[ \frac{\delta l}{\delta h} \right]
\]

(3.5)

where Q is the volumetric flow rate (m³/s), derived from the gravity models, A (m²) 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 3.3), \(\delta l\) is taken as the prism thickness. We assume fluid infiltrates in the direction of normal stress, along the plane of the fault. For \(\delta h\), we use the normal stress change (10 bar) from the fault opening (Figure 3.8) converted to head (\(\sim 101\) m). The pressure change across the subvertical fault means we are calculating the horizontal conductivity in the pressure change direction, \(K_{hx}\), rather than the vertical (i.e., gravity driven, \(K_z\)) conductivity. The fault zone is likely to be anisotropic so \(K_{hy}\) does not equal \(K_{hx}\). We use a fluid of density 1000 kg/m³ 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 volumetric flow rate ($Q$) of $4.9 \text{ m}^3/\text{s}$. Between 2014 and 2015 surveys, $Q = 1.1 \text{ m}^3/\text{s}$ from a mass change of $3.85 \times 10^{10}$ kg in 413 days, while between 2015 and 2016, $Q = 1.7 \text{ m}^3/\text{s}$ 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 $K_x$, $K_x = 6.65 \times 10^{-6}$, $9.92 \times 10^{-7}$, and $8.10 \times 10^{-7}$ m/s for the 2013 to 2014, 2014 to 2015, and 2015 to 2016 intervals, respectively.

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 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 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 we calculate are typical of volcanic material observed at similar volcanic centres.

The relationship between intrinsic permeability ($k$, in m$^2$) and hydraulic conductivity ($K$) was defined by Hubbert (1956) as:

$$K = k \frac{\rho_w g}{\mu}$$

(3.6)

where $\rho_w$ is the density of water (1000 kg/m$^3$), $g$ is gravitational acceleration (9.8 m/s$^2$) and $\mu$ is dynamic viscosity of water (1 x $10^{-3}$ kg/m.s at 20 °C). As we do not know the true temperature and hence density or dynamic viscosity of the water in the fault zone, the parameters we calculate are likely to be maximums. Solving for $k$, we obtain intrinsic permeabilities of $6.54 \times 10^{-11}$ m$^2$ for the 2013 to 2014 interval, $9.75 \times 10^{-12}$ m$^2$ between 2014 and 2015, and $7.96 \times 10^{-12}$ m$^2$ between 2015 and 2016, for our range of $K_x$ perpendicular to the fault plane.

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.
3.6 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 the mass addition is confined to a single geological layer, and is not associated with deep rooted dyke 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, modelled 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 volumetric flow rate within each time interval, we calculate hydraulic conductivity values perpendicular to the fault plane of $6 \times 10^{-6}$ to $9 \times 10^{-7}$ m/s. Permeabilities derived from the hydraulic conductivity are on the order of $6 \times 10^{-11}$ to $9 \times 10^{-12}$ 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 localisation 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 characterise their geometries, history and evolution.

3.7 Acknowledgments

We thank reviewers Jim Kauahikaua and Elske de Zeeuw-van Dalfsen for comments and suggestions that improved this manuscript. We also thank Michael Poland and Paul Tregoning for editorial handling and additional comment. Many thanks to the following for assisting with data collection; 2013 UW students, and Nathan Andersen, Alex DeMets, and Swetha Venugopal. We thank OVDAS staff, Maria Loreto Cordova, Carlos Cardona, Sergio Morales, and Diego Lillo, for CGPS and earthquake data, and ongoing field and logistical assistance. Thanks to Ministry of Public Works of Chile for the lake level data and to Sylvain Bonvalot and BGI for the absolute gravity measurements. We appreciate the hospitality
and logistical assistance of Don Luis Torres, and his help keeping us afloat on Laguna del Maule. Thanks to Jeff Witter for commenting on an early version of the manuscript and to Jeff Zurek for discussions on the source models. Field work was funded by NSF Integrated Earth Systems grant EAR-1411779 and EAR-1322595. C.M is supported by GNS Science Core Funding, EQC New Zealand, Mitacs Accelerate Canada, and Mira Geoscience. Gravity data are available in the supporting information.

3.8 References


Helsel, D., and R. Hirsch 1992, Statistical methods in water resources, Techniques of Water-
1269385.

Hildreth, W., E. Godoy, J. Fierstein, and B. S. Singer 2010, Laguna del Maule volcanic
field: Eruptive history of a Quaternary basalt to rhyolite distributed volcanic field on the
Andean range crest in central Chile., Servicio Nacional de Geologia y Mineria, Bolletin,
63, 145.

Hubbert, M. K. 1956, Darcy’s law and the field equations of the flow of underground fluids.,

Hummel, N., and S. A. Shapiro 2012, Microseismic estimates of hydraulic diffusivity in case
of non-linear fluid-rock interaction, Geophysical Journal International, 188(3), 1441–1453,

Johnson, D. J., A. A. Eggers, M. Bagnardi, M. Battaglia, M. P. Poland, and A. Miklius
2010, Shallow magma accumulation at Kilauea Volcano, Hawai’i, revealed by microgravity

Jónsson, S. 2009, Stress interaction between magma accumulation and trapdoor faulting on
2008.08.005.

Le Mével, H., K. L. Feigl, L. Córdova, C. DeMets, and P. Lundgren 2015, Evolution of unrest
at Laguna del Maule volcanic field (Chile) from InSAR and GPS measurements, 2003 to

Le Mével, H., P. M. Gregg, and K. L. Feigl 2016, Magma injection into long-lived reservoir
to explain geodetically measured uplift: application to the 2004–2015 episode at Laguna
del Maule volcanic field, Chile, Journal of Geophysical Research: Solid Earth, 121(8),

Lin, J. 2004, Stress triggering in thrust and subduction earthquakes and stress interaction
between the southern San Andreas and nearby thrust and strike-slip faults, Journal of

Manga, M., I. Beresnev, E. E. Brodsky, J. E. Elkhoury, D. Elsworth, S. E. Ingebritsen,
D. C. Mays, and C. Y. Wang 2012, Changes in permeability caused by transient stresses:
Field observations, experiments, and mechanisms, Reviews of Geophysics, 50(2), doi:

changes and CO₂ degassing at Laguna del Maule, Chile, accompanying rapid uplift., in
AGU Fall Meeting.

83

Peterson, D., C. Miller, N. Garibaldi, K. Keranen, B. Tikoff, and G. Williams-Jones 2016, Magma-tectonic Interaction at Laguna del Maule, Chile, in AGU Fall meeting, AGU, AGU, San Francisco.


This chapter presents the second case study on the architecture of volcanic systems, as modelled from gravity and magnetic data collected at Mt Tongariro, New Zealand. The results of this study address the third thesis objective in Table 1.1, ‘Define Mt Tongariro volcano and basement structure and hydrothermal system extent’, by answering questions on the shape of the basement, the influence of basement structures on vent locations, the extent of the hydrothermal system, and hazards associated with these features. This chapter also provides the context needed to understand the response of the hydrothermal system to the 2012 eruptions, as discussed in Chapter 5. In this chapter the following hypotheses are posed:

- Greywacke basement beneath Mt Tongariro is continuous.

- A broad region of hydrothermally altered rock underlies the TgVM, and thus is a site for potential phreatic eruption in the future.

- Fault movement can explain the magnitude of subsidence in the Ruapehu graben since 275 ka. Crustal flexure is not needed to accommodate the amount of observed subsidence in this region.

- The Waihi and Poutu basement faults do not act as magma ascent pathways in the TgVM.

- Magma rises to the surface in the TgVM in small, discrete batches with little to no shallow storage within the edifice.

- The volume of erupted material is larger than previous estimates.

Chapter 4 is published in the Journal of Volcanology and Geothermal Research, by Craig Miller and Glyn Williams-Jones. The journal formatted article is available at http://doi.org/10.1016/j.jvolgeores.2016.03.012. CM collected the new gravity dataset, processed, modelled and interpreted the data, and wrote the manuscript. GWJ commented on the manuscript before submission. Additional material is presented in Appendix D.
Chapter 4

Internal Structure and Volcanic Hazard Potential of Mt Tongariro, New Zealand, from 3D Gravity and Magnetic Models

Abstract

A new 3D geophysical model of the Mt Tongariro volcanic massif (TgVM), New Zealand, provides a high resolution view of the volcano’s internal structure and hydrothermal system, from which we derive implications for volcanic hazards. Geologically constrained 3D inversions of potential field data provide a greater level of insight into the volcanic structure than is possible from unconstrained models. A complex region of gravity highs and lows (±6 mGal) is set within a broader, ∼20 mGal gravity low. A magnetic high (1300 nT) is associated with Mt Ngauruhoe, while a substantial, thick, demagnetised area occurs to the north, coincident with a gravity low and interpreted as representing the hydrothermal system. The hydrothermal system is constrained to the west by major faults, interpreted as an impermeable barrier to fluid migration and extends to basement depth. These faults are considered low probability areas for future eruption sites, as there is little to indicate they have acted as magmatic pathways. Where the hydrothermal system coincides with steep topographic slopes, an increased likelihood of landslides is present and the newly delineated hydrothermal system maps the area most likely to have phreatic eruptions. Such eruptions, while small on a global scale, are important hazards at the TgVM as it is a popular hiking area with hundreds of visitors per day in close proximity to eruption sites. The model shows that the volume of volcanic material erupted over the lifespan of the TgVM is five to six times greater than previous estimates, suggesting a higher rate of magma supply, in line with global rates of andesite production. We suggest our model of physical property distribution can be used to provide constraints for other models of dynamic geophysical processes occurring at the TgVM.
4.1 Introduction

Knowledge of a volcano’s internal structure is important for many aspects of volcanology and volcanic hazard assessment. This is especially so in complex multi-vent systems where there is no central vent through which most eruptions occur and where multiple vents have been active in historic times. By geophysically imaging the volcano plumbing system and structures in the basement below the volcanic edifice, it is possible to assess the importance of these structures in controlling magma ascent paths and vent locations. In addition, knowledge of the extent of a volcano’s hydrothermal system provides important information on the likely style of eruptions. Hydrothermal systems often manifest as scenic surface features, attracting hikers and tourists, but when over-pressurised can produce small, but dangerous phreatic eruptions with very little warning (e.g., Raoul Island, Christenson et al. (2007); Te Maari, Procter et al. (2014); Ontake, Sano et al. (2015)) and are often overlooked in volcanic hazard assessments. As such, knowledge of the extent of a hydrothermal system and its interaction with magma pathways provides important information on the likelihood of such eruptions and allows suitable hazard mitigation to be put in place (Potter et al., 2014). Long-lived hydrothermal systems considerably alter and mechanically weaken large volumes of rock, which if coincident with steep slopes, presents a considerable landslide, lahar and flank collapse hazard (e.g., López and Williams, 1993; Day, 1996; Finn et al., 2001; Reid et al., 2002; Moon et al., 2005; Tontini et al., 2013).

Geophysical knowledge of a volcano’s internal physical property distribution also provides context within which processes that occur during volcanic unrest can be interpreted. Often, geophysical models of volcano unrest are limited by use of an unrealistic uniform halfspace or simple 1D model: the necessary geophysical context required for more detailed modelling is unknown (Cannavò et al., 2015). This results in inaccurate models which impedes scientists’ ability to make informed decisions during times of volcanic unrest.

Here we present a new, detailed, 3D geophysical model of the multi-vent Mt Tongariro volcanic massif (TgVM), New Zealand, combining an extensive new gravity dataset with aeromagnetic and geological data. We use a geologically constrained inverse modelling technique not previously applied to complex multi-vent andesite stratovolcanoes (c.f. Blaikie et al. (2014)), to produce a geologically sound and geophysically accurate model of the TgVM. This model enables examination of 1) the basement surface and faulting under the edifice, 2) the bulk internal structure of the volcano and 3) the extent of the hydrothermal system. Furthermore, we assess the volcanic hazard implications of features in our model. For example the distribution of hydrothermally altered rock has an influence on future landslide potential and we consider the likelihood of basement faults acting as future magma pathways.
4.2 Geologic Setting and Existing Geophysical Data

Interest in the TgVM has increased since early 2000 when unusual tornillo-type earthquakes were detected (Hagerty and Benites, 2003) around the Te Maari craters (Figure 4.1). In 2005–2009, a long sequence of small volcanic earthquakes occurred close to Mt Ngauruhoe (Jolly et al., 2012), 30 years after its last eruption, and in 2012 two eruptions occurred from the Upper Te Maari Crater, the first confirmed eruptions from this vent in over 100 years (Scott and Potter, 2014).

The TgVM lies at the southern end of the Taupo Volcanic Zone (TVZ), in a back-arc setting resulting from the westward subduction of the Pacific plate beneath the North Island of New Zealand. Within the back-arc setting is an extensional environment known as the Taupo or Ruanumoko Rift (Rowland and Sibson, 2001; Acocella et al., 2003). Extension across this rift is accommodated by segments or domains of sub-parallel north-west and south-east dipping normal faults (Seebeck et al., 2014). The TgVM is located at the northern end of the Ruapehu or Tongariro domain, an area dominated by andesitic volcanism, south of the dominantly rhyolitic Taupo domain. The geologic extension rate across the graben (Mt Ruapehu graben) formed by normal faulting in the Tongariro domain is estimated by Villamor and Berryman (2006b) to be 2.3 ± 1.2 mm/year. Several sub-parallel faults and fault zones delineate the graben in our study area; from west to east, these are the National Park fault, Waihi fault zone, Poutu fault zone and the inferred location of the northwest dipping Rangipo fault (Figure 4.1).

Jurassic age basement rocks of the Torlesse Terrane outcrop in the Kaimanawa Ranges on the east, while Waipapa Terrane rocks outcrop in the far west of the model area. In the centre of the Mt Ruapehu graben, basement rocks are inferred to be overlain by a thin layer (100 m) of Tertiary sediments. Tunnels drilled as part of the Tongariro power scheme in the far north-west of the study area intersected Waipapa Terrane greywacke beneath surface Tertiary sediments at a depth of around 100 m (Beetham and Watters, 1985).

Here we refer to the TgVM as the various eruptive centres that make up Mt Tongariro, including Mt Ngauruhoe (2280 m). The TgVM is constructed of at least 17 overlapping vents built during 6 main cone building episodes and covers an area of 5 by 13 km (Hobden et al., 1999). The massif has been extensively modified by glaciation since the first eruptions around 275 ka, thus surface exposures of early vents are obscured by later eruptions or have been removed by erosion.

Earliest activity began in the area of Lower Tama Lake (Tama 1) (Figure 4.1), followed by activity around 200 ka at a nearby centre, Tama 2. A long lived cone north of Oturere Valley (Northeastern Oturere, Mangahouhoumi lavas) was built between 105–130 ka during which time a vent near Pukekaikore was also active. Another centre, Tongariro Trig, formed between 65–110 ka, while contemporaneously a cone formed to the south of Oturere Valley (Southwestern Oturere, Waihohonu lavas) (Hobden et al., 1996). Around 25 ka, activity
Figure 4.1: Simplified geological map of the TgVM. The inset map shows the North Island of New Zealand with the TVZ and model area outlined in black. The dashed red line is the Pacific/Australian plate boundary. Coordinates are easting and northing in m using the NZTM projection.
started at Te Maari, Tama Lakes, Red Crater, North Crater, Blue Lake and Pukekaikiore (Nairn, 2000). Since around 7 ka, activity has been dominated by the growth of Mt Ngauruhoe cone (Moebis et al., 2011), while historic activity has been from Mt Ngauruhoe, Red Crater and Upper Te Maari (Scott and Potter, 2014).

Flank collapse has punctuated cone building episodes, either triggered by eruptions (Lecointre et al., 2002), or triggering eruptions by rapidly de-pressurising the hydrothermal system (Jolly et al., 2014). In both cases, the active hydrothermal system played an important role in mechanically weakening the rock prior to failure (Breard et al., 2014). Currently the largest surface hydrothermal features are at Red Crater, Ketetahi and Te Maari craters, although other mapped areas of alteration suggest a long history of hydrothermal activity in many locations on the massif. What is not documented from surface mapping is how extensive alteration is within the massif.

4.2.1 Previous geophysical studies

Previous geophysical studies have imaged the structure, magmatic and hydrothermal systems of the TgVM at varying degrees of resolution. Zeng and Ingham (1993) undertook two dimensional modeling of sparse gravity data along a profile south of Tama Lakes and suggested the presence of low density pyroclastic material overlying a dense basement. Walsh et al. (1998) summarised electrical resistivity data to delineate the extent of the hydrothermal system along a single profile; they found a shallow low resistivity layer, interpreted as geothermal condensate several hundred metres thick, overlying a vapour-dominated layer of unknown thickness. Rowlands et al. (2005) undertook a moderate resolution seismic tomography study and identified significant low velocity anomalies beneath Mts Ruapehu, Tongariro and Ngauruhoe which they interpreted as remnant magma batches and thick pyroclastic material from various volcanic sources. Cassidy et al. (2009) produced a more detailed 2D model across the Tama Lakes profile, from new gravity, aeromagnetic and magnetotelluric (MT) data. They also modelled the basement structure and inferred that the Waihi faults were pathways for magma intrusion into the TgVM. Johnson et al. (2011) and Johnson and Savage (2012) used seismic anisotropy measurements to map spatial and temporal changes in anisotropy; in particular they found a strong change in anisotropy north of Mt Ngauruhoe which they associated with the TgVM hydrothermal system. Hill et al. (2015) undertook a detailed 3D MT survey and found evidence for both shallow and deep conductive zones, interpreted as magma ascent pathways and deeper storage zones. In particular, they found a narrow (1 km) vertical conductive zone under Mt Ngauruhoe interpreted to represent the ascent path of magmatic fluids from a source at 4–12 km depth.

4.2.2 Physical property measurements

The TgVM consists of a variety of rock types including alternating layers of highly vesicular scoria and dense lavas, underlain by dense meta-sediments. To constrain the physical prop-
erties of different rock units for modelling, we extracted a dataset of 176 samples from the GNS Science PetLab database (http://pet.gns.cri.nz) from rocks on and around the TgVM. Physical properties include wet and dry density and magnetic susceptibility. We also incorporated physical property measurements from several studies of individual vents of the TgVM; including Tongariro (Hackett, 1985), North Crater (Griffin, 2007), Mt Ngauruhoe, (Krippner, 2009; Sanders, 2010), and Blue Lake (Simons, 2014), for a total of 288 measurements. For analysis purposes, we grouped samples into four main rock types: Andesite lava, referring to dense lava flows; Pyroclastic, a range of material from pumice and scoria to denser welded agglomerates; Greywacke, referring to basement Torlesse and Waipapa Terrane rocks; and Sandstone, Tertiary sandstones from the Taumarunui Formation.

For each rock type, we computed physical property histograms (Figure D.3) with the mean and standard deviation for each (Table 4.1). Wet densities better represent whole rock densities for rocks that are below the water table and are more suitable for gravity modelling. Depending on the porosity of the rock, dry vs wet densities in these samples can vary by as much as 340 kg/m³.

Magnetic susceptibilities of fresh volcanic rock samples range from 0.001 to 0.04 SI, while basement greywacke rocks are only very weakly magnetic (<0.001 SI) or non magnetic (below detection limit). Hunt and Mumme (1986) showed magnetisation intensities of young Mt Ngauruhoe lavas range from 0.7 to 49 A/m from unweathered samples. No samples of hydrothermally altered rock were available in the database and measurements of magnetic susceptibility in volcanic rocks may be dominated by remnant magnetisation, so are only used as a guide.

4.3 Geophysical Data Acquisition and Processing

Our study covers 504 km² within a rectangle 28 km x 18 km ranging in elevation from 600 m to 2300 m. This region encompasses all the lava flows from the TgVM and includes basement rocks outcropping to the east and west of the volcano.

4.3.1 Gravity survey design

We collected gravity data along radial traverses on foot, from the summit of the volcano massif (~2200 m) to the tree line (~1100 m) where thick vegetation prevented further surveying. This results in a 2–3 km wide region with no coverage from 1100 m to 700 m. Below 700 m, surveying resumed along the roads at the base of the volcano. The area with no coverage consists mostly of distal lava flows and pyroclastic deposits. Station spacing along the traverses is 500 m and traverses were located approximately 1 km apart; spacing reflects a trade off between completing coverage of the entire volcano and resolution of structures in the volcano. At 500 m station spacing, Nyquist theorem indicates we will be able to resolve features with a wavelength of >1000 m which is considered adequate for
<table>
<thead>
<tr>
<th>Rock Type</th>
<th>Dry density mean (kg/m³)</th>
<th>Dry density stddev (kg/m³)</th>
<th>Wet density mean (kg/m³)</th>
<th>Wet density stddev (kg/m³)</th>
<th>Magnetic susceptibility mean (SI)</th>
<th>Magnetic susceptibility stddev (SI)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Andesite Lava</td>
<td>2384 (n=132)</td>
<td>302</td>
<td>2535 (n=108)</td>
<td>205</td>
<td>0.022 (n=94)</td>
<td>0.017</td>
</tr>
<tr>
<td>Pyroclastic</td>
<td>1591 (n=143)</td>
<td>569</td>
<td>1931 (n=143)</td>
<td>267</td>
<td>0.009 (n=46)</td>
<td>0.010</td>
</tr>
<tr>
<td>All Volcanic</td>
<td>1971 (n=275)</td>
<td>607</td>
<td>2334 (n=251)</td>
<td>364</td>
<td>0.017 (n=140)</td>
<td>0.016</td>
</tr>
<tr>
<td>Greywacke</td>
<td>2706 (n=31)</td>
<td>56</td>
<td>2727 (n=30)</td>
<td>45</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Sandstone</td>
<td>2407 (n=3)</td>
<td>6</td>
<td>2517 (n=3)</td>
<td>6</td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

Table 4.1: Summary of physical properties for rock types within the TgVM. All Volcanic includes all Andesite Lava and Pyroclastic samples. Number of samples of each rock type is given by n.
the size of the volcano, but we would not be able to resolve small scale features such as individual feeder dykes as observed at Red Crater (Wadsworth et al., 2015). However, we can resolve large scale fault offsets and bulk rock physical property distributions related to different parts of the volcano and hydrothermal system. The station density across the survey area is 0.8 stations/km\(^2\) which increases to 1.4 stations/km\(^2\) on the upper flanks of the volcano.

### 4.3.2 Gravity datasets

Our gravity dataset contains data from three sources. The oldest data, sourced from the GNS Science New Zealand gravity station database ([http://gns.cri.nz/Home/Products/Databases/New-Zealand-Gravity-Station-Network](http://gns.cri.nz/Home/Products/Databases/New-Zealand-Gravity-Station-Network)), provides absolute gravity values referenced to the IGSN71 gravity datum (Morelli et al., 1974). These data provide far field, regional coverage to a distance of 40 km from the TgVM. Many of the 576 stations selected date from the 1960s, ‘70s and ‘80s and were located with accuracies of 100 m horizontally and >5 m vertically. Repeat occupation of a subset of these stations during the current survey reveals no systematic offset between current and old readings with most stations being repeatable to within 0.2 mGal. As a control on data quality, we excluded GNS Science data with measured elevations that are grossly different (10s m) from a 10 m Digital Elevation Model (DEM). These mostly occur in areas of steep terrain to the east of the study area where the poor horizontal positioning results in a large elevation difference.

The second dataset comprises 66 stations on the massif from Cassidy and Locke (1995) and Cassidy et al. (2009). These data were located using a mixture of barometry, precise levelling and differential GPS, hence height accuracies vary from ~5 m to <0.5 m.

The third dataset is new data we collected at 315 stations over 2 field campaigns in 2014 and 2015. Vertical and horizontal positions were determined using differential GNSS (using a Trimble Geoexplorer XH), operating in rapid static mode with 2 minute occupations and post processed with Trimble Pathfinder Office software using nearby GeoNet CGPS stations as reference stations. The short baselines between rover and base stations (<10 km) allows gravity stations to be located with vertical accuracies of better than 0.2 m.

To combine the data from the 3 surveys, we reoccupied the primary base station from the Cassidy survey and tied it to our newly established local base station. Repeat measurement of a selection of the Cassidy stations showed values agreed within 0.12 mGal. Finally, we tied our local base station back to the National Park reference station (GNS station ID 96), so that both our and the Cassidy stations were assigned an absolute gravity value, consistent with the GNS dataset.

### 4.3.3 Gravity data reduction and errors

We corrected the raw data from the 2014 and 2015 surveys for Earth tide, ocean loading and drift (e.g., Battaglia et al., 2012) to produce data relative to our local base. Average base
station loop closure errors after Earth tide and ocean loading are accounted for are 0.02 ± 0.03 mGal. We applied a correction scheme following that outlined in Hinze et al. (2005), across all generations of data, to compute a consistent dataset. Details of the correction scheme are in Appendix D.

Estimating the overall error in gravity values from closure, height, positioning and terrain errors gives a RMS value of 0.070 mGal for the 2014 and 2015 surveys. The Cassidy survey data have errors from 0.1 to 1 mGal, while errors from the GNS Science dataset are up to ~1 mGal, mostly due to the poor accuracy of the height determination.

One of the most important choices in gravity data reduction is the selection of the reduction density applied to the Bouguer and terrain corrections. The shape and amplitude of the resulting complete Bouguer anomaly can vary with choice of reduction density which directly influences the resulting models. Methods such as Nettleton (Nettleton, 1939), Parasnis (Parasnis, 1966) or their derivatives (Gottsmann et al., 2008), are often not valid in heterogeneous volcanic rock environments so we use our physical property dataset instead. See Appendix D for further discussion on the calculation of correction density from gravity measurements. We chose the mean volcanic rock wet density value of 2334 kg/m$^3$ (rounded down to 2300 kg/m$^3$), to represent the bulk density of the volcano massif, for computing the complete Bouguer anomaly. This density is valid for the mass above the reduction datum, i.e., the ellipsoid. Our study area contains a wide variety of volcanic and basement rock types above this datum, so finding a single density suitable for all rock types is not possible and may have resulted in parts of the dataset being over or under corrected. However we consider our chosen correction density to be in the middle of the range of all rock types, thus any error caused by over or underestimation of the correction density should be evenly distributed around the chosen value and not overly bias the results.

### 4.3.4 Complete Bouguer anomaly

We computed the complete Bouguer anomaly (CBA) on the dataset of 957 stations, covering an area of 70 km by 80 km in order to accurately determine the regional gravity in the area of interest around the volcanoes (Figure 4.2A). The regional CBA ranges from +41 mGal in the northwest to -54 mGal in the south and broadly consists of two gravity highs to the northwest and east. These highs correlate with mapped areas of outcropping basement Torlesse Terrane in the east and the Waipapa Terrane in the west (Figure 4.1). Between these highs is a broad gravity low defining the width of the Taupo Volcanic Zone. The TgVM is situated at a local maximum in this gravity low which decreases further to the north-east and also to the south, towards Mt Ruapehu. The strike of the gravity signal changes from NE - SW to E - W just to the south of Mt Ruapehu, representing the termination of the TVZ.

In the modelling area (dashed box in Figure 4.2A), the CBA shows a broad, asymmetric ‘U-shaped’ trend, descending steeply from a gravity high (28 mGal) in the north-west to a
Figure 4.2: Complete Bouguer anomaly data for A) Regional area around Mt Tongariro, contour interval 5 mGal. The detailed 28 x 18 km model area is shown in the black dashed rectangle. Black dots are GNS Science stations, blue dots are Cassidy stations, red dots are stations collected in this study. B) The residual CBA in the model area after removal of a 3rd order polynomial, contour interval 2 mGal. Vent locations are shown in white triangles and stations as for part A. C) The residual CBA low pass filtered to 10,000 m wavelength, contour interval 2 mGal. Shown in all figures are the active faults (white lines). Coordinates are easting and northing in m using the NZTM projection.
broad low (-4 mGal) and then increasing gradually to another high (12 mGal) in the east (Figure 4.2B). The TgVM is located on the west side of the gravity low, and is characterised by short wavelength anomalies with localised maxima and minima (±6 mGal). Local minima are associated with the cones of Mt Ngauruhoe, Red Crater, Mt Tongariro summit and Te Maari as well as with the Upper Tama Lake crater. Local maxima are located in the Oturere Valley to the east of Red Crater and to the east of Blue Lake. A small gravity high is also located in the Mangatepopo Valley.

Modelling requires removal of a regional field that creates a long wavelength gradient across the anomaly map reflecting the broad crustal structure relating to the TVZ and the subduction zone to the east. Zeng and Ingham (1993) and Cassidy et al. (2009) fit the regional field in the Tongariro area using a third order polynomial calculated from stations located on outcropping basement rocks. The use of a third order polynomial is common in gravity studies throughout the TVZ (e.g., Stern, 1979; Stagpoole and Bibby, 1999; Caratori Tontini et al., 2015). We remove the same regional field from our data and the resulting residual anomaly map is shown in Figure 4.2B.

### 4.3.5 Aeromagnetic data acquisition and processing

Approximately 510 line kilometres of aeromagnetic data were acquired in February 1995, along 19 flight lines, at a nominal 500 m spacing, of which only the southern most line is published in Cassidy et al. (2009). A proton precession magnetometer was towed 100 m behind a Cessna fixed-wing aircraft and data were acquired at 2 second intervals which for an average flight speed of 100 knots resulted in 1 sample approximately every 100 m. The survey was flown at a constant altitude configuration with a mean altitude of 2450 m, although turbulence meant that flight altitude could vary by as much as 100 m above or below the mean, along each line. Flight lines were oriented NW - SE perpendicular to the main strike of the TVZ. A single tie line was flown NE - SE along the central axis of Mt Tongariro (Figure 4.3A). We did not level the survey using crossover points or the tie line due to difficulties in maintaining a constant altitude over the mountainous terrain.

A base station installed 20 km from the centre of the survey area provided diurnal corrections. To produce a total magnetic intensity (TMI) anomaly, we subtracted the International Geomagnetic Reference Field (IGRF) (Thébault et al., 2015) from each sample point. To remove spikes in the data, we implemented a smoothing filter using a zero phase, 11 sample Hanning window (Jones et al., 2001). We removed a weak regional trend (~ 2 nT/km) from the smoothed TMI data by subtracting a second order polynomial, creating a residual TMI dataset suitable for inversion. For all magnetic models we used an ambient field with intensity 55,458 nT, inclination -64.58° and declination 20.67° as calculated from the IGRF model for February 1995.
4.3.6 Residual total magnetic intensity anomaly

The residual Total Magnetic Intensity (TMI) anomaly (Figure 4.3A) shows an elongated ellipsoid of magnetic material, with the long axis oriented parallel to the regional strike of the TVZ. The high flight altitude, at up to 1 km above the topography, results in a relatively low resolution image but still offers enough detail to distinguish larger scale structures. The relatively young age of the TgVM means the volcanics were erupted within the current normally magnetised Brunhes epoch that began ∼0.78 Ma and as such, no reversely magnetised material is expected and magnetic lows imply loss of magnetisation. To aid qualitative interpretation, we computed the reduced to pole (RTP) anomaly (Figure 4.3B) to centre the anomalies over their causative body, but for inversion, the residual TMI data are used.

The amplitude of the residual TMI reaches a maximum of ∼1300 nT at Mt Ngauruhoe, a significant proportion of which is expected to be caused by the topographic effect of the high standing cone in relation to the flight altitude. In areas away from Mt Ngauruhoe, the field intensity averages around 250 nT and fades to 0 nT on the flanks of the volcano. The magnetisation reaches background levels while still over mapped lava exposures; however this may be a result of the high flight height reducing the sensitivity of the measurements on the lower flank. Within the high intensity zone in the centre of the volcanic massif is a 9 km² area of very low magnetisation. This area is centred between Red Crater and Blue Lake and
extends south to Mt Ngauruhoe and north to Upper Te Maari Crater, coincident with the hydrothermal surface features. A ridge of magnetic high extends south of Mt Ngauruhoe over Upper and Lower Tama Lakes, while a small magnetic low is observed around the area of the Ketetahi hot springs on the northwest flank of the volcano. A weak positive anomaly is seen to the north of the Pukeonake cones, associated with the lava field from those cones (Figure 4.3).

4.3.7 Spectral analysis

In order to investigate the internal structure of the TgVM, and its relationship to the basement, we apply a low pass filter to separate the potential field signals into long wavelengths representing the deeper basement and shorter wavelengths representing shallower volcanic material. To determine the optimal filter characteristics, we computed a radially averaged power spectrum from a 2D fourier transform of the residual CBA gravity data and TMI data using the GMT function, `grdfit` (Wessel and Smith, 2013). Depth to the top of the source can be calculated by decomposing the radially averaged power spectra into linear segments where the depth is proportional to the slope of line segments (Spector, 1970). For TMI data, the source depth is calculated from depth = -s/4π where s is the slope of the natural log spectral power (SP) vs wave number (k) graph. For gravity data, a correction term, 2*ln(k), is added to the ln(SP) to convert the gravity data to pseudomagnetic data (Hinze et al., 2005), before calculating the source depth using the same formula as for TMI data. We use an elevation of 1500 m for the average topographic height to convert gravity source depths to elevations and use 2450 m (flight height) as the reference for converting TMI source depths to elevations. Note, however, that the wavelength filtering is still based on the uncorrected power spectrum; only the depth calculation for gravity data requires the correction. The wavelength filter cut off was chosen at slope changes in the uncorrected power spectrum.

The corrected radially averaged gravity power spectra (Figure 4.4A) is decomposed into 2 linear segments each representing a different source depth within the data. The top of the shallowest source is equivalent to the topography of the volcanic material, while the top of a second source at around sea level likely corresponds to the top of the greywacke basement. In the uncorrected data three wavelength segments are seen: longer than 10000 m, 10000 m to 3300 m, and less than 3300 m. We applied low pass filters of 10000 m and 3300 m and compared the resulting grids. The 3300 m filtered grid contained short wavelength anomalies that matched those visible in the unfiltered dataset and are spatially related to known vents, so we consider these anomalies to be volcano related, rather than the basement. We therefore applied a low pass Butterworth filter with a cut off of 10000 m to the gravity data in order to model the basement beneath the TgVM as shown in Figure 4.2C.

The radially averaged power spectrum of the residual TMI data is shown in Figure 4.4B. The deepest layer is around 50 m below sea level and likely represents the base of
Figure 4.4: A) Radially averaged power spectrum of complete Bouguer anomaly data. Both corrected (dots) and non-corrected data (triangles) are shown. Top of source elevation estimates are based on the corrected data, while wavelength filter characteristics are based on the non-corrected data. Vertical lines highlight the wavelength segments tested in filtering. B) Radially averaged power spectrum of TMI data with top of source elevation estimates. Annotated line segments represent elevations of source layers.
volcanic material as the basement is non-magnetic. This interface corresponds well with a basement source depth calculated from the gravity data. Shallower layers represent the bulk of the volcanic material and a layer of surface lava probably associated with high elevation material on Mt Ngauruhoe.

### 4.4 Geological Modelling and Geophysical Inversion

We constructed a range of models, starting with simple unconstrained apparent property models, followed by unconstrained 3D inversions, and finally geologically constrained 3D inversions. The work-flow and subsequent interpretation based on building models of increasing complexity allows the full dynamic range of the dataset to be explored.

All modelling in this study uses GOCAD® Mining Suite (www.mirageoscience.com) to construct the starting 3D geological model which is directly coupled to the VPmg (Vertical Prism Magnetics Gravity) inversion routines (Fullagar and Pears, 2007; Fullagar et al., 2008), providing two-way interaction between geology and geophysical data. VPmg allows for a variety of inversion types: Homogeneous property inversion, to determine the optimal physical property (density or magnetic susceptibility) of a single geologic unit; Heterogeneous property inversion, to find the optimal physical property distribution within a unit. This includes an apparent property inversion, where the voxel (a 3D regular grid-set of voxels, or volume-pixels) consists of a single vertical prism extending the full depth of the voxel. The misfit of the apparent property inversion is useful for quickly assessing the degree of three dimensionality in the data. Finally, geometry inversion of geological contacts optimise the shape of a unit while its physical property remains constant. Each type of inversion can be applied sequentially and in combination. VPmg models the subsurface as a set of vertical rectangular prisms whose top surface matches the topography. Internally, the prisms are divided into cells with arbitrary vertical dimension. Cell subdivisions can be based on geologic units and each unit can be assigned homogeneous or heterogeneous physical properties. Heterogeneous units can be inverted in a smooth sense via least squares or stochastically. In stochastic inversion, random perturbations are chosen for each cell of each geologic unit. The size of the random perturbations is governed by the a priori defined property distribution and limited to three standard deviations from the mean property value of the unit being inverted. The perturbation is accepted if it reduces the chi-squared misfit, and is rejected otherwise (Fullagar and Pears, 2007). The RMS misfit (mGal) is computed and recorded, where

\[
RMS = \sqrt{\frac{1}{N} \sum_{n=1}^{N} (O_n - C_n)^2}
\]  

where \(N\) is the number of data, \(O_n\) is the measured data and \(C_n\) the calculated model response.

Mathematical details of the inversion method are provided in Appendix D.
4.4.1 Model initialisation

Voxet cell sizes are 250 m (half gravity station spacing) in the east and north axes and 100 m in the vertical (depth) axis. VPmg mathematically extends the model volume to 25 km depth to ensure complete modelling of the data at all wavelengths. The voxet is then embedded in a halfspace so that the model does not terminate abruptly, reducing edge effects. The physical property of the halfspace is optimised by the inversion routine.

We gridded the observed gravity and magnetic data at a 250 m cell size to ensure that each vertical prism in the voxet is associated with 1 data point located in the centre of the prism. VPmg requires the input of a topographic surface as the top surface of the model, so that the topography is modelled directly. This was constructed using a point dataset from an 8 m DEM (down sampled to 24 m) as constraints for fitting a smooth surface using the DSI interpolator in GOCAD. The resulting surface consists of a mesh of equilateral triangles with \( \sim 100 \) m sides. This topographic surface is then down sampled to the 250 m voxet for modelling.

We begin with a simple two layer case, consisting of a basement unit and a cover unit of volcanics; the complex geological history of the TgVM makes constructing a more detailed model highly subjective. Instead, we use the inversion process to discover detail within the subsurface and focus on building geological constraints into the model from kilometre scale features.

To construct the basement surface, we imported into GOCAD shape files of surface fault traces from the GNS Science New Zealand Active Fault Database and outlines of geological units from GNS Science Hawke’s Bay QMAP (Lee et al., 2011). We used the mapped contacts to accurately define the basement - volcanics boundary in our model. We have not explicitly modelled the thin (\( \sim 100 \) m) overlying layer of Tertiary sediments as they have a similar density to the average volcanic rock density and distinguishing the two without other constraints is difficult. Using the DEM and the basement geology contact curves, we warped a flat starting surface to the shape of the outcropping basement topography. We then overlaid the surface fault traces of the National Park, Waihi, Poutu and Rangipo fault zones. These fault zones are made from numerous sub-parallel strands and modelling each individual strand is outside the resolution of our gravity data. As such, only the major fault strands were included. We assigned initial offsets to these faults based on the model of Cassidy et al. (2009) and consistent with the mapped continuation of the faults outside the study area by Villamor and Berryman (2006b). We then warped the basement surface to fit the fault offsets. As a sensitivity test of our model to the starting geometry, we also created a flat basement surface that only included the outcropping basement, i.e., with no fault steps in it.
4.5 Results

We began our exploration of the data by first performing apparent property inversions to determine the broad lateral distribution of physical properties. We then performed an unconstrained 3D inversion to investigate the approximate vertical extent of anomalous features. These results (see Appendix D) highlight the necessity to better constrain the depth to the basement interface which from our knowledge of the geology and petrophysical contrasts suggests should be a sharp interface.

4.5.1 3D geologically constrained density inversion

To begin the geologically constrained model, we first used a geometry inversion to adjust the shape of the starting basement surface to fit the low pass filtered gravity data. In the geometry inversion, the basement density contrast is fixed at 400 kg/m$^3$ (to represent an absolute value of 2700 kg/m$^3$ matching our petrophysical data) and the top of the basement in each voxel cell is allowed to vary vertically. The resulting RMS misfit for this model is 1.7 mGal. The areas of worst misfit are associated with older GNS Science stations that may have errors up to 1 mGal. To test for any density variations in the basement which may improve the fit of the model, we performed a second inversion on the starting basement surface, comprising a combined geometry and heterogeneous property inversion. In this approach, alternating steps of a single heterogeneous density inversion iteration, followed by a single geometry inversion iteration are run to produce a model that accounts for both the geometry and density contrasts within the basement. This improved the misfit RMS to 0.75 mGal.

With the shape of the basement now constrained, we model the cover unit using a heterogeneous density inversion. In this model, the best fit basement geometry and physical property distribution, as described above, is fixed and the initially homogeneous cover unit is converted to a heterogeneous unit. We then perform both a conventional inversion and a stochastic inversion in the cover unit using the unfiltered, full wavelength, gravity data. In this way we are fitting the remainder of the gravity signal, not accounted for by the basement model, by density variations in the cover unit. The final conventional inversion model of the full dataset has an RMS of 0.9 mGal while the stochastic inversion produces a model with RMS of 0.93 mGal, with the highest misfits associated with gravity stations with poorer elevation control. This misfit is due to the combination of the fit of the basement surface plus the misfit due to the heterogeneous cover model. When the basement misfit is subtracted, the misfit due to the cover unit alone is 0.15 mGal.

We extracted a series of elevation slices from the conventional and stochastic inversions at 1350, 1150, 750 and 550 m a.s.l. (Figure 4.5). The conventional and stochastic inversions show similar results; however, the stochastic model appears noisier as each cell is treated independently of its neighbour so no smoothing occurs.
Figure 4.5: Depth slices from the conventional (left column B - E) and stochastic (right column G - J) geologically constrained gravity inversions at 1350, 1150, 750, 550 m a.s.l. Residual anomaly maps are shown in A and F. Overlain in black lines are active faults and vent locations as white triangles. Coordinates are easting and northing in m using the NZTM projection.
A general NE/SW trending fabric is evident, parallel to the main TVZ structural trend. A low density root occurs under Mt Ngauruhoe extending from surface to the basement and a small low density root occurs beneath the recent Tama Lakes crater. A complex shaped low density body occurs under Red Crater, Central Crater, North Crater and Blue Lake and appears to terminate approximately at the western most Waihi fault. This body, coincident with the known hydrothermal surface features, is divided into two sub-areas by a narrow region of higher density material. To the west of the massif are several surficial, low density anomalies which may be small pockets of pyroclastic material infilling previous topographic lows.

Small high density features exist from surface to a few hundred metres depth. The largest of these is associated with the Te Tatu lavas from the recent NE Oturere vent (Figure 4.5). A small area of high density material is modelled at the head of the Oturere Valley, just to the east of Red Crater. This is likely to represent a thick sequence of overlapping lava flows from Red Crater. Small pockets of dense rock occur north and south of Mt Ngauruhoe and may represent accumulation of lavas from Mt Ngauruhoe. Dense rock to the south of Mt Ngauruhoe may also be related to thick lavas from ancestral cones Tama 1 and Tama 2. An area of above average density rock occurs south of Pukekaikiore around the Waihi fault, while the Poutu fault zone sits within lower than average density rocks. High density rocks in the far northwest and southeast correlate with outcropping and shallow basement.

### 4.5.2 3D geologically constrained susceptibility inversion

To construct the magnetic model, we imported the basement unit from the best fit gravity model and assigned it a zero susceptibility, consistent with basement physical properties. We then performed conventional and stochastic unconstrained inversions on the cover layer. The RMS of the conventional inversion is 7 nT and shows a normal distribution of residuals indicating no bias in the model. The stochastic inversion RMS is 49 nT and the inversion did not successfully converge. We therefore do not use the stochastic inversion results further in our study.

Figure 4.6 shows depth slices from the conventional inversion. A large very low susceptibility area (<0.02 SI) is imaged in the shallowest slices and is coincident with the surface manifestations of the hydrothermal system and the low density body in the gravity model. It is elongated in a NE - SW direction, sub-parallel to the main faults and is divided into two sub-areas by a narrow ridge of moderate susceptibility rock which is not present in deeper slices. This area extends to basement and is bound to the west by the Waihi fault. A small shallow region of low susceptibility is associated with Ketetahi hot springs.

Areas of moderate magnetic susceptibility (0.05 SI) form the flanks of the TgVM coincident with flank lava flows. Areas of high magnetic susceptibility (>0.075 SI) occur at Mt Ngauruhoe and Tama Lakes. At Mt Ngauruhoe a thick high magnetic susceptibility area extends to around sea-level. This area has two side lobes, a thick one to the north-west,
Figure 4.6: Depth slices from the conventional geologically constrained TMI inversion at 1350, 1150, 750, 550 m a.s.l. (B - E). Residual anomaly map shown in A. Overlaid in black lines are active faults and vent locations as white triangles. Coordinates are easting and northing in m using the NZTM projection.
and one with less vertical extent to the south-east of the cone. These features extend from surface and likely represent a combination of the central conduit and flank lava flows from Mt Ngauruhoe. Two more high magnetic susceptibility units exist around the Tama Lake vents. These are 300–500 m thick, extend from surface, and are associated with both the young Tama vents, and with lavas from the older Tama 1 and 2 cones.

4.5.3 Limitations and sensitivity of geophysical models

Our ability to determine subsurface structure of the TgVM using gravity and magnetics fundamentally relies on there being sufficient contrast in the physical properties of rock types that make up the massif. At the TgVM there is a wide range of physical property values and the basement is non-magnetic providing a good contrast with the overlying volcanics. However, when we model the depth extent of the demagnetised volcanic area, the non-magnetic basement may not provide sufficient contrast to determine if demagnetised volcanic rocks extend into the basement.

The basement is also generally denser than the overlying volcanics; however, some of the less vesicular lavas have similar densities to greywacke, so if the basement is directly overlain by a thick pile of lava, accurately modelling the location of the basement contact will be more difficult. Similarly, density contrasts between solidified intrusions and greywacke are likely to be low (<100 kg/m$^3$) making them difficult to detect within the basement using gravity. However, they should provide good magnetic targets if they are of suitable size and have not been hydrothermally altered. Low density roots of magma conduits or low density hydrothermally altered rocks should have good contrast within the basement.

End members of the range of constrained and unconstrained models are shown in Appendix D.

4.6 Discussion

4.6.1 Basement structure

Our gravity model shows that the basement forms a continuous, but faulted, surface beneath the TgVM. The best fitting basement model includes a subtle E to W density gradient across the basement surface from 2670 kg/m$^3$ to 2730 kg/m$^3$ (Figure 4.7). This is within the range of physical property measurements and likely reflects the change from Torlesse Terrane to Waipapa Terrane, respectively. McNamara et al. (2014) noted a systematic difference in the mechanical behaviour of the Torlesse and Waipapa Terranes, where the Waipapa Terrane appears to be mechanically stronger. They attributed this difference to Waipapa greywacke having coarser grain size and a more mafic composition compared to the finer grained, felsic composition of the Torlesse. These attributes would also make the Waipapa Terrane rocks more dense than the Torlesse, which corroborates our model and justifies the inclusion of a
variable density basement. We are not able to distinguish any sharp boundary between the two terranes, under the TgVM.

The basement is mostly flat-lying with a slight dip (<5°) to the north and south away from the TgVM. Throws across the National Park, Waihi 1 and Waihi 2, and Poutu faults are modelled as ~50 m, ~70 m, ~200–300 m, and ~200 m, respectively. There is no discernible offset on the Rangipo fault in the model area which has decreased from ~800 m in the south. The throws on these faults are similar to those obtained by Cassidy et al. (2009); however they did not model the Poutu fault. While we have modelled the main fault strands, the Waihi and Poutu fault zones are made up of numerous sub-parallel strands which may accommodate the fault movement in a more complex way than can be resolved by our gravity model.

The gravity model shows that surfaces between the fault zones are gently dipping (<5°) towards the centre of the TgVM, indicating that each block of basement between the faults may be tilted as well as being faulted. The total basement subsidence beneath the TgVM, determined from the basement gravity model and taking into account discrete faulting and surface tilting, is 500–700 m.

Villamor and Berryman (2006a) infer initiation of faulting 554 ± 323 ka. If we assume that the onset of extension across the Ruapehu graben was synchronous with the onset of volcanism, then a long term subsidence rate from the gravity models of 1.8–2.5 mm/year
since 275 ka is calculated. This is similar to the $4 \pm 1 \text{ mm/year}$ of subsidence Villamor and Berryman (2006a) calculated for the Ruapehu graben to the south of the TgVM.

The general good agreement of geophysically-derived subsidence rates with those measured from fault outcrops shows that most, if not all, subsidence can be accounted for by fault movement. There is no need to invoke additional mechanisms for subsidence such as crustal flexing under the volcanic load, as has been observed for large stratovolcanoes erupted into weak sediments (e.g., Concepción and Maderas in Nicaragua; van Wyk de Vries and Borgia (1996)). The mechanical stiffness of greywacke is sufficient to support the weight of volcanic material and any unsupported load is taken up by fault movement. We are unable to determine the effect adding volcanic load has had on the rate of fault movement throughout the lifetime of the TgVM; however, our geophysically-derived subsidence rates are slightly less than the geologically observed rates so it is expected that loading has had little impact on subsidence rate which is largely a result of tectonic extension.

3D gravity and magnetic inversions did not resolve any discrete bodies within the basement, which is not surprising considering there is minimal density contrast between solidified andesite and greywacke and the size and depth of solidified magnetic intrusions may not be resolvable by the aeromagnetic survey. Conversely, it suggests that any partially molten, low density magma bodies are too small or too deep to be detected by our measurements. This is in agreement with the petrologic model of Hobden et al. (1999) that suggests small discrete magma batches are responsible for recent magmatic activity.

### 4.6.2 Volcanic edifice structure

We calculate a volume of $\sim 350 \text{ km}^3$ of volcanic material between the topographic surface and the basement. Most of this material is sourced from the TgVM; however, a small portion (estimated at $<50 \text{ km}^3$) will be sourced from outside the TgVM, from the Taupo Caldera and from nearby Mt Ruapehu. We calculate a long-term volumetric eruption rate of $1.3 \times 10^{-3} \text{ km}^3$ per year since initiation of volcanism at 275 ka. Hobden (1997) estimated a total volume of around $60 \text{ km}^3$ for eruptive products since the initial Tama 1 eruptions, erupting at an average rate of $0.17 \times 10^{-3} \text{ km}^3$ per year, while eruption rates for the Holocene formation of Mt Ngauruhoe are around $0.3 \times 10^{-3} \text{ km}^3$ per year. However, Nairn (2000) and Pardo et al. (2012) provide evidence for much larger eruption rates of the TgVM (Pahoka-Mangamate sequence of $6 \text{ km}^3$ in 200–400 years) and Mt Ruapehu (0.1 to $10 \text{ km}^3$ from individual Plinian eruptions since 27 ka), suggesting that the rate of magma supply has been variable over time.

The total eruptive volume and rate derived from our geophysical model is five to six times larger than those estimated from field observations and suggests a significant amount of material has been removed by erosion and glaciation, or that the volumes of older cones have been significantly underestimated because of poor surface exposure and a lack of knowledge about the true basement depth beneath the TgVM. These revised volumetric output rates compare well to the global average rate for andesite volcanoes of $2.3 \pm 0.8 \times$
10^{-3}\text{km}^3/\text{year} \ (\text{White et al., 2006})$, and imply a greater rate of magma production, albeit with a large amount of temporal variability.

Our model resolves some structures relating to the long term individual cone building episodes outlined by Hobden (1997), particularly the thick lavas associated with the Tama 1 and 2 centres, south of Mt Ngauruhoe. Features associated with younger volcanic vents (Figure 4.8A) include a small low density volume coincident with the young Pukekaikiore cone and similar volumes mapped at Tama Lakes, Half Cone and the young NE Oturere vent. These low density features likely represent the tops of vesicular magma feeder systems. The lack of high density anomalies coincident with these vents suggests that magma drainage from the feeder post-eruption, as seen at Red Crater (Wadsworth et al., 2015), may be common at other vents. This, in turn, suggests these vents were erupted at a low magma supply rate, likely originating from a small discrete magma batch. The only high density anomalies are associated with mapped thick lava flows around old Pukekaikiore, at the head of the Oturere Valley and in the head of the Mangahouhoumiui Valley.
A large low density and magnetic root (Figure 4.8B) extends to below basement depth under Mt Ngauruhoe which we attribute to a substantial magma feeder system. This cone shaped feature is geographically and geophysically consistent with the low velocity zone imaged by Rowlands et al. (2005) and the low resistivity zone imaged by Hill et al. (2015).

An extensive irregularly shaped low density area is mapped under the Red Crater, Central Crater and North Crater area, extending to Ketetahi hotsprings to the north. This feature is broadly coincident with a low magnetic susceptibility volume and is interpreted as altered rock from the hydrothermal system. Hydrothermal systems can produce either high or low density alteration depending on whether void space is mineralised by circulating fluids and thereby increasing density, or if the host rock is altered to clay minerals (Allis, 1990). As the hydrothermal system at the TgVM is vapour dominated (Hochstein, 1985), there is not likely to be large fluid circulation precipitating minerals into void space. Alteration of rocks by hot acid gases and condensate into lower density clay minerals is therefore the most likely mechanism for lowering density. Other low density regions on the flanks of the massif are shallow, have small vertical extent and are interpreted as variations in accumulations of tephra.

Our model provides limited support for the conclusion of Cassidy et al. (2009) that the basement faults acted as pathways for magma ascent. While some high density and magnetic material exists in conjunction with the southern end of the Waihi fault, this is also coincident with thick lavas from Pukekaikiore to the north of their profile. Our models do not image any thin, dense vertical structures extending into the basement. To be resolved by our surveys, feeder dykes would need to be a minimum of several hundred metres wide which is geologically implausible based on comparison with outcropping dykes and the amount of extension across the faults required to be accommodated by dyke infilling during a single eruptive episode. The other fault systems around TgVM show no evidence of dyke intrusion so it appears that while these faults are important tectonic features, they are not critical in determining the location of eruption sites. It may be that the faults are not zones of weakness but rather, are strongly coupled and resistant to dyke intrusion.

Neither the gravity nor the magnetic models imaged any large intrusive bodies within the edifice. This is in contrast to the dense dyke network imaged at the Pouakai and Kaitake volcanoes in Taranaki, to the west of the TVZ (Locke et al. (1993)). The extensive Taranaki volcano feeder systems suggest a fundamentally different magma supply regime compared to the TgVM and again corroborates petrophysical evidence that the TgVM vents are fed from small discrete magma batches that have risen from depth with little intermediate storage.

4.6.3 Hydrothermal system

We interpret the large magnetic low and the complex region of low density north of Mt Ngauruhoe extending to Upper Te Maari as representing the extent of TgVM hydrothermal system (Figure 4.9). The total volume of hydrothermally altered rock is around 20 km$^3$. 

111
Figure 4.9: The TgVM hydrothermal system as shown by iso-surfaces of low magnetic susceptibility (0.025 SI) in yellow, and low density (2250 kg/m$^3$) in cyan. The area outlined in red represents the hydrothermal system as shown in Figure 4.10. Perspective view looking from the east south-east. No vertical exaggeration.

This volume is smaller than the approximately 100 km$^3$ estimated by Caratori Tontini et al. (2010) for the hydrothermal system at Marsili volcano in the Tyrrhenian Sea, but is significantly larger than the 1.5–3 km$^3$ estimated by Finn et al. (2001, 2007) for the volumes of altered rock at Mt Rainier and Mt Adams, Washington. The low density and demagnetised area is broadly consistent with a region of strong seismic anisotropy change observed by Johnson and Savage (2012), which they attributed to increased fracturing associated with the hydrothermal reservoir. Such fracturing would help account for the observed gravity low, in addition to the formation of low density clay minerals as a by-product of hydrothermal alteration (Allis, 1990). Within this area are zones of more intensely demagnetised rock (0.001 SI). These areas are on the south flank of the Tongariro summit ridge, around Blue Lake and the slopes above Upper Te Maari crater and coincide with areas of considerable surface alteration. A smaller area of demagnetised rock is associated with Ketetahi hotsprings where there is an extensive area of surface alteration. Ballistic blocks from the 2012 Upper Te Maari eruption show varying degrees of alteration, from fresh to extensively altered (Breard et al., 2014), while mineral component analysis by Pardo et al. (2014) found unaltered magnetite phenocrysts within the ejecta. We interpret these observations as being consistent with Te Maari’s location on the edge of the hydrothermal system where a mixture of fresh and altered rock occurs.

As the basement is non-magnetic, it is difficult to determine how far into the basement the hydrothermal system extends solely on the basis of the magnetic anomaly. The geo-
logically constrained model suggests that the low magnetisation zone ends approximately 200 m above the basement surface, while the unconstrained model suggests it may continue to 2500 m below sea level (see Appendix D). The low density (2250 kg/m$^3$) region coincident with the low magnetisation area, extends to 500 m below the basement in the unconstrained model, suggesting that alteration might occur in the basement rock to a depth of a few hundred metres. This is not surprising given that the basement is likely to be highly fractured from repeat eruptions through it, allowing fluids to more easily circulate in otherwise impermeable rocks (cf. the basement-hosted Kawerau geothermal field, Milicich et al. (2013)). All models show that the hydrothermal system appears to truncate against the Waihi faults in the west. Faulted low permeability basement rocks may form a seal for the hydrothermal system to the west which agrees with the earlier observation that faults are tightly coupled and impermeable to fluid or magma injection. This may also explain why there are no outflow hotsprings or other thermal features outside these faults.

The southern boundary of the hydrothermal system ends on the north side of Mt Ngauruhoe. This appears unusual given that Mt Ngauruhoe has been the dominant cone building centre during the last 7 ka and would be expected to have a well developed hydrothermal system. Several explanations are possible. Firstly, the hydrothermal system does extend under Mt Ngauruhoe, but demagnetised areas are masked by the strongly magnetised historic lavas. Our modelling shows that strongly magnetised rocks exist to basement depth under Mt Ngauruhoe, so masking of altered rocks seems unlikely. Secondly, it is possible that either the present Mt Ngauruhoe cone, or remnants of older cones (Tama 1 and 2) act as a physical barrier to fluid movement. However, the current Mt Ngauruhoe cone is low density implying high porosity rocks that should be capable of supporting a hydrothermal system, and seismic evidence suggests hydrothermal fluid movement beneath the cone (Jolly et al., 2012). Alternatively, it may be that Mt Ngauruhoe is simply too young to have been sufficiently hydrothermally altered to produce a measurable demagnetisation. Estimating the rate of rock dissolution in hydrothermal systems is difficult as rates of chemical reactions that drive dissolution are highly dependent on temperature, surface area of exposed rock and fluid flux rates through the rock. Caratori Tontini et al. (2015) estimated a dissolution rate of 50,000 m$^3$/year for the Rotomahana geothermal field based on comparison of rates for other hydrothermal systems. For instance at Poás volcano, Rowe et al. (1992) calculated a rate of $\sim 1650$ m$^3$/year, and at White Island, Giggenbach (1987) estimated 22,000 m$^3$/year of rock dissolution. The topographic volume of the Mt Ngauruhoe cone is approximately 2.2 km$^3$ which at a dissolution rate of 50,000 m$^3$/year would require 44,000 years to demagnetise. Dissolution rates would need to be an order of magnitude higher, for sufficient alteration to have developed within the $\sim 7$ ka life span of the cone. Hence, while it is likely the hydrothermal system does extend under Mt Ngauruhoe, the rocks there are simply too young to have been sufficiently demagnetised to be imaged by aeromagnetic surveys. We can apply the same argument in reverse to conclude that the hydrothermal system producing
the large demagnetised area to the north of Mt Ngauruhoe must be long-lived, on the order of $10^4$ to $10^5$ years.

### 4.6.4 Implications for volcanic hazards inferred from geophysical models

Our geophysical models provide first order constraints on volcanic hazard potential at TgVM and offer some possibility for improvements in hazard monitoring. Use of simple 1D seismic velocity models in volcanic earthquake location algorithms can impact real-time hazard assessment in times of volcanic unrest if those models are oversimplified, resulting in incorrect hypocenter locations. We compare our 3D density model to the 1D velocity model calculated by Jolly et al. (2014) for an area on the north flanks of Te Maari. Converting P-wave velocity ($V_p$) to density (Brocher, 2005), we find excellent agreement below 1 km depth between densities in the 3D gravity model and the $V_p$ converted densities (Table 4.2). Our gravity model resolution is less sensitive to changes in the top 1 km which may account for the discrepancy in the upper layer. We could therefore reasonably convert our 3D density model to a volcano wide high resolution 3D velocity model for improved earthquake locations. This would be especially useful for high frequency earthquakes within the volcanic edifice (Hurst et al., 2014), with seismic wavelengths short enough to be influenced by a heterogeneous seismic velocity distribution.

<table>
<thead>
<tr>
<th>Depth (km)</th>
<th>1D $V_p$ (km/s)</th>
<th>Converted densities (kg/m$^3$)</th>
<th>Gravity model densities (kg/m$^3$)</th>
</tr>
</thead>
<tbody>
<tr>
<td>0</td>
<td>1.8</td>
<td>1810</td>
<td>2200</td>
</tr>
<tr>
<td>1</td>
<td>3.6</td>
<td>2330</td>
<td>2334</td>
</tr>
<tr>
<td>2</td>
<td>5.9</td>
<td>2690</td>
<td>2700</td>
</tr>
</tbody>
</table>

Table 4.2: P wave velocities from Jolly et al. (2014) converted to density using the relationships in Brocher (2005).

Our basement faulting model shows little evidence for the Waihi and Poutu faults acting as magmatic pathways and suggests that magma intrudes the basement between the faults, rather than along them. These fault structures are therefore considered low probability areas for future eruption locations.

The delineation of the extensive hydrothermal system identifies areas most at risk of phreatic eruption, either as a result of natural fluctuations in hydrothermal activity or those perturbed by magmatic intrusion. While phreatic eruptions are possible wherever there is interaction between magma and water, intrusions into the base of the hydrothermal system have greater ability to provide early warning of future eruption than intrusions into shallow groundwater aquifers (e.g., Hurst et al., 2014). The presence of large volumes of hydrothermally weakenend rock further promotes the chances of phreatic activity due to lowered confining strengths of these rocks (Heap et al., 2015). The large number of tourists...
that cross the TgVM on hiking trails each year, means that even small phreatic eruptions can present a high risk to those in close proximity.

The TgVM has a history of landslides and flank collapse. The largest example is the 0.5 km$^3$ Te Whaiau formation, formed by collapse of the northwestern flank around 55–60 ka (Lecointre et al., 2002), while the most recent example is the 2012 Te Maari eruption, initiated by a small landslide of ∼0.0007 km$^3$ (Procter et al., 2014). Both landslide deposits show extensive evidence of weak hydrothermally altered material being a contributing factor in the failure. The volume of alteration at the TgVM is considerably larger than that found at other studied andesite volcanoes and constitutes a considerable potential hazard. Our gravity and magnetic models suggest that highly altered surface zones surround a core of more moderately altered rock that extends to basement depths. Similarly, variably altered cores have been mapped at other andesite volcanoes (Finn et al., 2001; Mayer et al., 2015), and are recognised as potential sources of future landslides. As alteration extends to basement depths at the TgVM, the risk of large scale flank collapse is increased compared to volcanoes with only shallow alteration. Alteration at depth is susceptible to failure by mechanisms generated within the volcanic edifice, such as dyke intrusion or changes in pressurisation of the hydrothermal system, as well as surface based processes. In addition, once surface initiated flank collapse is under way, there is no unaltered core of competent rock to impede collapse retrogression, and limit the amount of material available to form debris flows.

To determine to a first order, landslide risk areas on the TgVM, we undertook a slope angle analysis of the DEM to identify steep slopes coincident with altered rock, which are likely to be susceptible to failure (Figure 4.10). Steep slopes (>30°), coincident with regions of hydrothermal alteration, are at greater risk of failure (Moon et al., 2005) due to the lower friction properties of altered rocks. Slope failure often depends on the level of the groundwater system, or on the internal fluid pore pressure, both of which can be raised through injection of fluids as a result of dyke intrusion, such as occurred in 2012. A suitable trigger mechanism may then be small local earthquakes or shaking from a larger regional event or increased rainfall adding extra gravitational load (Voight and Elsworth, 1997).
Figure 4.10: Slope angle (calculated on 15m DEM) and extent of demagnetised hydrothermal system (outlined in red). Areas of greatest risk of collapse are steep slopes within the hydrothermal system outline. Coordinates are easting and northing in m using the NZTM projection.

The areas of coincident steep slope and hydrothermal alteration occur on slopes of all aspects and size. The most at risk slopes, and those representing the highest hazard, are high on the massif where they have large potential energy and the longest runouts. Such slopes occur on the north flanks above Upper Te Maari and on the west flanks of Tongariro summit, above the Mangatepopo Valley and around the head of the Mangahouhoumui and Oturere Valleys. A more detailed ground study of alteration and slope stability in these areas would provide a better assessment of the hazard these slopes present.

4.7 Conclusions

Geologically constrained geophysical inversions of an extensive potential field data set at the TgVM have successfully mapped the basement structure beneath the volcano, identified magmatic plumbing system roots, delineated the extent of a large hydrothermal system, highlighted areas at risk from various volcanic hazards and offer improvements to volcanic unrest monitoring capability.
Our model shows a continuous dense, non-magnetic basement beneath the volcano, and suggests places where it is pierced by the magmatic plumbing system. The basement is extensively down faulted to a depth of around 100 m below sea level under the TgVM, a total to 500–700 m displacement across the graben. We calculate that the volume of volcanic material above the basement is five to six times larger than previous geologically based estimates, requiring a higher rate of magma supply than previously thought. However, the lack of discrete magma bodies within the model indicates that when magma is supplied to the surface it is only via relatively small batches. We find only minor evidence for the Waihi faults acting as preferential magma pathways for the Pukekaikore vents and conclude that if feeder dykes have intruded along them, they are likely too small to be resolved by our surveys. Thus, the main bounding faults are considered low probability areas for future eruptions. We image a low density and highly magnetic root beneath Mt Ngauruhoe which we interpret to represent the main system of magmatic feeder conduits beneath the volcano.

The zone of hydrothermally demagnetised rocks extends to around basement depths and is bound to the west by the Waihi fault system, which acts as an impermeable barrier to fluid movement. The hydrothermal system is not imaged beneath Mt Ngauruhoe, so while there is other evidence for hydrothermal activity there, fluid circulation within the cone is likely not developed enough to cause sufficient alteration to be detected by our survey. Where the hydrothermal system intersects steep topographic slopes, we map areas most at risk from flank collapse. Flank collapse potential may be elevated by pressurisation of the hydrothermal system following dyke intrusion, or by over-saturation of pore space during heavy rainfall or snow melt. Finally, we propose that our high resolution density model could act as a proxy for a new 3D velocity model to improve earthquake locations and enhance volcanic unrest monitoring capability.

4.8 Acknowledgements

Corinne Locke and John Cassidy are thanked for their generous access to unpublished gravity and aeromagnetic data. C.M. is funded by GNS Science Core Funding, EQC New Zealand, Mitacs Accelerate Canada and Mira Geoscience. Many field assistants helped with gravity data collection including Nellie Olsen, Natalia Deligne, Sophie Pearson, Alex Kmoch, Tom Ayling, Janvion Cevuard, Matt Stott and Nick MacDonald. Thanks to Vaughan Stagpoole for running the terrain corrections on the GNS system. Thanks to Harry Keys at the Department of Conservation for permitting assistance and to helicopter pilots Keith McKenzie and Andrew MacIntosh for logistical support. Thomas Campagne, Peter Fullagar, Stanislawa Hickey and Shannon Frey at Mira Geoscience provided many hours of VPmg and GOCAD assistance. Thanks to Jeff Zurek and Jeff Witter for discussions on the models and to Bruce Christenson, Dougal Townsend and Graham Leonard for discussions on the hydrothermal system and geology of the TgVM. We thank Fabio Caratori-Tontini and an
anonymous reviewer for comments that improved the manuscript. Figures and analysis were made using open source software, Python, Matplotlib (Hunter, 2007), Inkscape and QGIS.

4.9 References


Hochstein, M., 1985. Steaming Ground at Red Crater and in the Te Maari Craters. 7th NZ Geothermal Workshop 177–180


In the preceding chapter I presented a model of the volcanic architecture of Mt Tongariro, that defined the extent of the hydrothermal system within the volcanic pile above the basement rock. In 2012, two eruptions occurred at Upper Te Maari crater, driven by dyke intrusion into the hydrothermal system. In this chapter I use microgravity and ground displacement measurements to address the fourth thesis objective, ‘understand the response of the Tongariro hydrothermal system to eruptions in 2012’ (Table 1.1). I present a model of mass transfer processes occurring after the eruption, applicable to many volcano hydrothermal systems, and assess whether the Tongariro system is repressurising or depressurising. In this chapter I pose the following hypotheses:

- The same source and physical process can explain both the measured deformation and gravity change at Upper Te Maari crater, post–2012 eruption.

- More than one source is responsible for the observed gravity changes at Upper Te Maari crater.

Chapter 5 is submitted to the Journal of Volcanology and Geothermal Research, by Craig Miller, Gilda Currenti, Ian Hamling, and Glyn Williams-Jones. The author contributions are as follows. CM designed the gravity network, collected, processed, modelled, interpreted the data, and wrote the manuscript. GC assisted with the FEM model under the direction of CM, IH provided deformation data, and ran the deformation model. All authors commented on the manuscript prior to submission.
Chapter 5

Mass Transfer Processes in a Post Eruption Hydrothermal System: Parameterisation of Microgravity Changes at Te Maari Craters, New Zealand

Abstract

Fluid transfer and ground deformation at hydrothermal systems can occur both as a precursor to, or as a result of an eruption. Typically studies focus on pre-eruption changes to understand the likelihood of unrest leading to eruption; however, monitoring post-eruption changes are important for tracking the return of the system towards background activity. Here we describe processes occurring in a hydrothermal system following the 2012 eruption of Upper Te Maari crater on Mt Tongariro, New Zealand, from observations of microgravity change and deformation. Our aim is to assess the post-eruption recovery of the system, to provide a baseline for longer term hazard assessment. We model microgravity changes using a range of analytic solutions to determine the most likely geometry and source location. Using a multiobjective inversion, we test whether the gravity change models are consistent with the observed deformation. We conclude that the source of deeper seated subsidence is separate from the location of mass addition. From this unusual combination of observations, we develop a model of fluid transfer within a condensate layer, occurring in response to eruption-driven pressure changes. We find that depressurisation drives the evacuation of pore fluid, either exiting the system completely as vapour through newly created vents and fumaroles, or migrating to shallower levels where it accumulates in empty pore space, resulting in positive gravity changes. Evacuated pores then collapse, causing subsidence. We explore the likely microgravity and deformation responses to a range of other processes occurring within the system, as well as the contributions of external fluid inputs and losses.
Long term combined microgravity and deformation monitoring will allow us to track the re-sealing and re-pressurisation of the hydrothermal system and assess what hazard it presents to 1000s of hikers who annually traverse the volcano, within 2 km of the eruption site.

5.1 Introduction

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

Microgravity is a useful tool to monitor hydrothermal systems (e.g. Carbone et al., 2017), as it can detect mass changes that can be interpreted in terms of fluid flux or phase changes (condensation, etc.), as well as distinguish between fluids of magmatic or hydrothermal origin (Battaglia et al., 2006). While deformation measurements alone can locate pressure sources, interpreting those sources in terms of fluid movement, or thermal changes, in poroelastic media is more challenging (Fournier and Chardot, 2012). This is especially so for hydrothermal systems, where there are unlikely to be large cavities or chambers often used to simulate deformation measurements. In hydrothermal systems, microgravity gives us a way to reinterpret pressure cavity based deformation models, in terms of fluid transfer in a porous medium, using analytic (e.g. Allis and Hunt, 1986) or numerical models (Rinaldi et al., 2011; Coco et al., 2016; Currenti and Napoli, 2017). Additionally, the combination of deformation and microgravity can determine if a single source model can explain both sets of observations or if there are multiple processes occurring, requiring a combination of sources (e.g. Miller et al., 2017a). Hence, tracking changes in both deformation and microgravity sources over time provides a richer understanding of the system dynamics.

Mount Tongariro Volcanic Centre hosts a shallow, vapour dominated hydrothermal system (Walsh et al., 1998), as mapped from the extent of demagnetised and altered rocks (Hill et al., 2015; Miller and Williams-Jones, 2016). The hydrothermal system is likely confined to more permeable volcanic material above a low permeability greywacke basement surface modelled at an elevation of around 0 m a.s.l. beneath the Te Maari area (Miller and Williams-Jones, 2016). Upper Te Maari crater, on Mt Tongariro, erupted just before midnight (NZST) on the 6th August 2012 (Crouch et al., 2014), after 24 days of unrest (Hurst et al., 2014) that followed 125 years of quiescence (Scott and Potter, 2014). On the 21st November 2012, upper Te Maari erupted again at 1 pm, this time without warning. Ballistic blocks from the August eruption punctured the roof and floor of an unoccupied hiking hut 2 km away. Dyke injection into the base of the hydrothermal system (Christen-
son et al., 2013) increased pore pressures within mechanically weak, hydrothermally altered rocks (Montanaro et al., 2016). A landslide triggered the eruption by unloading the system by $\sim$0.5 MPa (Procter et al., 2014) resulting in rapid boiling and expansion of the over-pressure hydrothermal fluids. The intruded volume of magma was small enough that no local deformation was detected prior to eruption (Jolly et al., 2014), and no traces of fresh magma were found in the erupted deposits (Pardo et al., 2014). The eruption enlarged the existing Upper Te Maari crater and created a fissure 30 m deep and 400 m long (Figure 5.1), south of the main crater.

Hamling et al. (2016) modelled a post eruption subsidence trend at Te Maari from persistent scatter InSAR data, and suggested it was due to de-pressurisation of the hydrothermal system. They modelled the deformation as the closing of a sill like crack, located at around 500 m depth beneath Upper Te Maari. We extend that work with new microgravity, deformation and fumarole temperature measurements taken after the eruption, to better understand the processes driving the observed subsidence and gravity increases. The combination of gravity increases with subsidence is unusual (Carbone et al., 2017) (see Jousset et al. (2000); Crider et al. (2008) for examples), and we perform an extensive suite of analytic solutions to test the likely source parameter distribution and to determine if subsidence and gravity change sources are co-located. We initially model the gravity and deformation data independently to determine the optimal solutions for each dataset. We then investigate joint inversion, to see if a single source, and hence process, can explain both sets of observations. We use a numerical, finite element, forward model to test if the effects of topography and geological heterogeneities are significant. Finally, we semi-quantitatively describe the likely combination of processes occurring within the hydrothermal system that contribute to the observed data. Our aim is to gain a fuller understanding of the hydrothermal processes occurring post eruption at Te Maari, to better inform volcano monitoring efforts and hazard assessment at a volcano that is visited by 1000s of hikers annually.

5.2 Data

5.2.1 Gravity measurements

In February 2014 we established a network of 14 gravity benchmarks on accessible ground to the north and east of Upper Te Maari crater, at a spacing of 100 to 300 m (Figure 5.1). Steep ground to the south of the Upper Te Maari crater limits data coverage in this area. We installed benchmarks by driving stainless steel rods to refusal into the pyroclastic deposits from the 2012 eruptions. We installed a reference benchmark, TGKB, off the volcano, approximately 5 km to the north, adjacent to highway 46 (Figure 5.1). Access to the network is by helicopter, and then on foot. We used a single LaCoste and Romberg gravity meter (G106) for all surveys, with each benchmark surveyed independently 2 or 3 times in separate loops, from local base benchmark TGM03. We repeated the network at 11 to
Figure 5.1: Location of microgravity network overlain on 0.5 m resolution DEM of the pre 2012 eruption landscape. Red triangles are the gravity benchmarks. Red lines in the smaller figure indicate active faults, and white lines are roads. The reference station, TGKB, is shown as a red triangle in the insert figure. The purple shaded area is the vent and fissure formed during the August 2012 eruption.

Legend

- 2012 eruption vent
- microgravity benchmarks
- roads
- lake
- faults
- fumarole

130
12 month intervals in January 2015, January 2016 and December 2016 to determine mass changes over time. We refer to the December 2016 measurement as ‘2017’ for convenience.

Using Gtools (Battaglia et al., 2012), we corrected daily measurements for Earth tide and ocean loading, along with linear drift corrections, to determine gravity changes relative to TGM03. TGM03 is tied to the reference benchmark TGKB each day so that all final gravity changes are relative to TGKB. We combined multiple repeat readings to a single value through a least squares adjustment, with standard deviations and standard errors calculated for each benchmark (Table 5.1). After 2015, benchmark TGM10 was buried by a small landslide. We used a locally measured free air gradient of -0.3047 mGal/m to correct for the observed subsidence (Battaglia et al., 2008).
Table 5.1: Microgravity change data from 2014 to 2017. Microgravity change values (Deltag) are in $\mu$Gal, as are the standard error (STDER) values. TGM10 was not able to be surveyed after 2015. Height change values (DeltaH) are in m and are computed at the benchmark locations using the deformation model of Hamling et al. (2016). Height change errors are nominally 0.01 m.

<table>
<thead>
<tr>
<th></th>
<th></th>
<th></th>
<th></th>
<th></th>
<th></th>
<th></th>
<th></th>
<th></th>
<th></th>
<th></th>
<th></th>
<th></th>
</tr>
</thead>
<tbody>
<tr>
<td>TGM01</td>
<td>-39.1037</td>
<td>175.6745</td>
<td>1531</td>
<td>-0.020</td>
<td>3.3</td>
<td>14</td>
<td>-0.015</td>
<td>44.3</td>
<td>16</td>
<td>0.012</td>
<td></td>
<td></td>
</tr>
<tr>
<td>TGM02</td>
<td>-39.1092</td>
<td>175.6747</td>
<td>1523</td>
<td>-0.020</td>
<td>49.8</td>
<td>12</td>
<td>-0.017</td>
<td>43.9</td>
<td>13</td>
<td>0.013</td>
<td></td>
<td></td>
</tr>
<tr>
<td>TGM03</td>
<td>-39.1085</td>
<td>175.6726</td>
<td>1512</td>
<td>-0.020</td>
<td>53.3</td>
<td>6</td>
<td>-0.018</td>
<td>45.7</td>
<td>8</td>
<td>0.014</td>
<td></td>
<td></td>
</tr>
<tr>
<td>TGM04</td>
<td>-39.1083</td>
<td>175.6707</td>
<td>1517</td>
<td>-0.020</td>
<td>55.5</td>
<td>15</td>
<td>-0.011</td>
<td>82.1</td>
<td>16</td>
<td>0.006</td>
<td></td>
<td></td>
</tr>
<tr>
<td>TGM05</td>
<td>-39.1065</td>
<td>175.6690</td>
<td>1442</td>
<td>-0.015</td>
<td>1.6</td>
<td>10</td>
<td>-0.007</td>
<td>42.3</td>
<td>12</td>
<td>0.002</td>
<td></td>
<td></td>
</tr>
<tr>
<td>TGM06</td>
<td>-39.1050</td>
<td>175.6685</td>
<td>1437</td>
<td>-0.009</td>
<td>5.1</td>
<td>10</td>
<td>-0.002</td>
<td>20.8</td>
<td>12</td>
<td>0.001</td>
<td></td>
<td></td>
</tr>
<tr>
<td>TGM07</td>
<td>-39.1087</td>
<td>175.6759</td>
<td>1528</td>
<td>-0.020</td>
<td>42.9</td>
<td>11</td>
<td>-0.013</td>
<td>92.4</td>
<td>11</td>
<td>0.008</td>
<td></td>
<td></td>
</tr>
<tr>
<td>TGM08</td>
<td>-39.1065</td>
<td>175.6736</td>
<td>1508</td>
<td>-0.020</td>
<td>38.9</td>
<td>14</td>
<td>-0.026</td>
<td>52</td>
<td>16</td>
<td>0.023</td>
<td></td>
<td></td>
</tr>
<tr>
<td>TGM09</td>
<td>-39.1019</td>
<td>175.6722</td>
<td>1501</td>
<td>-0.020</td>
<td>33.6</td>
<td>14</td>
<td>-0.024</td>
<td>52.1</td>
<td>15</td>
<td>0.019</td>
<td></td>
<td></td>
</tr>
<tr>
<td>TGM10</td>
<td>-39.1056</td>
<td>175.6727</td>
<td>1339</td>
<td>-0.017</td>
<td>-</td>
<td>-</td>
<td>-0.017</td>
<td>-</td>
<td>-</td>
<td>-0.007</td>
<td></td>
<td></td>
</tr>
<tr>
<td>TGM11</td>
<td>-39.1041</td>
<td>175.6731</td>
<td>1341</td>
<td>-0.018</td>
<td>-15.7</td>
<td>10</td>
<td>-0.025</td>
<td>-11.6</td>
<td>10</td>
<td>0.014</td>
<td></td>
<td></td>
</tr>
<tr>
<td>TGM12</td>
<td>-39.1080</td>
<td>175.6732</td>
<td>1309</td>
<td>-0.020</td>
<td>-10.5</td>
<td>15</td>
<td>-0.034</td>
<td>34.6</td>
<td>16</td>
<td>0.030</td>
<td></td>
<td></td>
</tr>
<tr>
<td>TGM13</td>
<td>-39.1070</td>
<td>175.6757</td>
<td>1479</td>
<td>-0.020</td>
<td>-15.5</td>
<td>15</td>
<td>-0.037</td>
<td>1.5</td>
<td>16</td>
<td>0.031</td>
<td></td>
<td></td>
</tr>
<tr>
<td>TGM14</td>
<td>-39.1105</td>
<td>175.6745</td>
<td>1539</td>
<td>-0.020</td>
<td>14.4</td>
<td>11</td>
<td>-0.005</td>
<td>19.2</td>
<td>12</td>
<td>0.002</td>
<td></td>
<td></td>
</tr>
</tbody>
</table>
5.2.2 Deformation measurements and models

We calculated the vertical displacement at each benchmark location from a forward model of the best fit deformation model of Hamling et al. (2016), as this proved more reliable than short GNSS occupations for small displacements. We updated this model for each gravity measurement interval using ALOS-2, InSAR observations. We processed ALOS-2 data from two descending tracks in stripmap mode to measure the deformation over Tongariro using the InSAR Scientific Computing Environment (ISCE) software (Rosen et al., 2012). Topographic corrections were made using the 30 m SRTM data (Farr et al., 2007). Both of the interferograms are filtered using a power-spectrum filter (Goldstein et al., 1988), and unwrapped using SNAPHU (Chen and Zebker, 2002). To model the data, the line-of-sight displacements are resampled onto a 120 m grid in the vicinity of the eruption site and every ~1 km in the far-field.

The vertical displacement is used to apply free air corrections to the gravity data. The modelled vertical displacements have an uncertainty of ±0.01 m, or approximately 3 µGal (3 x 10^{-8} m/s^2). Between 2014 and 2015, approximately 9 to 20 mm of vertical subsidence is modelled across the gravity network and the centroid of the displacement source is coincident with the main vent area. Between 2015 and 2016, maximum subsidence increased to 37 mm across the gravity network, and the center of displacement moved ~450 m to the north-east, away from the vent. Between January 2016 and December 2016, vertical subsidence of 1 to 33 mm is calculated at the gravity benchmarks and the source of subsidence is similar to the 2016–2015 source (Figure 5.2). In January 2016 and December 2016, we surveyed benchmarks TGM04 and TGM13 with static GNSS measurements to provide independent measurements of deformation. Vertical displacements of -51 and -5 mm were observed at these benchmarks respectively; however, the velocities are poorly constrained from only two measurements, so we do not use these observations. Table 5.2A summarises the deformation source parameters used to fit the InSAR data in each time period using a sill like source (Okada, 1985).

For the joint deformation and gravity inversion (Section 5.5), we use the line of sight (LOS) persistent scatter dataset from Hamling et al. (2016), updated for each gravity measurement interval. Positive LOS displacement indicates movement away from the satellite or subsidence. We track persistent scatterers or coherent pixels in the region of the observed gravity changes. As there is only one look direction available, we are only able to use LOS displacements, which limits the accuracy of the source depth and geometry; however, the spatial coverage is improved from using only the gravity benchmarks. Figure 5.2 shows the LOS displacement pixels for each gravity measurement interval.
Figure 5.2: Gravity changes (shaded contour lines) and line of sight (LOS) displacements (coloured dots) from Hamling et al. (2016) for the intervals, 2014 to 2015, 2015 to 2016 and January 2016 to December 2016. Positive LOS indicates subsidence. Black triangles are gravity benchmarks, purple shaded area is vent region. The red dot is the location of the point source mass (see Figure 5.3) and the red rectangle is the outline of the sill dislocation model from Hamling et al. (2016).
Table 5.2: Source parameters for the deformation data, from A), the finite element model of Hamling et al. (2016), updated for the gravity time periods and B), the best fit deformation model from the NSGA-II joint inversion. Xc is the centroid Easting, Yc is the centroid Northing. The reference elevation for depth conversion is 1490 m a.s.l.
5.3 Residual Gravity Change Results

The sub-optimal station coverage (Figure 5.1), with no stations to the south of the vent, means the gravity change anomaly is not fully defined and should be considered a minimum value. In the time intervals studied, the maximum gravity change is at stations closest to the Upper Te Maari eruption vent (Figure 5.2). In 2014 to 2015, the gravity changes are broadly coincident with the areas of subsidence observed in the InSAR data, but cover a smaller area. In later time intervals, the gravity and deformation signals are similar in spatial extent, but the maximum gravity changes are \( \sim 450\) m to the south west of the maximum subsidence.

5.3.1 2014 to 2015

Residual gravity changes in the 12 months from February 2014 to January 2015 show a positive gravity change with a maximum of \( 51 \pm 18 \) \( \mu \text{Gal} \) (standard error) at benchmark TGM04. The gravity anomaly is oriented NE-SW, broadly along strike of the main tectonic features. A Wilcoxon signed-rank test shows the 2015 results are significantly different to the 2014 results at \( p = 0.001 \).

As the anomaly is not completely defined, we are not able to calculate the mass change through Gaussian integration of the anomaly. Instead, we assume a point source, and invert our data for the mass of the point source. We use the `curvefit` function of python module, `scipy.optimise`, to fit the equation of the gravity effect of a point source, \( \Delta g_z \) (Battaglia et al., 2008), to our observed gravity changes, including the standard error for each measurement:

\[
\Delta g_z = Gm \frac{d}{(r^2 + d^2)^{3/2}}
\]

where mass \( m = \rho \Delta V \) and \( G = 6.67 \times 10^{-11} \text{Nm}^2\text{kg}^{-2} \), \( d \) is the depth of the point source, and \( r \) is the radial distance from the surface projection of the source. The average elevation of the benchmarks is 1490 m a.s.l. (variation 1340 to 1540 m), which is used as the reference for depth below surface. The nonlinear least squares minimization requires starting estimates for the fit parameters. To avoid convergence in local minima, we randomly select the starting parameters within a range of reasonable values and run 2000 iterations of the inversion to test the distribution of the parameter space. In all cases the inversion converged on the same result, regardless of the starting model.

For the 2014 to 2015 interval, a mass of \( 1.08 \times 10^9 \pm 5.0 \times 10^2 \) kg at a depth of \( 433 \pm 1 \) m (\( \sim 1057 \) m elevation) is calculated from the point source curve fit, with \( R^2=0.62 \) (Figure 5.3).
Figure 5.3: Curve fit for point source solution for all intervals. Dots are data points with error bars; curved lines are the fit of the data. For 2015–2014 $R^2=0.62$, in 2016–2015 $R^2=0.71$, and in 2017–2016 $R^2=0.85$. 

137
5.3.2 2015 to 2016

Residual gravity changes in the 12 months from January 2015 to January 2016 show a positive gravity change of $55 \pm 15 \mu \text{Gal}$ at benchmark TGM04, decreasing to $-10 \pm 15 \mu \text{Gal}$ at benchmark TGM11. A Wilcoxon signed-rank test shows the 2016 results are significantly different to the 2015 results at $p = 0.009$. The 2015 to 2016 anomaly gradient is steeper than the 2014 to 2015 anomaly, suggesting a shallower source. The best fit point source solution ($R^2=0.71$) yields a mass change of $3.68 \times 10^8 \pm 1.7 \times 10^2 \text{ kg}$ at a depth of $182 \pm 1 \text{ m (\sim 1308m elevation)}$, (Figure 5.3).

5.3.3 2016 to 2017

Residual gravity changes in the 11 months from January 2016 to December 2016 show a positive gravity change of $92 \pm 11 \mu \text{Gal}$ at benchmark TGM07, decreasing to $-11 \pm 9 \mu \text{Gal}$ at benchmark TGM11. A Wilcoxon signed-rank test shows the December 2016 results are significantly different to the January 2016 results at $p = 0.002$. The best fit point source solution ($R^2=0.85$) yields a mass change of $1.50 \times 10^9 \pm 4.0 \times 10^2 \text{ kg}$ at a depth of $269 \pm 1 \text{ m (\sim 1221m elevation)}$, (Figure 5.3).

5.4 Gravity Analytic Source Inversion

To test whether the point source solution is an appropriate geometry, we model a range of analytic solutions and compare the fit of each model to the observed data. 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, vertical rectangular prism and horizontal sill (Clark et al., 1986; Okubo, 1992). Genetic algorithms are a class of Monte Carlo algorithm (Sambridge and Mosegaard, 2002), 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 non-unique solutions are inevitable. The GA randomly generates an initial population of 200 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, such as mutation, crossover and selection, for 100 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 thousands of times for each geometry. In this way we generate populations of final models, where each individual model has been selected from 200 individuals, chosen from a randomised starting model. This approach allows us to test the stability and sensitivity of the models over a wide range of randomised starting parameters and statistically determine which geometry best explains the observed data.
5.4.1 Gravity source models

Analytic solutions of simple geometries are well suited for this study, given the relatively few survey points and incomplete spatial coverage. Even though the anomaly is poorly constrained, analytic models allow us to make an estimate of the most likely source geometry, i.e. are more spherical source shapes preferred to more rectangular source shapes? In all cases, the southern extent of the geometry and location parameters is poorly constrained in the models. 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), vertical prism (8) and horizontal sill (7). The sphere geometry is defined by its centroid coordinates (Xc, Yc, depth to centre), axis radius and density change. The prolate spheroid is defined by its centroid coordinates (Xc, Yc, depth to centre), 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 both minor axes are free to adjust. The oblate triaxial spheroid restricts one minor axis ratio so that an oblate geometry is produced. The vertical prism geometry is defined by its centroid coordinates (Xc, Yc, depth to top), length, width, strike, thickness and density change, with dip fixed at 90 degrees. The sill geometry is defined by the centroid coordinates (Xc, Yc, depth to top), length, width, strike, and UM (the thickness and density product). The gravity change of a rectangular prism linearly depends on the product (UM) of thickness and density, and the inverse problem becomes undetermined if the two are solved for separately. If an assumption about the density of the source fluid is made (i.e. water at 1000 kg/m$^3$) then an estimate of the thickness can be calculated from UM / density. The location of the source centroid is confined to be within the gravity anomaly area and the depth is constrained from 1400 m a.s.l. to 0 m a.s.l. (≈ 100 to 1500 m depth).

To determine the best-fit source geometry, after removing non-physical models from the population, we calculate the reduced chi-squared ($\chi^2_{red}$) statistic on each of the individual models for each geometry, defined as:

$$\chi^2_{red} = \frac{1}{v} \sum_{k=1}^{n} \frac{(O_k - E_k)^2}{\sigma_k^2}$$

where $O_k$ are the observed data, $E_k$ are the calculated data, $\sigma_k$ is the standard deviation of the observation, $v$ 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 a F-test (Wackerly et al., 2007), implemented in the python scipy.levene module, on the $\chi^2_{red}$ values to determine which model geometry population best fits the data. Figure 5.4 shows box plots of $\chi^2_{red}$ calculated for each model type, coloured by observation interval. The sphere, vertical prism and horizontal sill show the lowest $\chi^2_{red}$ at all time intervals, so we only consider these models from here-on.
Figure 5.4: Box and whiskers plots showing the $\chi^2_{red}$ for the range of gravity only model geometries tested. In each geometry segment the three observation intervals are shown. The box shows the quartiles of the data set, while the whiskers extend to show the rest of the distribution. Dots are outlier values.

To assess the variability of each model 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), implemented in the seaborn python module. We refer to the peak of the KDE as the mode from here-on.

A summary of the depths of the sill and sphere models for each time interval is shown in Figure 5.5, with the deformation source depth for comparison.

2014 to 2015 source model parameters

The analytic sphere, vertical prism and sill models have similar $\chi^2_{red}$ (Figure 5.4). The sill has the lowest mean $\chi^2_{red}$ and the smallest standard deviation of the $\chi^2_{red}$ distribution. An F-test shows that the sill model is a better fit to the data than the prism model with an F statistic of 510 at p <0.01, and to the sphere with an F statistic of 3577 at p <0.01. The sill RMS fit is 7 $\mu$Gal (Figure 5.6D). The analytic sill models have a mode elevation of 1251 m a.s.l. (approximately 239 m depth), mode length of 586 m, mode width of 559 m, and mode UM of 1995 kg/m$^2$ (Figure 5.6A). The length parameter of the sill is poorly constrained by the lack of station coverage to the south. From the UM, length and width product, we calculate a mass change of $1.01 \times 10^9$ kg. As the sill formulation is unable to calculate an independent density change parameter, we use the sphere model, and obtain a density change of 65 kg/m$^3$. By comparison, the sphere model has a mode depth of 530 m (960 m...
Figure 5.5: Distribution of model depths for each time interval. The sphere and sill model depth distributions are shown as green and blue KDE plots from the gravity only inversions, while the best fit deformation solution is shown as a black dot.
Table 5.3: Summary of the sill parameters for A) the gravity only model, and B) the best gravity model from the NSGA-II joint inversion. In A) All values are the peak KDE or ‘mode’. * indicates a double peaked distribution. Xc is the centroid Easting, Yc is the centroid Northing, UM is the thickness and density product. The reference elevation for depth conversion is 1490 m a.s.l.

elevation), with a mode radius of 141 m and mass of 7.6 x 10^8 kg. Table 5.3A summarises the sill model parameters.

2015 to 2016 source model parameters

The rectangular models are generally a better fit than the spherical models for the 2015 to 2016 interval. The sill is a better fit than the prism model with an F statistic of 145 at p <0.01, and is also a better fit than the sphere model, with an F statistic of 908 at p <0.01. The RMS fit of the sill is 12 μGal (Figure 5.6E). The sill models have a mode elevation of 1385 m (~105 m depth), mode length of 795 m, mode width of 272 m, and mode UM of 2779 kg/m² (Figure 5.6B). The length parameter of the sill is poorly constrained by the lack of station coverage to the south. From the UM, length and width product, we calculate a
Figure 5.6: Summary of the sill and sphere gravity model locations (A, B, C) and observed vs calculated data (D, E, F) for each gravity observation interval. The blue rectangles are the sill outlines, green dots are the sphere centroids. Gravity change contours are shown in black lines, with benchmarks as yellow triangles. The purple outline is the vent area. The black dashed square in A–C is the extent of the position ranges for the inversion. The observed data in plots D–F are shown with standard error bars. The calculated data are for the sill model.
mass change of $6.43 \times 10^8$ kg. By comparison, the sphere model has a mode depth of 183 m (1307 m elevation), mode radius 120 m, bi-modal density change of 23 or 264 kg/m$^3$, and mode mass addition of $1.5 \times 10^9$ kg.

2016 to 2017 source model parameters

In the 2016 to 2017 interval, the sill model is no longer the best fit geometry. The sphere is a better fit to the sill with an F statistic of 123 at $p < 0.01$ and also to the prism with an F statistic of 945 at $p < 0.01$. The RMS fit of the sill is $10 \mu$Gal compared to $3 \mu$Gal for the sphere (Figure 5.6F). For comparison with the previous intervals, we plot the sill solutions along with the sphere (Figure 5.6C). The sill models have a mode elevation of 1291 m (approximately 199 m depth), mode length of 754 m, mode width of 436 m, and mode UM of 5852 kg/m$^2$. From the UM, length and width product, we calculate a mass change of $1.80 \times 10^9$ kg. The better fitting sphere models are centred at a mode depth of 250 m (1240 m elevation) with a radius of 176 m and density change of 122 kg/m$^3$, resulting in a mass addition of $2.78 \times 10^9$ kg.

5.5 Joint Gravity and Deformation Analytic Source Inversion

The best fit deformation solution of Hamling et al. (2016) has similar geometry and location to the preferred sill gravity models from the 2014 to 2015 interval, suggesting a common source may be responsible for mass and deformation changes. In later intervals there is less overlap, but there may still be a solution that adequately fits both data sets. Therefore we jointly invert the gravity and deformation data to determine the likelihood that a co-located source produces the observed gravity and deformation changes. We use the sill solution, as this best fits the individual gravity and deformation datasets for the majority of the time intervals. The combined gravity and deformation sill solution utilises analytic expressions to account for the effects arising from tensile cracks buried in a homogeneous half-space (Okada, 1985), and calculates the displacement due to the crack opening (or closing). The Okubo (1992) solution calculates the corresponding gravity changes from contributions due to 1) the density changes due to compressibility of the medium, 2) the surface mass distribution and 3) the input of new mass into the cavity generated by tensile opening (Carbone et al., 2008) (Figure 5.7). Coupling these sources as a multi-objective problem, allows us to simultaneously model the displacement and gravity changes due to a tensile source. Since the tensile dislocation (U3) is small (a few cm) in comparison to the thickness of the source required to fit the gravity data, we assumed that the density changes occur in a thicker source (UM) than the displaced sill. Assuming a water-filled pore space, the volume in which the displacement changes occur in the best fit Hamling et al. (2016) model, contains a mass of between 1 to $4 \times 10^7$ kg in the three time intervals investigated.
Figure 5.7: Parameters of the Okada and Okubo models for joint gravity and deformation inversion. In our model the sill is horizontal and there are no strike or dip slip components. UM (density and thickness product) is required to be approximately 100 times greater than U3 (tensile displacement) to fit the observed gravity data (the diagram is not to scale). Note that U3 in our models is negative, representing a closure of the sill, not opening. Hydraulic conductivity directions are shown as $K_x K_y K_z$. This is around 100 times smaller than the observed mass changes ($1 \times 10^9$ kg), suggesting the gravity change source region may be thicker than the displacement source by a factor of around 100. Indeed in the gravity-only sill solution for 2015–2014, if water is assumed to be the fluid, the UM parameter (1995 kg/m$^2$), divided by the water density (1000 kg/m$^3$), gives a thickness of around 2 m, which is 100 times greater than the U3 (compression closing) parameter (-0.024 m) from the deformation model.

For the joint inversion we use the NSGA-II algorithm (Non-dominated Sorting Genetic Algorithm), devised by Deb et al. (2002) and demonstrated for joint gravity and deformation by Carbone et al. (2008). The joint inversion minimises a multi-objective function, to solve for both gravity and deformation. In a multi-objective geophysical inversion, it is likely no single solution exists that simultaneously minimises each objective. Rather there exist a family of solutions known as a Pareto-optimal set (Fonseca and Fleming, 1993), each of which represents a trade off of one objective against the other. Each point in the Pareto set is optimal in the sense that no improvement can be achieved in one component without degradation of another. An advantage of the NSGA-II approach is there is no requirement for manual weighting of the objective function towards either dataset. Instead the NSGA-II algorithm uses the Pareto frontier to test the dominance of one solution over another, based
on the $\chi^2$ result of each solution, and thus determine if a solution conflicts or cooperates with the objective functions. The solutions are iterated using a Genetic Algorithm approach similar to that used for the gravity only solutions, i.e. using selection, crossover and mutation operators on real numbers (Carbone et al., 2008). Similar to the individual inversions, the NSGA-II generates a range of optimal solutions, consistent with the inherent non-uniqueness of potential field data. The Pareto frontier is visualised by plotting the $\chi^2$ or RMS of each objective against each other. A Pareto frontier with widely dispersed solutions indicates a low likelihood of a single source fitting both objectives, while a compact front indicates the solutions which most likely fit both data sets. The NSGA-II was run for 600 generations using a population of 500 individuals.

As we have a population of optimal solutions, we can examine those that best fit the data by examining all the models whose $\chi^2$ value is less than a chosen $\chi^2_\alpha$ given by the $F$ ratio test. For our number of observations and model parameters, we calculate the $\chi^2_{opt}$ for the optimal models of the gravity and deformation solutions (using scipy.f.ppf) at a chosen confidence limit (95%). All the models with $\chi^2 < \chi^2_\alpha$ are consistent with the optimal models at the 95% confidence limit, where:

$$\chi^2_\alpha = \chi^2_{opt} \left[ 1 + \frac{p}{n - p} F(p, n - p, 1 - \alpha) \right]$$

and $p$ is the number of model parameters, $n$ is the number of data and $F$ is the F distribution with $p$ and $n-p$ degrees of freedom (Carbone et al., 2008).

Figure 5.8 shows the results of the NSGA-II joint inversion for each time interval. In parts A–C, the Pareto optimal set of solutions is shown in grey rectangles with the best fit gravity solution in blue and best fit deformation in red, as also shown in the Pareto front plots in D–F. The Pareto fronts for all time intervals show poor clustering where the best fit gravity and deformation solutions (larger dark blue and dark red dots) are located at extremes of the distribution, indicating conflicting behaviour of the objective functions. Solutions within the 95% confidence limits for the individual gravity and deformation data sets are shown as light blue and light red dots, respectively, in D–F. In the 2015–2014 and 2017–2016 intervals, there is a minor overlap of models at the ends of the individual 95% confidence limits; however, in the 2016–2015 interval there is a clear break between the distributions. The model locations and geometries from the individual best fits in A–C are clearly separated. The best deformation solution is generally similar to the stand alone deformation results (Table 5.2), as is the best gravity solution to the gravity only models (Table 5.3). From the multi-objective inversion we conclude that the gravity and deformation sources occur in close proximity to each other, but they are in fact separate sources, and hence require different physical processes to be driving them.
Figure 5.8: Summary of the joint gravity and deformation NSGA-II inversion models. A,B,C show the locations of the joint sill models for the 2015–2014, 2016–2015 and 2017–2016 time intervals, respectively. Grey rectangles represent the optimal Pareto set of models with the best gravity model in blue and best deformation in red. Observed gravity change contours are shown in black lines, with coloured dots as the observed line of sight (LOS) displacement. Positive LOS indicates subsidence. Yellow triangles are gravity benchmarks. D,E and F show the Pareto front of the RMS gravity vs RMS deformation solutions. Light blue dots represent gravity solutions within the 95% confidence limit of the best gravity solution, and light red dots are the equivalent for the deformation solutions. The best individual gravity and deformation solutions are shown in larger dark blue and dark red dots, respectively. The widespread distribution of the Pareto front indicates separate gravity and deformation sources.
5.6 Numerical Forward Model to Investigate Effects of Topography and Heterogeneity

Volcanic systems rarely have isotropic or homogeneous physical property distributions, as assumed by the analytical models. The sill shaped models also produce a non-zero internal gravity change when they deform, and steep topography can modify the observed deformation from shallowly buried sources. To test whether topography, heterogeneities or internal deformation causing gravity changes affect our analytic models, we construct a 3D finite element model using COMSOL Multiphysics (v4.3b) to forward model the deformation and gravity changes. We use the best fit deformation solution of the NSGA-II inversion from 2015 to 2016 as the source (Table 5.2B), as this is the shallowest model, most likely to be affected by topography. We incorporate the topography, and subsurface density boundaries derived from a Bouguer gravity model of Mt Tongariro (Miller and Williams-Jones, 2016) as well as different mechanical properties for the greywacke basement and overlying volcanics (Mielke et al., 2016) (Table 5.4). The model is a 2 layer model with a gently dipping basement surface at around 0 km a.s.l., above which is the volcanics layer. The gravity model of Miller and Williams-Jones (2016) extends only 5 km north of the Te Maari vent, so we use infinite elements to mathematically extend the 10 x 10 x 6 km model volume to ensure boundary effects at the edges of the model do not affect the results of the model interior. The sill source is modelled as a rectangular discontinuity surface by means of ‘pair elements’. The elements are added in pairs along the surface rupture, where a normal tensile dislocation is assigned (Currenti et al., 2008). We then calculate the deformation and gravity response of this numerical model and compare it to the NSGA-II analytic solutions.

Using the formulation of Currenti (2014) we calculate gravity changes $\Delta g$ caused by 3 internal density changes to investigate the effect of the heterogeneities and topography on the gravity models:

$$\Delta \rho(x, y, z) = -\mu \cdot \nabla \rho_0 + \Delta \rho_m - \rho_0 \nabla \cdot \mu$$  \hspace{1cm} (5.4)

where $\rho_0$ is the medium density, and $\rho_m$ is the density change inside the ellipsoid and $\mu$ is the displacement vector. The first term produces gravity change $\Delta g_1$ and accounts for displacement of density boundaries originating from the free surface, density discontinuities and contraction of the source wall. The second term produces gravity change $\Delta g_2$ and is caused by the change of mass in the volume. The third term produces gravity changes arising from the compressibility of the surrounding medium. The observed gravity change $\Delta g$ is the sum of these contributions:

$$\Delta g = \Delta g_1 + \Delta g_2 + \Delta g_3$$  \hspace{1cm} (5.5)
The sum of the $\Delta g_1$ and $\Delta g_3$ contributions in our sill model is maximum $-5 \mu \text{Gal}$, around benchmarks TGM12 and TGM13, indicating $\Delta g_1$ and $\Delta g_3$ make a minimal contribution to the observed gravity, and can be ignored in the analytic gravity only solutions. The displacement RMS of the numerical forward model is 7mm, compared to 6mm in the NSGA-II model.

The small subsidence signal, topography and crust heterogeneities contribute only minor gravity changes that are within the noise level of our measurements. As such, there is no advantage gained by inverting the displacement and gravity data in a computationally expensive numerical model.

5.7 Discussion

The Tongariro hydrothermal system is described as vapour dominated (Hochstein and Bromley, 1979) and DC resistivity and MT measurements (Walsh et al., 1998; Hill et al., 2015) map a thick condensate layer a few hundred metres below the surface, overlying the vapour dominated zone (Figure 5.9). Outflows of this condensate are observed at Ketetahi springs 3km to the west of Te Maari, but no outflows are observed at Te Maari and the two systems are considered to be separate. The shallow gravity and deformation sources, and low density changes modelled, suggest processes occurring in the condensate layer, or the vadose zone above this, are responsible for the observed changes. Our model analysis suggests that separate gravity and deformation sources are required to adequately explain the observed data. The analytic models presented are over simplifications of the likely situation as they neglect the role of fluids and pore space.

Below we discuss a generalised conceptual model that accounts for the observed subsidence and gravity increases, in terms of fluid fluxes into and out of a hydrothermal system, driven by temperature and pressure changes post eruption. In a porous reservoir, liquids absorb stress, creating a fluid pressure or hydraulic head. When a reservoir seal is pierced by an eruption, the fluid pressure will be significantly lowered, reducing its ability to support the overlying rock, causing subsidence above the reservoir. Any process that causes fluid withdrawal increases the effective stress in the host rock, and subsidence then happens by compaction, following Biot poroelasticity theory (Wang, 2017). The hydrothermally altered rock within the reservoir has reduced capability to support itself and requires the fluid pressure for support, compounding the effect of fluid withdrawal.

We propose a scenario where generally deeper sourced subsidence is caused by matrix compaction (causing a reduction in pore space as the grains move together), from pressure driven fluid loss, some of which occurs as vapour exiting the fumaroles and vent, while the mass increase is shallowly sourced, and driven by a mix of buoyant upward migration of expelled pore fluid from compaction, influx of meteoric fluids through the broken seal, vapour condensation, and liquid density changes related to a cooling system (Figures 5.9 and
Figure 5.9: Schematic cross section showing locations of gravity and deformation sources in relation to the hydrothermal system. The cross section shows a region of higher permeability rock (orange shading) above the gravity source allowing greater influx of meteoric fluids. Red shading above the dyke shows an area of higher temperature, where the liquid phase has been pushed out immediately after the eruption. Over time this region contracts (red arrows) as the dyke cools, allowing the liquid condensate to return, adding mass to the system (black arrows). Blue arrows on the gravity and subsidence sources show the direction of movement of these sources over time. The depth of the dyke intrusion is inferred from seismicity (Hurst et al., 2014) and the buoyancy boundary created by the large density contrast between greywacke basement and overlying volcanics. Greywacke basement and extent of altered vs unaltered volcanic rock from Miller and Williams-Jones (2016).
Table 5.4: Summary of physical properties used in finite element model.

<table>
<thead>
<tr>
<th>Layer</th>
<th>Depth m a.s.l.</th>
<th>Density kg/m³</th>
<th>Poisson’s Ratio</th>
<th>Young’s Modulus Pa</th>
<th>Porosity</th>
<th>Permeability m²</th>
</tr>
</thead>
<tbody>
<tr>
<td>Volcanics</td>
<td>&gt;0</td>
<td>2200</td>
<td>0.25</td>
<td>2 x 10⁹</td>
<td>0.2</td>
<td>1 x 10⁻¹⁵</td>
</tr>
<tr>
<td>Basement</td>
<td>&lt;0</td>
<td>2670</td>
<td>0.15</td>
<td>12 x 10⁹</td>
<td>0.03</td>
<td>1 x 10⁻¹⁷</td>
</tr>
</tbody>
</table>

5.10. We discuss the contribution of each of these mechanisms in the following sections. Table 5.5 summarises mechanisms in operation, and whether they are likely to result in gravity increases or decreases, accompanied by inflation or deflation.

5.7.1 Spatial evolution of sources over time

The preferred geometry and depth of the gravity sources varies over time (Figure 5.5). In the first two intervals, more rectangular (i.e sill) shapes are preferred, based on reduced chi-square testing of model results for each geometry. The sphere is the best fit of the non-rectangular geometries. In the last interval, there is a strong preference for the sphere model over the other geometries, highlighted by the RMS difference of 10 µGal (sill), compared to 3 µGal (sphere). Gravity sources generally become shallower from the first, to second and third time intervals; however, there is some overlap in possible depth ranges. Mass varies within a single order of magnitude and the lateral location of the source stays approximately constant.

By contrast, the deformation source becomes progressively shallower over time, from 500 m depth in the first interval, to 178 m depth in the third time interval, using the model of Hamling et al. (2016). The deformation source also becomes narrower and shorter over time, decreasing from 1100 m width to ~340 m, and 1500 m length to ~600 m between the first and third intervals.

In the 2014 to 2015 interval, the gravity and deformation sources laterally overlap, however the deformation source is much broader. In the later intervals, the gravity source location remains centred on the vent, but the deformation source moves 400 to 500 m north east, towards Lower Te Maari Crater. The spatial separation of deformation and gravity sources is highlighted by the joint gravity and deformation inversions. The set of optimal solutions generated by the NSGA-II algorithm is diffuse, and no ‘best fit’ solution exists. Rather, the best gravity source and best deformation source are distinctly separate in the Pareto front.

We explain the source separation as reflecting variations in the rock physical properties within the reservoir. Fluid is drawn from pores in different parts of the reservoir based on permeability, porosity and pressure gradients, where deeper, higher pressure pore space is evacuated ahead of shallow, lower pressure pores (see section 5.7.2 Pore Pressure Changes and Compaction). Over time, different portions of the reservoir compact preferentially,
dependent on the rock strength, degree of alteration and porosity. Initially pores close to the vent are evacuated, but the zone of evacuation migrates away from the vent as the initially evacuated pores come into pressure equilibrium with the vent region or are fully compacted with all fluid expelled. Fluid is then drawn from more distal pores, changing the spatial pattern of subsidence to further from the vent, as is often seen in producing geothermal systems (Narasimhan and Goyal, 1982), where subsidence can occur some distance from the production wells.

5.7.2 Mass balances

Using the gravity only sill model we calculate a mass change rate of 34 kg/s from the mass increase of $1.01 \times 10^9$ kg in 344 days between the 2014 and 2015 observations. The mass flux decreased to 20 kg/s in the 2015 to 2016 interval and increased again to 62 kg/s in the January to December 2016 interval, or 83 kg/s using the better fitting sphere source. The density changes from the sphere gravity models increase from 65 kg/m$^3$ to 146 kg/m$^3$ throughout the study period, reflecting a cooling system. If the fluid is water with a density of 1000 kg/m$^3$, then filling $\sim 6$ to 15 % void space would account for the observed gravity changes. Montanaro et al. (2016) measured porosity of hydrothermally altered rocks from Te Maari of 7 - 28 %, indicating that void space is only partially saturated by incoming fluids, and that the reservoir is not fully pressurised, allowing subsidence to continue, while mass is added.

Observed gravity changes in a hydrothermal system are the result of a combination of processes that contribute mass gains and losses to the system, through the filling or emptying of pore space, with vapour or liquid fluids. The observed mass change is the net of gains - losses. Gains include inputs from redistribution of expelled pore fluids, rainfall infiltration, condensation within the system, and density changes caused by pressure and temperature effects, while losses include fumarole emissions, changes in density due to pressure loss and pore compaction (Table 5.5). We estimate the relative contributions of each to the observed gravity changes.

Pore pressure changes and compaction

Thermo-poro-elastic numerical models of hydrothermal systems (e.g Rinaldi et al., 2011; Fournier and Chardot, 2012; Currenti and Napoli, 2017) usually simulate the injection of new fluids into a hydrothermal system, as a cause of volcanic unrest. These models show that as pore pressure increases, ground inflation occurs soon after, followed by a later, thermally derived inflation signal. Initially the deformation is located close to the fluid injection site, and later becomes more dispersed. The converse situation is true for fluid withdrawal causing depressurisation. As observed at Te Maari, initially the deflation is located symmetrically around the vent and fumaroles, but later shifts away from the vent.
The effect of fluid injection or withdrawal is to change the effective stress \( \sigma' \) acting on the rock matrix. In unconsolidated materials such as soils or volcanic deposits, on a horizontal plane, the effective stress, \( \sigma' \), can be expressed as:

\[
\sigma' = \sigma - \alpha P \tag{5.6}
\]

where \( \sigma \) is the total stress, \( \alpha \) is the Biot coefficient, \( \alpha = 1 - (K_d/K_o) \), where \( K_d \) is the rock bulk modulus and \( K_o \) is the mineral bulk modulus (typically \( \alpha = 1 \) in unconsolidated material (Terzaghi, 1943)) and \( P \) is the pore-fluid pressure (Jousset et al., 2000). In hydrothermal systems the effective stress, is usually increased by removing fluid and lowering the pore-fluid pressure. The increase in effective stress causes matrix compaction, reducing the pore size and promoting subsidence. The 2012 eruption caused a pore-fluid pressure drop in the reservoir, raising the effective stress and initiating matrix contraction which later develop to compaction as fluid is drained from the pores. After the initial pressure drop, deeper pores under greater hydrostatic pressure are evacuated ahead of shallower pores, because of the greater pressure gradient between the pore and the now open vent conditions. Subsidence begins in the deeper pores and migrates upwards, as deeper pores evacuate and then compact (Figure 5.10). Decreasing the pore pressure lowers the density of the fluid and causes a gravity decrease. Hamling et al. (2016) calculated a pressure drop of 0.09 MPa / year, which for a 1000 m thick aquifer causes only 1 µGal gravity decrease from the pressure associated density change (e.g. Allis and Hunt, 1986, eqn 12). The effect of pore pressure decrease on gravity changes is therefore negligible.

Jousset et al. (2000) observed subsidence at Mount Komagatake, Japan, and derived a model where loss of pore fluid occurred through evaporation following heating caused by deep magma injection. They calculated a 4 m water drop in a 200 m thick porous layer would produce 5 cm subsidence after 1 year. The subsidence rates, rock properties and source depths at Mount Komagatake are similar to those at Te Maari. If the 4 m water drop occurs in a layer with 0.15 porosity (as required to explain the observed mass increases), then theoretical gravity changes from an infinite Bouguer slab, \( \Delta g = 2\pi \rho G \Delta h \phi \) approximation, will be \( \sim 25 \mu \text{Gal} \). If this expelled pore fluid remains in the system (there is no observed liquid outflow at Te Maari) and migrates upwards due to buoyancy from its decreased density, then it will add to the shallow gravity signal. The fluid then also cools and becomes more dense, contributing to the observed gravity signal.

When the fluid is displaced vertically, the deeper sourced gravity decrease from the pore fluid loss will be masked by the shallower mass addition, but when the fluids are displaced laterally a negative gravity signal should be seen where fluid has been expelled. In the 2015 to 2017 time interval, there is some evidence for the development of a negative gravity change in a similar position to the centre of the subsidence source (Figure 5.2),
Saturated
Unsaturated
T0
< 2012
Vadose
Condensate
Vapour

System at equilibrium.
Sealed hydrothermal system with pressurised condensate layer.

Condensate seal broken.
System pressure drops.
Condensate pores evacuated and matrix around the vent starts compacting.
Pore fluid redistributed from deeper pores.

P atmospheric
Meteroric water influx

Gravity source
Deformation source vertically below gravity source.

Porous condensation

Precipitation infiltration adds shallow mass in vadose zone or top of condensate.
Pore fluid redistributed from deeper pores.
Previously compacted pores (2014 to 2015 interval).

Gravity sources
- Precipitation
- Condensation
- or pore fluid migration

Deformation sources
- Actively evacuating pores
- Previously compacted pores

Figure 5.10: Conceptual evolution of the gravity and deformation sources after the August 2012 eruption. Open blue ellipses represent evacuated pores undergoing compaction, while narrower ellipses are already compacted pores. Light blue coloured circles are pores filled by precipitation infiltration, while darker blue circles are pore fluid expelled from deeper pores or condensation. The solid or dashed thick pink line represents the intact or broken condensate seal, respectively.
which may reflect the pore fluid expulsion. Migration of pore fluid therefore may account for a significant proportion of the observed signal.

**Precipitation recharge**

The shallow gravity change source suggests that groundwater changes (e.g. Hemnings et al., 2016) might be responsible for a portion of the observed mass increase. Jolly et al. (2012) suggested melting snow percolating into the hydrothermal system beneath neighbouring Mt Ngauruhoe, as a mechanism for generating low frequency volcanic earthquakes recorded between 2005 and 2009. Vapour dominated systems such as Te Maari, typically require a long-lived heat source at depth, and long lived, low permeability boundaries to seal the entire system and prevent flooding by adjacent ground water (Allis, 2000). The Te Maari system is likely to have sealed in the 125 years following the previous eruptions, as evidenced by very low fumarolic activity prior to the 2012 eruption, but that seal would have been broken by the 2012 eruptions, allowing vertical and lateral ingress of meteoric fluids. Rainfall records from a weather station approximately 20 km to the south, on Mt Ruapehu, show an average annual rainfall of approximately 2 m. Additionally, in the winter months approximately 2 m of snow falls, which may represent an additional 20–50 cm of water. Assuming a total precipitation of 2.5 m and a recharge rate of 20 % precipitation, covering the catchment area around the Te Maari vent of 2 km$^2$, a total mass of $1 \times 10^9$ kg meteoric water is added to the system per year. This is the same order of magnitude as the observed mass changes and suggests increasing saturation of the either the vadose zone or the top of the condensate layer. If we consider a smaller region of rainfall recharge, limited to the area directly above the source models (approximately 500 m x 1000 m), then using the same rainfall and recharge rates, we calculate $2.5 \times 10^8$ kg of mass added to the system per year. This is around 1 order of magnitude smaller than the observed mass addition.

If precipitation infiltration into the hydrothermal system is widespread on the volcano, then we would expect a spatially broader gravity signal. Instead the gravity anomaly is only around the vent region, suggesting that increased fracturing around the vent allows additional infiltration. Precipitation ingress likely represents an important contribution to the observed mass increases.

**Condensate formation**

De-pressurisation of the hydrothermal system following eruption increases condensate production as the vapour pressure and temperature drop. Using a conservation of mass argument at neighbouring Ketetahi springs, Walsh et al. (1998) suggested a condensation of vapour rate of 6.3 kg/s, occurring at the base of the condensate layer / top of the vapour layer. At this rate, condensation would add approximately $1.87 \times 10^8$ kg per year. This condensation rate is for a system in steady state, so is likely to be higher for a system that has been actively decompressed, increasing condensate formation. De-pressurisation of the
system will produce a pressure gradient within the condensate layer, promoting the lateral movement of fluid into pore space previously evacuated. The location or rate of condensate formation, however, is not sufficient to refill pores emptied by the eruption induced pressure drop, and does not prevent subsidence. Condensation is likely a smaller contributor to the observed gravity changes than pore fluid transfer and groundwater influx.

**Temperature changes**

We propose the shallow (1.5 to 2 km) emplacement of a high temperature dyke (Hurst et al., 2014) and subsequent eruption through the hydrothermal system changed location of the vapour - condensate boundary beneath Te Maari, and created a new vapour zone in a radius around the vent region, similar to the vapour chimney proposed by Reyes et al. (1993) for volcanic hydrothermal systems. At the start of the gravity measurements, the temperature of the intruded dyke is well below the solidus, so the input of magmatic fluids is likely to be minimal and the effects of the dyke are limited to adding heat to the system. The temperature of fumarole F3 decreased from 415 °C in September 2012 to 309 °C in November 2016 (Figure 5.11). The effect of dyke cooling will be two fold. Firstly the vapour chimney around the conduit will decrease in radius, allowing condensate fluid to migrate closer to the vent, and secondly, the temperature of the condensate outside the vapour region will cool, increasing in density. Both effects result in gravity increases.
Most of the F3 temperature trend can be fitted by a 1D conductive cooling curve (Turchette and Schubert, 2014); however, the most recent measurement is well below that curve suggesting additional cooling from convection or advective heat transfer, possibly caused by influx of water, either meteoric or expelled pore fluid, into the condensate layer or shallow subsurface around the vent.

Gravity changes caused by liquid density changes can be approximated by $\Delta g = 0.015 \Delta T h$ (e.g. Allis and Hunt, 1986, eqn 10) for liquids between 180 and 230 °C and porosity 0.3. A 100 m thick condensate layer cooling by 10 °C will result in a 15 µGal gravity increase. The phase transformation associated with condensation requires absorption of the latent heat of vaporisation and results in cooling of the surrounding fluid. While we do not know the temperature change of the condensate, it has likely cooled in concert with the dyke, so it is possible that a measurable portion of the gravity changes are due to temperature effects of the cooling dyke, increasing the density of the condensate fluid.

**Fumarole emissions**

We calculate a rough mass loss from fumarole emissions, using sparse emission rate measurements of CO₂ and SO₂ since the eruption, of 10 and <1 tonnes/day, respectively (GeoNet website, www.geonet.org.nz, accessed June 2017). These measurements indicate that emission of these gases contributes relatively minor amounts of mass flux (0.1 and 0.01 kg/s). These gases are likely deeper sourced from the remanents of dyke intrusion. There are no direct measurements of water vapour loss; however, we estimate the steam flow rate from visual observations of the main fumarole, F3, and calculate an approximate flux rate on the order of 10 kg/s, assuming a density of steam at 100 °C of 0.6 kg/m³, a fumarole diameter of 2 m and a flow rate of around 5 m/s. By comparison, the H₂O emission rate at the actively degassing White Island volcano, at the northern end of the TVZ was estimated by Werner et al. (2008) to be 2600 tonnes/day (30 kg/s) in 2008. White Island is a much larger and more active system than Te Maari, so our estimate of fumarolic mass losses at Te Maari is probably approximately correct. Our estimate is similar to values calculated by Walsh et al. (1998) for fumaroles at Ketetahi (3 to 8 kg/s), 3km to the west, and is consistent with typical H₂O:SO₂ flux ratios of 100:1 typically reported (Giggenbach, 1996).

In summary, relatively minor mass loss from fumarole emissions (~10 kg/s) is exceeded by mass addition (20–60 kg/s) from the migration of expelled pore fluid, meteoric fluid influx through the broken hydrothermal system seal, condensation of cooling vapour around the vent, and increase in density of cooling condensate. The deviation of the F3 temperature trend below the conduction cooling curve, coincides with the change in preferred source type to a sphere, and more rapid mass flux rates, and likely indicates a change to a convective or advective cooling regime, driven by increased water influx into the reservoir around the fumarole. Precise estimation of the contributions of each mechanism require more detailed numerical modelling and will be the focus of future studies.
Gravity increases
- Liquid infiltration / increases in saturation
- Liquid cooling, increasing density
- Vapour to liquid condensation
- Increased pore pressure and liquid density
- Mineralisation

Associated deformation change
- Inflation
- None or deflation
- None or inflation
- Inflation
- None, but may slow subsidence rates

Gravity decreases
- Liquid withdrawal / changes in saturation
- Liquid heating, decreasing density
- Liquid to vapour conversion
- Decreased pore pressure and liquid density
- Pore compaction

Table 5.5: Summary of processes occurring in hydrothermal system and if they are likely to result in gravity increases or decreases and inflation or deflation.

5.7.3 Reservoir hydraulic conductivity and permeability estimates

From our gravity change models we know the amount of liquid added to the hydrothermal reservoir and at what depth, i.e. the hydraulic head. From Darcy’s Law and the sill models, we calculate the vertical hydraulic conductivity ($K_z$ in m/s) of the reservoir for each interval:

$$K_z = -\frac{Q}{A} \left[\frac{l}{h}\right]$$

(5.7)

where $Q$ is the volumetric flow rate ($m^3/s$), derived from the gravity models, $A$ ($m^2$) is the cross sectional area, $l$ is the flow path length (m) and $h$ is the head (m) (See Figure 5.7). A is determined from the cross sectional area of the sill model (see Figure 5.7), $l$ is taken as sill thickness, calculated from the UM and a water density of 1000 kg/m$^3$. The water density is a maximum value and geothermal fluid will have a lower density resulting in $K_z$ being a maximum value. For $h$, we use depth of the source converted to head. $K_z$ varies from $8.73 \times 10^{-10}$ in the first interval to $2.55 \times 10^{-9}$ in the second and $5.68 \times 10^{-9}$ m/s in the 3rd measurement interval.

The relationship between intrinsic permeability ($k$, in $m^2$) and hydraulic conductivity ($K$) was defined by Hubbert (1956) as:

$$K = k \frac{\rho_w g}{\mu}$$

(5.8)

where $\rho_w$ is the density of water (1000 kg/m$^3$), $g$ is gravitational acceleration (9.8 m/s$^2$) and $\mu$ is dynamic viscosity of water ($1 \times 10^{-3}$ kg/m.s at 20 °C). Solving for $k$, we obtain intrinsic permeabilities of $8.58 \times 10^{-15}$ m$^2$ for the 2014 to 2015 interval, $2.51 \times 10^{-14}$ m$^2$ between 2015 and 2016, and $5.58 \times 10^{-14}$ m$^2$ between January 2016 and December 2016, for our range of $K_z$. Our permeabilities are similar to the $9 \times 10^{-15}$ m$^2$ values estimated by
Walsh et al. (1998) for permeability of the condensate layer. These values are well below the minimum permeability calculated by Schubert et al. (1980) \((4 \times 10^{-8} \text{ m}^2)\), as necessary to maintain a gravitationally stable condensate layer, above a vapour layer. Compared to permeabilities and conductivities calculated using a similar method by Miller et al. (2017a) for fluid intrusion into a magma stressed fault zone at Laguna del Maule, Chile, the Te Maari hydrothermal reservoir values are lower and may reflect high clay content caused by hydrothermal alteration.

5.8 Conclusions

Gravity increases accompanied by subsidence as recorded at Te Maari crater are unusual, and require a combination of processes to explain our observations. A multiple parameter dataset is required to distinguish processes causing subsidence from those causing mass addition. Here the combination of microgravity (even with a limited dataset) and deformation modelling revealed processes that either method on its own would not discover, and provided a more complete picture of the hydrothermal system dynamics. Multi-objective inversion fully tests if the deformation and mass change sources are coincident, and is important to help guide an accurate interpretation of the datasets. Finally, finite element modelling incorporates existing physical property knowledge (from previous Bouguer gravity surveys) to build a comprehensive model testing the validity of the analytic solutions.

We propose a model of eruption initiated, de-pressurisation driven fluid transfer, producing subsidence and shallow mass addition. The eruption through the previously sealed hydrothermal system caused a decrease in fluid pore pressure in the shallow condensate layer. Initial matrix contraction and subsequent drainage of pores promotes subsidence which migrates both laterally and vertically over time, exploiting changing pressure gradients between the condensate layer and the atmosphere. As fluid is expelled from deeper pores it depressurises, becomes less dense, and rises where it accumulates at its new level of buoyancy, or is trapped beneath still sealed parts of the system. The fluid then cools and becomes more dense, contributing to the observed gravity signal. The resulting shallow mass increase is recorded as positive gravity changes. Increased fracturing around the vent, also allows greater meteoric recharge into the condensate layer, contributing to the observed mass addition. Pore fluid redistribution and meteoric water infiltration are likely the dominant two mass addition processes. Gravity increases are also caused by cooling of the system following dyke injection, increasing conversion of vapour to condensate, while also increasing the density of existing condensate as the vapour phase around fumarole F3 contracts. Later, more rapid cooling of the hydrothermal system around the vent is shown by departure of fumarole temperatures from conductive cooling trends, reflecting convective cooling aided by increased liquid phase refilling pore space. Mass gains to the system are greater than relatively minor losses from fumarole emissions.
The maintenance of a vapour dominated system in a volcanic environment requires rapid resealing through mineralisation, to prevent flooding of the system by meteoric groundwater. As observed at Raoul Island (Christenson et al., 2007), sealed hydrothermal systems present eruption hazards, so joint microgravity and deformation monitoring can give information on the time scale of the resealing processes and help with long term hazard assessment. As such it compliments the standard suite of volcano monitoring techniques (e.g. Miller and Jolly, 2014) used at volcanoes with hydrothermal systems. By analysing gravity and deformation changes, we can consider whether the system has become sealed, and may be pressurising in the lead up to a future eruption. Until the reservoir seal is re-established, subsidence is likely to continue, along with gravity changes from the redistribution of mass from within the system, and addition of new mass from outside. Mineralisation of fractures is the most likely cause of resealing, however this process operates on timescales of years to decades so long term subsidence and mass monitoring is required. Continuing microgravity and deformation measurements will allow us to monitor the long-term hazard the Tongariro hydrothermal system presents, while future studies using thermo-poroelastic numerical models, constrained by the analytic models, will test the conceptual framework presented here, and expand our understanding of the evolution of the system, post eruption.

5.9 Acknowledgements

Thank you to Department of Conservation with support from Harry Keys, for permission to undertake the gravity surveys in Tongariro National Park. Thank you to Sigrun Hreinsdottir for processing the GPS results and to Bruce Christenson for discussions about the hydrothermal system at Te Maari. Thank you to field assistants, Nellie Olsen, Lauriane Chardot, Natalia Deligne, Tom Ayling, and Alex Kmoch. F3 temperature measurements and gas flux data are from GeoNet with thanks to Karen Britten. CM is funded by the Earthquake Commission, GNS Science core funding, and Mitacs Accelerate, Canada. Most of the figures were made using python and matplotlib (Hunter, 2007).

5.10 References


Terzaghi, K., 1943. Theoretical soil mechanics. volume 18. Wiley Online Library.


Chapter 6

Conclusions

In this thesis I examine two contrasting volcanic systems, Laguna del Maule volcanic field (LdMVF), Chile, and Mt Tongariro volcanic massif (TgVM), New Zealand, and illustrate how knowledge of volcanic architecture is critical to understanding the volcanic unrest occurring at those volcanoes. At each volcano I use a range of potential-field geophysical methods, and several modelling approaches, to elucidate volcanic architecture and active, time varying processes that occur within that structure. To conclude this thesis, I summarise and extend the key findings of the four papers presented here, with reference to the research objectives outlined in Chapter 1. Finally, I discuss the broader scientific implications of my work, and suggest directions for future research.

6.1 Conclusions to Research Objectives

The two main goals of the research objectives are firstly to define the internal structure of volcanoes, by modelling the 3D distribution of physical properties such as density and magnetic susceptibility, as measured by gravity and magnetic surveys. The second goal is to incorporate the knowledge of that internal structure into models and interpretations of unrest processes. Where possible, internal structure is incorporated directly as physical property parameter distributions in the time varying models, or if this is not possible, indirectly, by guiding the interpretation of the active processes through better understanding the context within which they occur. Chapters two and four address the first theme, and Chapters three and five, the second. In the preamble of each chapter I posed specific hypotheses, that the associated chapter developed and tested within the framework of the broader science questions (Table 1.1). In Table 6.1 I resolve the hypotheses with brief answers.

In Chapter 2, I use Bouguer gravity measurements and a novel 3D inversion routine to image a shallow, volatile rich, crystal poor, 30 km$^3$ magma system, beneath Laguna del Maule. This achieves the first research objective to ‘Understand the current configuration of the magma reservoir at the LdMVF’. The interpretation of the gravity model, leading
Laguna del Maule hypotheses

<table>
<thead>
<tr>
<th>Hypothesis</th>
<th>Outcome</th>
</tr>
</thead>
<tbody>
<tr>
<td>There is a silicic magma body that underlies the LdMVF.</td>
<td>True. The body extends between 2 and 5 km below the lake.</td>
</tr>
<tr>
<td>An inflating sill at 5km depth is the only active magma body at LdMVF.</td>
<td>False. There is a 30km$^3$ silicic body above the sill, within a larger 115km$^3$ mush reservoir.</td>
</tr>
<tr>
<td>The present-day magma body under LdMVF is not large enough to have been the source of all Holocene eruptions at LdMVF.</td>
<td>True. The melt area is likely to have migrated within the mush over time.</td>
</tr>
<tr>
<td>The present-day magma body under LdMVF is not overpressurised.</td>
<td>True. Lithostatic pressure exceeds buoyancy by $\sim$3x.</td>
</tr>
<tr>
<td>Microgravity changes through time at LdMVF can be explained by the deformation changes observed by InSAR.</td>
<td>False. The source of deformation is not the source of mass changes.</td>
</tr>
<tr>
<td>Microgravity changes through time at LdMVF can be explained by mass changes within the shallow hydrothermal system.</td>
<td>True. Mass changes are located at 1 to 1.5km depth, along the Troncoso fault zone.</td>
</tr>
<tr>
<td>Sill opening at depth changes the stress field around regional scale faults at LdMVF, allowing fluids to migrate in.</td>
<td>True. Coulomb and mean stress changes along the Troncoso fault induce fluid movement.</td>
</tr>
</tbody>
</table>

Mt Tongariro hypotheses

<table>
<thead>
<tr>
<th>Hypothesis</th>
<th>Outcome</th>
</tr>
</thead>
<tbody>
<tr>
<td>Greywacke basement beneath Mt Tongariro is continuous.</td>
<td>True. The gravity model shows subtle density differences between the two greywacke terranes beneath TgVM.</td>
</tr>
<tr>
<td>A broad region of hydrothermally altered rock underlies the TgVM, and is thus a site for potential future phreatic eruptions.</td>
<td>True. This is delineated by low density, and demagnetised rock, and must be long-lived ($10^4$ to $10^5$ years).</td>
</tr>
<tr>
<td>Fault movement can explain the magnitude of subsidence in the Ruapehu graben since 275ka. Crustal flexure is not needed to accommodate the amount of observed subsidence in this region.</td>
<td>True. Modelled fault offsets are consistent with geologically determined offsets.</td>
</tr>
<tr>
<td>The Waihi and Poutu basement faults do not act as magma ascent pathways in the TgVM.</td>
<td>True. Within the resolution of the models there is no evidence for magma intrusion along these faults.</td>
</tr>
<tr>
<td>Magma rises to the surface in the TgVM in small, discrete batches with little to no shallow storage within the edifice.</td>
<td>True. Within the resolution of the models there is no evidence for magma intrusion at shallow depth.</td>
</tr>
<tr>
<td>The volume of erupted material is larger than previous estimates.</td>
<td>True. Five to six times larger than previous geological estimates.</td>
</tr>
<tr>
<td>The same source and physical process can explain both the measured deformation and gravity change at Upper Te Maari crater, post-2012 eruption.</td>
<td>False. The deflation source is deeper than the mass addition source.</td>
</tr>
<tr>
<td>More than one source is responsible for the observed gravity changes at Upper Te Maari crater.</td>
<td>True. Rainfall recharge, deep to shallow fluid migration, condensate formation and temperature changes all contribute to the gravity signal.</td>
</tr>
</tbody>
</table>

Table 6.1: Summary of hypotheses developed and tested in this thesis, with brief outcome description.
to the conclusion of a dominantly liquid magma body, is driven by thermodynamic models derived from the chemical composition of erupted lavas. The thermodynamic model gives the critical insight that the low density body modelled by gravity is likely composed of a high liquid-proportion magma, and contains a significant volatile component. I show the eruptible magma system overlies the source of previously reported deformation (Feigl et al., 2014) and sits within a broader (115 km$^3$) region of high crystal-content magma mush. This model provides a rare geophysical image of a rhyolitic magma reservoir, and is consistent with cutting-edge petrologic models that suggest liquid-dominated magma bodies may be maintained within a larger region of crystal mush (e.g. Parmigiani et al., 2016; Cashman et al., 2017; Rubin et al., 2017).

To explain vent locations in the last 25 ka, I use this magma reservoir image, and simple dyke propagation models, to conclude that the active melt-rich portion of the reservoir must have been either bigger, or has reformed elsewhere within the crystal mush over time, or that significant (km) lateral transport of magma occurred from the present day magma location. Using a simple 1D elastic model approximation I propose that in its current state, the magma system is not critically overpressured, and requires either further volatile input from crystallisation or from deeper mafic volatile influx, for buoyancy to become the driving eruptive force. Alternatively, the sudden release of volatiles, driven by an external source, such as earthquake shaking on nearby faults may be required. External shaking releasing volatiles is implicated as an eruption trigger mechanism at other volcanoes, (e.g. Christenson et al., 2007). In addition, the intrusion of large volumes of magma in the last 10 ka as evidenced from the tilted shoreline, are likely to have thermally conditioned the surrounding crust into a state where small volumes (10-100 km$^3$) of magma remain stored rather than erupted. These initial conclusions are important for managing the public response to the well known, rapid uplift, and have been communicated to Sernageomin, Chile, and discussed with ONEMI (Chilean civil protection agency). However more detailed modelling incorporating non-elastic crustal rheologies and thermal priming of the crust, as well as determining the local geotherm beneath the lake are required for a more comprehensive view of this important issue.

In Chapter 3, I measured gravity changes over a 4 year period, and put forward the hypothesis that observed gravity increases are directly caused by the injection of magma, modelled as responsible for the large scale uplift by Le Mével et al. (2016). However, I conclude this hypothesis is incorrect, and a shallower interplay between magma injection and local fault structures, not evident in the deformation model, is at work.

I propose a mechanism whereby the injection of magma at depth, changes the stress and strain fields around nearby faults, lowering pore pressure within them, and facilitating the migration of hydrothermal fluid into existing and newly created pore space. The fluid injection in turn creates the observed gravity changes as it adds mass to the system. This model shows that the deformation and gravity change sources are spatially separate, but
linked by the mechanism of stress transfer, driving permeability changes. Gravity change measurements are the only way of detecting this fluid flow, and provide unique insight into how magma intrusion interacts with regional fault structures. Without the detailed knowledge gained in the first chapter, showing the location of a magma reservoir adjacent to a regional scale fault system, interpreting the gravity change results is more challenging. Instead, the combination of volcano structure, and time varying models, provides greater insight than possible from either alone. These conclusions achieve the second objective to ‘Understand the causes of magmatic and/or hydrothermal unrest’ at the LdMVF.

Additional evidence for the past interplay of magma and faults at LdMVF is presented in Appendix C. Here I briefly describe the results of a magnetic survey on the lake, showing that remanently magnetised dykes are coincident with lake bed faults, mapped by seismic reflection. I conclude these faults are active, long-lived structures, and played important roles in accommodating past magma injection, as well as being the focus of fluid injection in the present day.

In Chapter 4, I use a combination of gravity and magnetic data, constrained by mapped geology, to build a geologically constrained, geophysical model of the Mt Tongariro volcanic massif. The gravity model outlines the basement structure beneath Mt Tongariro, resulting in a new, larger estimate of the volume of erupted material, and hence a new estimate on the long-term magma supply rate. The location of the boundary between dense, low permeability greywacke, and higher permeability, low density volcanics, is important for interpretation of many geophysical data sets. The contact creates a strong physical property contrast, at fairly shallow depth, that violates many of the assumptions of a homogeneous halfspace often used in models of active processes. Unlike at LdMVF, no evidence for shallow magma systems is observed, and major bounding faults at Mt Tongariro do not appear to be major conduits for fluid movement. In fact, as the magnetic model shows, they may be relatively impermeable, and act to confine the hydrothermal system. The striking feature of the magnetic model is a large region of demagnetised rock that I interpret as representing the hydrothermal system. Circulation of hot, acid fluids acts to physically and chemically alter the host rock, dissolves magnetite, and produces an obvious ‘burnhole’ in the magnetic data. The combinations of low density and low magnetisation models successfully outlines the hydrothermal system, and answers the 3rd research objective to ‘Define Mt Tongariro volcano and basement structure, and hydrothermal system extent’. I propose that the 3D gravity model be converted to a new high resolution 3D seismic velocity model enabling more accurate earthquake location, especially for shallow earthquakes that may precede future dyke injections that lead to eruption as occurred in 2012. This will in turn enable more accurate eruption forecasting during times of unrest.

In Chapter 5, time-varying gravity measurements taken after the 2012 Te Maari eruptions show ongoing mass addition, and the continuation of subsidence measured by Hamling et al. (2016). Joint, multi-objective inversion of gravity and deformation changes show that
separate sources are required, and that the mass addition occurs shallower than the subsidence. I propose a model whereby mass addition is driven by a combination of external meteoric fluids, entering through eruption-produced fractures that pierce the hydrothermal system cap, as well as transfer of pore fluids from deep to shallow, and increased condensation and cooling of vapour from the eruption-induced pressure drop. The mass additions outweigh mass lost from steam exiting fumaroles. I conclude that as part of a comprehensive monitoring programme, continued monitoring of gravity and deformation changes will allow progress on the resealing of the hydrothermal system to be judged, helping to determine the likelihood of future steam-overpressure-driven eruptions. The new model of the hydrothermal system completes the last research objective to ‘Understand the response of the Tongariro hydrothermal system to eruptions in 2012’.

6.2 Additional Contributions

In addition to conclusions drawn from the specific research objectives, I describe broader scientific contributions in two themes modelling techniques and the role of fluids.

6.2.1 Modelling techniques

Throughout this thesis I use a variety of modelling techniques to quantify, interpret and understand the observations. Not all techniques are applicable in all cases, but by selective use, the information gained from each dataset can be maximised. The first example is a contribution to volcanology using modelling techniques commonly applied in the mineral exploration and mining industry, but seldom applied in volcanic research. These techniques, shown in Chapter 4, build geologically constrained geophysical models that directly incorporate known geological structure and physical property ranges. In this method, a geological boundary is directly inverted to match the appropriate potential field data, while alternating iterations of geometry and physical property, allow heterogeneities within the unit to be investigated. At Mt Tongariro, this approach produced a well constrained depth-to-basement map that contained a subtle density contrast, consistent with the two different basement terrains outcropping on each side of the central graben. In areas of well mapped geology, this class of model allows detailed images of the subsurface to be developed that retain a high degree of confidence in their geological accuracy.

However, there are many volcanoes where knowledge of the detailed geology is nonexistent, or highly irregular and difficult to model. In this case, I applied 3D inversion techniques to provide realistic simulations of the volcanic structure. For example, in the LdMVF magnetic data (Appendix C) an objective function rotated along the geologic strike, combined with mixed norm regularisation acting on both model values and gradients, better modelled long, narrow dykes than a traditional ‘smooth’, or $L_2$ norm inversion. In the gravity model of the LdMVF magma reservoir (Chapter 2), I am inspired by leading edge
conceptual and petrological models of magma systems (e.g. Cashman et al., 2017), and use a mixed norm inversion (Fournier et al., 2016) to simulate a compact central body (\(L_0\) norm) representing a region of high melt proportion, with a short gradational zone (\(L_1\) norm on the model gradients) representing a crystal mush zone surrounding the reservoir. The mixed norm approach allows greater flexibility for inversion and interpretation than traditional ‘smooth’ (\(L_2\) norm) or ‘compact’ (\(L_0\)) only regularisations, and is highly applicable to studies of volcanoes where physical property boundaries may be either sharp, or gradual, within the same model. To promote this modelling technique among the volcanological community, the source code is available at http://docs.simpeg.xyz/content/examples/index.html#gravity.

Even using sophisticated inversion techniques, interpretation of models of physical property distributions is improved by knowledge gained from other modelling techniques. In the Laguna del Maule study I use the rhyolite-MELTS (Gualda and Ghiorso, 2015) thermodynamic model to obtain densities of magma phases and investigate their impact on the whole system density. This technique extends previous approaches by considering not only melt density, but critically the impact volatiles have on modifying the density of the whole magma system. The method is widely applicable to interpretation of gravity results from other volcanic systems, and may result in significantly different conclusions, compared to if volatile phases are ignored, with important implications for hazard assessment.

Despite new approaches to modelling dense data sets, analytic models are still highly applicable for modelling time-varying processes from microgravity data, mostly because the time intensive nature of data collection results in sparser coverage, compared to more detailed Bouguer gravity data sets. To better parameterise the analytic models, I use a Monte Carlo type approach and run many thousands of simulations seeded with randomised starting parameters, so that the physical property distributions are viewed in a probabilistic sense. Additionally, I improve the interpretation of the time varying analytic models, by viewing them in context of the models that describe the internal structure. This contextual approach reduces the limitations of an isotropic half-space assumption, inherent in analytic models. For silicic volcanoes, topography is often relatively subdued, so the limitation of analytic sources that often assume a flat, or smoothly varying topography are less important.

Deformation data are common at volcanoes through the now near ubiquitous InSAR, but 4D gravity measurements less so. However, as I show in Chapters 3 and 5, deformation measurements may not tell the full story, as they reflect primarily pressure changes, but are not sensitive to mass changes that actually drive the pressure changes. Combining deformation and gravity data in either analytic or finite-element models (Chapter 5) helps constrain both pressure and mass change sources. If distributions of physical properties are available from other models, then finite element models of gravity and deformation that incorporate these, along with topography, can make more realistic simulations, especially in stratovolcanoes, where topography can have an important effect.
At Laguna del Maule, the time-varying gravity model revealed a distinctly spatially separate source to the deformation model, and I linked the two through a stress change model that revealed a previously unknown interaction with local faults (Chapter 3). At Te Maari Crater (Chapter 5), the gravity and deformation sources are much closer together and I used a NSGA-II multi-objective inversion algorithm to assess if the same source could reproduce the gravity and deformation data. From this modelling I concluded that the two sources are separate at better than 95% confidence limit. The two sources are however related, in that fluid movement from deep to shallow accounts for the deeper deformation source and a proportion of shallow mass addition. In both cases combining deformation and gravity data produced a more detailed understanding of active processes than possible from either dataset alone.

### 6.2.2 The importance of fluids

Fluids significantly influence all the models presented in this thesis. At Laguna del Maule, the exsolution of a discrete volatile phase, significantly lowers the density of the magma body. My initial attempts to model the gravity anomaly, without considering the role of volatiles were unsuccessful, as they could not reproduce the low density required to match the observed data. The presence of a free volatile phase in a shallow magma body results in a much different conclusion about the hazard potential that magma body represents, than if only the melt and crystal phases are considered. Likewise at Mt Tongariro, hot and acidic hydrothermal system fluids played a key role in physically altering their host rock, allowing the hydrothermal system to be delineated by gravity and magnetic models. The destruction of magnetite in hydrothermal systems means magnetic surveys are an important tool in quantifying the volumes of altered and weakened rock that may represent a substantial landslide threat.

Fluids not only influence the internal structure of volcanoes, but fluid movement within volcanic edifices produces time-varying potential-field signals. Migration of fluids into a fault zone is responsible for gravity changes observed at Laguna del Maule, while vertical migration of depressurised pore fluids, combined with infiltration of rainwater and condensation of vapour, causes gravity increases at Mt Tongariro. At LdMVF the fluid migration revealed the previously undetected interaction between the ongoing sill intrusion and the nearby Troncoso fault. At Mt Tongariro, the continued monitoring of fluid flux will help determine the state of resealing of the hydrothermal system following eruption, and will help with long term volcanic hazard forecasts of future, often unpredictable phreatic eruptions.

### 6.3 Future Work

I would be interested to see future combined gravity and deformation modelling work at the LdMVF focus on finite-element models. A fluid-flow model, set in a thermo-poroelastic
medium, simulating pore pressure changes from the deformation source, to unify the gravity
and deformation observations in a single model, would have great predictive power for
determining long term outcomes of unrest. This would be very important for more accurately
determining the overpressure state of the magma reservoir. The most recent gravity
observations in January 2017 show no gravity change from the 2016 survey. This coincides
with a change in seismicity pattern (C. Thurber 2017 pers. comm. 12 August 2017) and
may strengthen the argument presented in Chapter 3 that gravity changes are related to
changing stress regimes, influenced by earthquake shaking.

To unify other geophysical observations (from other researchers working at LdMVF),
joint inversions of gravity and seismic data are being planned to determine the sensitivity of
each of those techniques to different physical properties of the magmatic system. This would
create a fuller picture of the volcanic field, than from each method individually. Preliminary
seismic tomography models show a low Vs zone overlapping the Bouguer gravity low. Joint
inversion of the seismic and gravity data will allow better constraints on both Vs and density
and hence likely melt proportions than from gravity or seismology alone. Finally, the shallow
volatile rich magma body may contain a gas-rich cap at its shallowest point that could be
detected by seismic reflection as a high amplitude ‘bright spot’.

At Te Maari, a two-phase fluid-flow model (from Tough2 code), coupled to finite-element
models of gravity and deformation would be a powerful method to simulate changes in the
hydrothermal system reservoir, and anticipate its long-term recovery post-eruption. Further
work into the influence of the deeper magmatic system on the hydrothermal system would
allow a more complete picture of the Tongariro Volcanic Centre to be determined.

Overall, my work showcases how geophysical models of volcanic architecture and time
varying processes, when integrated with ground deformation data and thermodynamic mod-
els, produce greater insight into the causes of volcanic unrest, and reveal previously hidden
processes. Combining traditional potential-field geophysical methods, with state-of-the-art,
multi-discipline, modelling techniques, creates a powerful and effective toolbox for the 21st
century volcanologist!

6.4 References

Cashman, K. V., R. S. J. Sparks, and J. D. Blundy 2017, Vertically extensive and unstable
magmatic systems: A unified view of igneous processes, Science, 355(6331), doi:10.1126/
science.aag3055.

Christenson, B. W., C. A. Werner, A. G. Reyes, S. Sherburn, B. J. Scott, C. Miller, M. J.
Rosenburg, A. W. Hurst, and K. A. Britten 2007, Hazards From Hydrothermally Sealed
1029/2007EO050002.


Parmigiani, A., S. Faroughi, C. Huber, O. Bachmann, and Y. Su 2016, Bubble accumulation and its role in the evolution of magma reservoirs in the upper crust, Nature, 532, 492–495, doi:10.1038/nature17401

Appendix A

Supplementary Material for ‘3D Gravity Inversion and Thermodynamic Modelling Reveal Properties of Shallow Silicic Magma Reservoir Beneath Laguna del Maule, Chile’

A.1 Gravity Measurements

Three different LaCoste and Romberg gravity meters (G127, G943 and EG26) were used in the survey and we inter-calibrate the measurements using two absolute gravity stations MAUL and LDMA (S. Bonvalot, pers. comm.), approximately 30 and 2 km, respectively, from our local base station. These absolute stations also served to tie our local survey to absolute gravity values.

Following the scheme outlined in Hinze et al. (2005), we process the gravity data as follows. We correct for latitude variations using the theoretical gravity value from 1980 GRS Geodetic Reference System (GRS80), Somigliana formula. To correct for elevation differences, we apply an atmospheric correction to account for the changing density of atmosphere with elevation and a free air correction using the full equation with second order terms. We apply a Bouguer correction, with the ‘Bullard B’ curvature correction included. Finally, we apply terrain corrections in a three step process. Firstly, we estimate local terrain corrections out to Hammer Zone C (53 m) in the field and then calculate the effect using Hammer’s formula (Hammer, 1939). We then apply inner corrections using the ASTER GDEM2 1 arc sec (30 m) digital elevation model (DEM) out to 1000 m and outer corrections out to 167 km from a down sampled (3 arc sec, 90 m) version of the DEM using the Geosoft terrain correction software (Whitehead, 2006). We estimate the terrain corrections to be accurate to within
0.1 mGal, with most of the uncertainty coming from limitations of the 30 m DEM in the inner zone radius in areas of steep topography.

### A.1.1 Bulk density from gravity measurements and seismic velocity models

The Nettleton (1939) and Parasnis (1966) methods provide means of estimating bulk density of the terrain above the correction datum (ellipsoid) from analysis of Bouguer gravity correlation with topography (Figure A.1). For Nettleton’s method, we calculate the Pearson correlation coefficient between the Bouguer gravity and elevation for a range of Bouguer and terrain correction densities. The minimum correlation coefficient 0, corresponds to the bulk terrain density of 2375 kg/m$^3$. We used the same data set for Parasnis method and ran an ordinary least squares regression on the free air anomaly against the elevation data. The density is retrieved by the relation, density = regression slope / 2πG, where G is the gravity constant. The retrieved density is $2361 \pm 61$ kg/m$^3$, within error of the Nettleton result.

Additionally, we analyse an unpublished, commercially collected gravity dataset, from the Mariposa geothermal prospect ~ 10 km NW of the study area, using both Parasnis and Nettleton methods, and obtained densities of 2450 and $2452 \pm 24$ kg/m$^3$ respectively (Supplementary Figure A.1B).

Finally we compare our calculated densities to those derived from the conversion of the OVDAS 1-D Vp velocity model to density (Brocher, 2005). The top 3 km of this model gives densities of 2429 kg/m$^3$, increasing to 2448 kg/m$^3$ at 7 km depth. Averaging the five bulk density calculations gives a value of $2413 \pm 42$ kg/m$^3$ which we round to 2400 kg/m$^3$ for use in the gravity reduction.

### A.2 Inversion Method

The inversion script can be viewed on-line at [http://docs.simpeg.xyz/content/api_core/api_Examples.html](http://docs.simpeg.xyz/content/api_core/api_Examples.html). A jupyter python notebook is also included in the supplementary files.

The L$_2$ norm is defined as:

\[
\phi_m(m) = \frac{1}{2} \| W_s (m - m_{ref}) \|^2 + \sum_{i=x,y,z} \frac{1}{2} \| W_i G_i m \|^2
\]  

(A.1)

where $W_s$, $W_x$, $W_y$, $W_z$ are cell-based weights used to incorporate a depth weighting (Li and Oldenburg, 1998) to overcome the intrinsic loss of sensitivity with depth of potential field data, and $m_{ref}$ is a reference model which we set to zero. $G_x$, $G_y$, $G_z$ are forward difference operators measuring the spatial gradients of the model in Cartesian coordinates. In addition, we incorporate hard constraints on each cell in the mesh by imposing upper
Figure A.1: A) Parasnis (left) and Nettleton (right) plots for the LdMVF dataset. B) Parasnis (left) and Nettleton (right) plots for the Mariposa dataset.
and lower bounds on the allowable density. We also allow designated cells to be fixed, so that they contribute to the simulated gravity field, but are not adjusted in the inversion.

The discrete $L_p$ norm measure can be written as:

$$
\phi_m = \sum_{j=1}^{N} |m_i|^p + \sum_{i=x,y,z} \sum_{j=1}^{N} \left| (G_i m)_j \right|^q
$$

(A.2)

which we approximate with the function:

$$
\phi_m (m) = \sum_{j=1}^{N} \frac{m_j^2}{(m_j^2 + \epsilon_p)^{(1-\frac{p}{2})}} + \sum_{i=x,y,z} \sum_{j=1}^{N} \frac{(G_i m)_j^2}{(G_i m)_j^2 + \epsilon_q^{(1-\frac{q}{2})}}
$$

(A.3)

where $\epsilon$ is a small number to ensure zero division errors do not occur. For $p=2$, we recover the $L_2$ norm measure and $p \leq 1$ returns a compact model. The $L_p$ norm can be applied on both the model values ($m$) or the spatial gradients. We choose $p=0$ and $q_x=q_y=q_z=1$ as we want a model that retains a mix of compact and smooth features to reflect our conceptual notion of a magma chamber, that has a short but gradational density boundary from the reservoir to the host rock, representing an increasing degree of crystallisation from the center to the edge of the reservoir. Supplementary Figure A.2 shows the distribution of model values from the $L_2$ norm and $L_p$ norm after choosing the 95th percentile of the $L_2$ as a threshold.

The objective function for the $L_p$ norm is rewritten as:

$$
\phi (m) = \frac{1}{2} \left\| W_d (F[m] - d_{obs}) \right\|_2^2 + \frac{1}{2} \left\| W_s R_s (m - m_{ref}) \right\|_2^2 + \sum_{i=x,y,z} \left\| W_i R_i G_i m \right\|_2^2
$$

(A.4)

We solve the inverse problems described above using a scaled iterative re-weighted least squares (scaled-IRLS) optimisation approach (Fournier et al., 2016).

The scaled-IRLS weights $R_s$ and $R_i$ are defined as diagonal matrices:

$$
R_{sii} = \left[ \left( m_j^{(k-1)} \right)^2 + \epsilon_p^2 \right]^{(\frac{p}{2}-1)/2}
$$

(A.5)

$$
R_{iij} = \left[ \left( G_i m_j^{(k-1)} \right)^2 + \epsilon_q^2 \right]^{(\frac{q}{2}-1)/2}
$$

(A.6)

To run the inversion, we search for a perturbation of the model that reduces the objective function. The iterative optimisation process continues until the algorithm converges to a minimum and the misfit tolerance is achieved.
Figure A.2: Histograms showing the distribution of model density values from the $L_2$ norm (top) and $L_p$ norm (lower) after applying a compaction to values outside the 95th percentile indicated by red lines in the $L_2$ norm histogram.
Figure A.3: A) Rhyolite and rhyodacite densities at 90 MPa with 5 wt % H$_2$O. B) Andesite mcp and asd densities at 90 MPa and 3 wt% H$_2$O. The green shading shows the preferred temperature range from Fe-Ti geothermometer and the orange shading in the top plot indicates the preferred density model range.

A.3 Magma Densities

Supplementary Figure A.3 shows the density of (A) rhyolite and rhyodacite magmas at 90 MPa with 5 wt % H$_2$O and (B) two end member andesite magmas (asd and mcp) at 90 MPa and 3 wt % H$_2$O.

A.4 Role of Water and Brines in Magma Density

Supplementary Figure A.4 shows the variation in density of rhyolite magma with different water proportions. The shading indicates the preferred temperature range and density contrasts and shows that a minimum of 4 wt % H$_2$O is required. This is in agreement with the measured H$_2$O contents of >4.75 wt %.

The question arises to what influence magmatic brines have in the interpretation of the density model. It is difficult to judge the influence of brines for several reasons, firstly because we have hardly any evidence for their existence in detectable quantities. Firstly,
brines are good electrical conductors and should show up in the MT model. However the MT data show only a surface and shallow conductor interpreted as lake sediments and hydrothermal cap, and no large deeper conductor that could be a brine rich magma. Indeed the high melt content silicic magma system proposed would be not that conductive which could explain why MT does not image it, especially if it is concealed beneath two shallow conductors. Second, evidence from fluid inclusions. The crystal poor nature of the lavas at LdmVF results in very few fluid inclusions, making assessment of the composition of brines difficult.

The lack of surface geothermal expression means the top of the magma/hydrothermal system is probably closed. This might result in brines accumulating in significant quantity beneath the system cap. Hydrothermal brines outside of the magma system would only exist in filling existing pore space and would have little effect on the overall density model. For example, hydrothermal brines intruding into the Troncoso Fault as discussed in Chapter 3 are not detectable in the Bouguer gravity data so would not be detectable elsewhere in the model space. Magmatic brines existing in the melt/mush matrix are approximated to a first order by the MELTS modelling which includes H₂O. Replacing pure H₂O with brine would result in a lower density magmatic fluid and lowering the amount of melt required. However, the densities obtained from MELTS modelling with 5% H₂O provide good first order agreement with the preferred gravity model densities, so any brines present are probably only in small quantities and are not required to satisfy the gravity model.

In summary, although magmatic brines probably exist in the magma system there is very little geophysical or petrological evidence on which to assess their influence on the interpretation of the gravity model. It is however definitely worth keeping them in mind for other scenarios where more data exist.
Figure A.5: A) Histogram of observed - calculated values show a normal distribution. B) Map of residuals at each gravity station show no obvious geographic bias.

**A.5 Gravity Model Residual**

Supplementary Figure A.5 shows the distribution of residuals (observed - calculated values) from the final inversion. The residuals show a normal distribution (A) and no obvious geographic bias (B).

**A.6 End Member Gravity Models**

Supplementary Figure A.6 shows gravity model end members of -400 kg/m$^3$ (larger body) to -800 kg/m$^3$ (small body with mesh). The overall shape remains similar, but the volume of the body increases with lower density contrast.
Figure A.6: End member gravity models from -400 (larger body) to -800 kg/m$^3$ (smaller body). The sections are from the -600 to + 300 kg/m$^3$ model where the -200 kg/m$^3$ zone is interpreted as the crystallised rim of the active magma system.

A.7 Alternate Model

Supplementary Figure A.7 shows the results of a model where the top 500 m of the model are fixed at -300 kg/m$^3$ and the remainder of the cells are allowed to vary. A considerable amount of low density material is required beneath the surface layer, showing that a low density surface layer alone is unable to fit the data.

A.8 References


Figure A.7: Inversion slices and sections of alternate model with 500 m of -300 kg/m$^3$ material fixed at the surface. The remainder of the model is allowed to vary within the bounds -600 to 300 kg/m$^3$. A considerable volume of low density material beneath the surface layer is required to fit the data.


Appendix B

Supplementary Material for ‘Microgravity Changes at the Laguna del Maule Volcanic Field: Magma Induced Stress Changes Facilitate Mass Addition’

B.1 Network Design and Installation

In March and April 2013, we installed a 35 benchmark precisely-positioned microgravity network around LdM. The network benchmarks are stainless steel pins drilled and cemented into lava flows or large stable boulders, located around the southern edge of LdM and along Highway 115 to the north. Access to benchmarks on the southern lake shore is by boat and hiking, whilst those along Highway 115 are accessible by 2WD vehicle. The average distance between benchmarks is around 1km on the lake edges. We chose the locations of the benchmarks to cover the extent of the large deformation signal initially reported by Fournier et al. (2010). We installed additional benchmarks closer to the identified centre of uplift and also to the NW and NE away from the lake, to provide reference points away from the area of uplift. The reference benchmark, GLMREF2 was located in Argentina, outside the area of deformation. This benchmark was destroyed after the 2015 survey by roadworks, and GLM21 became the new reference benchmark.

B.2 Gravity Data Reduction

To produce precise residual gravity change ($\Delta g$) measurements, a number of corrections must be applied to the data to remove the effects of instrument drift, Earth tides and ocean loading, ground deformation and water table changes (Figure B.1), (e.g. Battaglia et al., 2008).
Figure B.1: Lake level variation between 2013 and 2016 as measured daily at the dam by the Ministry of Public Works in Chile. From May 2013 to January 2014 and March 2014 to March 2015, data are decimated to one point per month, on the first day of each month. The vertical bars show the observation time periods of each survey. The maximum lake level gravity correction in $\mu$Gal is labelled between each survey.
We controlled daily survey loops for drift and tares by repeat occupations of benchmark BASE, in the northwest of our study area. We occupied BASE at a minimum at the start and end of each day, and in the middle of each day if logistics permitted. For the final gravity value, we referenced our data to benchmark GLMREF2 which is outside the deformation area. We made several BASE-GLMREF2 ties with each gravity meter to ensure the reference benchmark is well constrained for each survey. After the 2015 survey, GLMREF2 was destroyed by road works, so GLM21 became the new reference for the 2016 survey. The 2015 data were recalculated relative to GLM21 to produce values for the 2016–2015 survey interval. It was not practical to recalculate the entire dataset to GLM21 as in earlier surveys GLM21 had relatively few observations (Figure B.2), making it less well constrained and unsuitable for a reference.

Raw gravity data were corrected for the effects of solid Earth tides and ocean loading using Gtools (Battaglia et al., 2012). Ocean loading was corrected using the FES2004 global model. Our inland location means that ocean loading effects are typically <3 µGal (1 µGal = 10^{-8} m/s^2). We applied linear drift and tare corrections after the tidal corrections. We repeated this process for each of the 3 gravity meters independently so that for each benchmark we calculate independent values for that benchmark, one from each gravity meter. For example

![Figure B.2: Number of occupations of each benchmark in the microgravity network per year. The top cell above GLM35 is GLMREF2. See Table B2 for details.](image-url)
Table B.1: Inter-meter calibration values for each survey year. Columns are Meter, referenced to D217, Correlation coefficient, R-squared fit, number of readings per calibration, 95% confidence intervals. ABS is the calibration of D217 to absolute benchmarks.

<table>
<thead>
<tr>
<th>Year</th>
<th>Meter (D217)</th>
<th>Coeff</th>
<th>R-Squared</th>
<th>n</th>
<th>95% Conf. Int.</th>
</tr>
</thead>
<tbody>
<tr>
<td>2013</td>
<td>G127</td>
<td>1.0046</td>
<td>1</td>
<td>6</td>
<td>1.004 - 1.005</td>
</tr>
<tr>
<td></td>
<td>G19</td>
<td>1.0030</td>
<td>1</td>
<td>12</td>
<td>1.002 - 1.004</td>
</tr>
<tr>
<td>2014</td>
<td>G127</td>
<td>1.0016</td>
<td>1</td>
<td>15</td>
<td>1.001 - 1.002</td>
</tr>
<tr>
<td></td>
<td>EG26</td>
<td>0.9954</td>
<td>1</td>
<td>24</td>
<td>0.995 - 0.996</td>
</tr>
<tr>
<td>2015</td>
<td>G127</td>
<td>0.9968</td>
<td>1</td>
<td>4</td>
<td>0.994 - 0.999</td>
</tr>
<tr>
<td></td>
<td>EG26</td>
<td>0.9935</td>
<td>1</td>
<td>45</td>
<td>0.993 - 0.994</td>
</tr>
<tr>
<td>2016</td>
<td>G127</td>
<td>1.0555</td>
<td>1</td>
<td>42</td>
<td>1.055 - 1.056</td>
</tr>
<tr>
<td></td>
<td>G943</td>
<td>1.0217</td>
<td>1</td>
<td>47</td>
<td>1.021 - 1.022</td>
</tr>
<tr>
<td></td>
<td>ABS</td>
<td>1.0005</td>
<td>1</td>
<td>5</td>
<td></td>
</tr>
</tbody>
</table>

A single repeat occupation of a benchmark results in 6 independent measurements from 3 gravity meters at that benchmark.

Each of the gravity meters employed in the survey has slight calibration differences that need to be considered for calculating precise gravity values. We performed a cross meter calibration using data from benchmarks common to all meters. To calculate the calibration factor, we performed an ordinary least square regression (using statsmodels python module) where meter D217 is kept as the dependent variable and the other meters are regressed against it. We chose D217 as the reference meter as it consistently had the lowest closure errors during all surveys (typically <15 µGal). We then applied the calibration factors to readings from each meter. Coefficients for cross meter calibration are shown in Table B.1. We found that the calibration coefficient between meters varied slightly from year to year, so we recalculate the calibration coefficient for each survey year. In 2016 we calibrated D217 to two absolute benchmarks, MAUL and LDMA (S. Bouvalot, pers. comm.) and found a calibration factor for D217 of 1.0005. As the calibration of D217 has likely varied over time we do not apply this absolute calibration to our dataset. We estimate an error of no more than 10 µGal between the absolute and relative gravity measurements, over our survey range of approximately 40 mGal.

Following calibration of each measurement from each gravity meter we undertook a second round of data quality control to eliminate those measurements that were obvious outliers from the other measurements at that benchmark. Typically we kept measurements that clustered within 20 µGal of each other, however we accepted a larger range if there were only a few readings for each benchmark. Poor clustering of individual readings may indicate tares in a measurement that are not removed in the linear drift correction model. We then recalculate the calibration factors with the outliers removed, and then calculate the gravity change map to check the overall smoothness of the gravity change field, aiming
to minimise ‘bullseyes’ at individual benchmarks, on the assumption that a gravity source
should produce a smoothly varying gravity field.

The rapid rate of uplift (∼0.001 m/day) occurring at some gravity benchmarks means
that measurable deformation occurs during our 2–3 week long survey campaigns. Hence
benchmarks that we occupied multiple times during a survey may have moved appreciably
between individual readings. In the worst case scenario, 0.021 m of uplift occurred between
the start and end of a campaign, which results in ∼7 µGal of gravity change if data are
combined from a benchmark measured on the first and last day of the survey. This gravity
change is incorporated in the error estimate of the affected benchmarks.

B.3 Survey Error Propagation

Errors are propagated as follows. Individual samples from each meter at each benchmark
are averaged to a mean value and standard deviation and standard error calculated to
provide one gravity value per benchmark per meter. Following between meter calibration
and quality control, an average value of the each benchmark is calculated and the individual
meter standard deviations are propagated using root mean square error propagation to give
a single gravity value and error value per benchmark. When the benchmark values are
converted to GLMREF2 as zero, the error from GLMREF2 is propagated to the other
benchmarks. We calculate a single gravity value per benchmark with standard deviation
and standard error. Errors in the height measurements are converted to gravity values via
the free air gradient. This height error is propagated via root mean square to the gravity
error estimate to give a final error value for the benchmark. Standard errors range from 12 to
47 µGal in 2014–2013 (average 19 µGal), 14 to 40 µGal in 2015–2014 (average 19 µGal) and
13 to 36 µGal (average 21 µGal) in 2016–2015. For benchmarks with only 1 meter occupation
(see Figure B.2) the standard error of the gravity change is made up of the standard error
of the individual samples comprising the reading (typically <5 µGal) plus the height change
error. These standard errors should be treated as minimum values. Gravity change data are
presented in Table B.2.

B.4 Deformation Models of the Gravity Change Source

We model the deformation produced by gravity change source prisms if they are treated as
intruding dykes, rather than the passive pore filling mechanism proposed. For each gravity
change interval we model the mode gravity source prism using the Okada (1985) solution
for a tensile crack oriented vertically and compare it to a horizontally oriented tensile crack
representing the sill that best fits the observed deformation (Feigl et al., 2014). See Figures
B.3, B.4, B.5.

These models show that in 2014–2013 the dyke representation of the best fit gravity model
(Model I) would produce a maximum displacement of ∼1.6 m (Figure B.3B). The subsidence
lobe is centered beneath the lake and may not be detected however the western inflation
lobe would be clearly detectable at CGPS station MAU2 (gravity benchmark GLM09) on
the peninsula. When combined with the sill model a total vertical displacement of ∼1.8 m is
Table B.2: Gravity data table. Gravity values ($\Delta g$) and standard error (STDER) in $\mu$Gal. Height change values ($\Delta H$) in m.

<table>
<thead>
<tr>
<th></th>
<th></th>
<th></th>
<th></th>
<th></th>
<th></th>
<th></th>
<th></th>
<th></th>
<th></th>
<th></th>
<th></th>
<th></th>
</tr>
</thead>
<tbody>
<tr>
<td>BASE</td>
<td>-36.01716</td>
<td>-70.5595</td>
<td>2215.1</td>
<td>0.064</td>
<td>10.6</td>
<td>12.4</td>
<td>0.065</td>
<td>4.4</td>
<td>11.2</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>GLM03</td>
<td>-36.0176</td>
<td>-70.5055</td>
<td>2215.1</td>
<td>0.084</td>
<td>10.6</td>
<td>12.4</td>
<td>0.065</td>
<td>4.4</td>
<td>11.2</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>GLM04</td>
<td>-36.0176</td>
<td>-70.5055</td>
<td>2215.1</td>
<td>0.084</td>
<td>10.6</td>
<td>12.4</td>
<td>0.065</td>
<td>4.4</td>
<td>11.2</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>GLM05</td>
<td>-36.0176</td>
<td>-70.5055</td>
<td>2215.1</td>
<td>0.084</td>
<td>10.6</td>
<td>12.4</td>
<td>0.065</td>
<td>4.4</td>
<td>11.2</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>GLM06</td>
<td>-36.0176</td>
<td>-70.5055</td>
<td>2215.1</td>
<td>0.084</td>
<td>10.6</td>
<td>12.4</td>
<td>0.065</td>
<td>4.4</td>
<td>11.2</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>GLM07</td>
<td>-36.0176</td>
<td>-70.5055</td>
<td>2215.1</td>
<td>0.084</td>
<td>10.6</td>
<td>12.4</td>
<td>0.065</td>
<td>4.4</td>
<td>11.2</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>GLM08</td>
<td>-36.0176</td>
<td>-70.5055</td>
<td>2215.1</td>
<td>0.084</td>
<td>10.6</td>
<td>12.4</td>
<td>0.065</td>
<td>4.4</td>
<td>11.2</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>GLM09</td>
<td>-36.0176</td>
<td>-70.5055</td>
<td>2215.1</td>
<td>0.084</td>
<td>10.6</td>
<td>12.4</td>
<td>0.065</td>
<td>4.4</td>
<td>11.2</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>GLM10</td>
<td>-36.0176</td>
<td>-70.5055</td>
<td>2215.1</td>
<td>0.084</td>
<td>10.6</td>
<td>12.4</td>
<td>0.065</td>
<td>4.4</td>
<td>11.2</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>GLM11</td>
<td>-36.0176</td>
<td>-70.5055</td>
<td>2215.1</td>
<td>0.084</td>
<td>10.6</td>
<td>12.4</td>
<td>0.065</td>
<td>4.4</td>
<td>11.2</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>GLM12</td>
<td>-36.0176</td>
<td>-70.5055</td>
<td>2215.1</td>
<td>0.084</td>
<td>10.6</td>
<td>12.4</td>
<td>0.065</td>
<td>4.4</td>
<td>11.2</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>GLM13</td>
<td>-36.0176</td>
<td>-70.5055</td>
<td>2215.1</td>
<td>0.084</td>
<td>10.6</td>
<td>12.4</td>
<td>0.065</td>
<td>4.4</td>
<td>11.2</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>GLM14</td>
<td>-36.0176</td>
<td>-70.5055</td>
<td>2215.1</td>
<td>0.084</td>
<td>10.6</td>
<td>12.4</td>
<td>0.065</td>
<td>4.4</td>
<td>11.2</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>GLM15</td>
<td>-36.0176</td>
<td>-70.5055</td>
<td>2215.1</td>
<td>0.084</td>
<td>10.6</td>
<td>12.4</td>
<td>0.065</td>
<td>4.4</td>
<td>11.2</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>GLM16</td>
<td>-36.0176</td>
<td>-70.5055</td>
<td>2215.1</td>
<td>0.084</td>
<td>10.6</td>
<td>12.4</td>
<td>0.065</td>
<td>4.4</td>
<td>11.2</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>GLM17</td>
<td>-36.0176</td>
<td>-70.5055</td>
<td>2215.1</td>
<td>0.084</td>
<td>10.6</td>
<td>12.4</td>
<td>0.065</td>
<td>4.4</td>
<td>11.2</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>GLM18</td>
<td>-36.0176</td>
<td>-70.5055</td>
<td>2215.1</td>
<td>0.084</td>
<td>10.6</td>
<td>12.4</td>
<td>0.065</td>
<td>4.4</td>
<td>11.2</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>GLM19</td>
<td>-36.0176</td>
<td>-70.5055</td>
<td>2215.1</td>
<td>0.084</td>
<td>10.6</td>
<td>12.4</td>
<td>0.065</td>
<td>4.4</td>
<td>11.2</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>GLM20</td>
<td>-36.0176</td>
<td>-70.5055</td>
<td>2215.1</td>
<td>0.084</td>
<td>10.6</td>
<td>12.4</td>
<td>0.065</td>
<td>4.4</td>
<td>11.2</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>GLM21</td>
<td>-36.0176</td>
<td>-70.5055</td>
<td>2215.1</td>
<td>0.084</td>
<td>10.6</td>
<td>12.4</td>
<td>0.065</td>
<td>4.4</td>
<td>11.2</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>GLM22</td>
<td>-36.0176</td>
<td>-70.5055</td>
<td>2215.1</td>
<td>0.084</td>
<td>10.6</td>
<td>12.4</td>
<td>0.065</td>
<td>4.4</td>
<td>11.2</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>GLM23</td>
<td>-36.0176</td>
<td>-70.5055</td>
<td>2215.1</td>
<td>0.084</td>
<td>10.6</td>
<td>12.4</td>
<td>0.065</td>
<td>4.4</td>
<td>11.2</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>GLM24</td>
<td>-36.0176</td>
<td>-70.5055</td>
<td>2215.1</td>
<td>0.084</td>
<td>10.6</td>
<td>12.4</td>
<td>0.065</td>
<td>4.4</td>
<td>11.2</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>GLM25</td>
<td>-36.0176</td>
<td>-70.5055</td>
<td>2215.1</td>
<td>0.084</td>
<td>10.6</td>
<td>12.4</td>
<td>0.065</td>
<td>4.4</td>
<td>11.2</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>GLM26</td>
<td>-36.0176</td>
<td>-70.5055</td>
<td>2215.1</td>
<td>0.084</td>
<td>10.6</td>
<td>12.4</td>
<td>0.065</td>
<td>4.4</td>
<td>11.2</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>GLM27</td>
<td>-36.0176</td>
<td>-70.5055</td>
<td>2215.1</td>
<td>0.084</td>
<td>10.6</td>
<td>12.4</td>
<td>0.065</td>
<td>4.4</td>
<td>11.2</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>GLM28</td>
<td>-36.0176</td>
<td>-70.5055</td>
<td>2215.1</td>
<td>0.084</td>
<td>10.6</td>
<td>12.4</td>
<td>0.065</td>
<td>4.4</td>
<td>11.2</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>GLM29</td>
<td>-36.0176</td>
<td>-70.5055</td>
<td>2215.1</td>
<td>0.084</td>
<td>10.6</td>
<td>12.4</td>
<td>0.065</td>
<td>4.4</td>
<td>11.2</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>GLM30</td>
<td>-36.0176</td>
<td>-70.5055</td>
<td>2215.1</td>
<td>0.084</td>
<td>10.6</td>
<td>12.4</td>
<td>0.065</td>
<td>4.4</td>
<td>11.2</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>GLM31</td>
<td>-36.0176</td>
<td>-70.5055</td>
<td>2215.1</td>
<td>0.084</td>
<td>10.6</td>
<td>12.4</td>
<td>0.065</td>
<td>4.4</td>
<td>11.2</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>GLM32</td>
<td>-36.0176</td>
<td>-70.5055</td>
<td>2215.1</td>
<td>0.084</td>
<td>10.6</td>
<td>12.4</td>
<td>0.065</td>
<td>4.4</td>
<td>11.2</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>GLM33</td>
<td>-36.0176</td>
<td>-70.5055</td>
<td>2215.1</td>
<td>0.084</td>
<td>10.6</td>
<td>12.4</td>
<td>0.065</td>
<td>4.4</td>
<td>11.2</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>GLM34</td>
<td>-36.0176</td>
<td>-70.5055</td>
<td>2215.1</td>
<td>0.084</td>
<td>10.6</td>
<td>12.4</td>
<td>0.065</td>
<td>4.4</td>
<td>11.2</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>GLM35</td>
<td>-36.0176</td>
<td>-70.5055</td>
<td>2215.1</td>
<td>0.084</td>
<td>10.6</td>
<td>12.4</td>
<td>0.065</td>
<td>4.4</td>
<td>11.2</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>GLMREF2</td>
<td>-36.0176</td>
<td>-70.5055</td>
<td>2215.1</td>
<td>0.084</td>
<td>10.6</td>
<td>12.4</td>
<td>0.065</td>
<td>4.4</td>
<td>11.2</td>
<td></td>
<td></td>
<td></td>
</tr>
</tbody>
</table>
Figure B.3: Deformation models of the 2014–2013 gravity change source. Note color scale changes between images. A) InSAR sill solution (red rectangle) from Feigl et al. (2014) that best fits the observed vertical displacement between 2013 and 2014. B) Vertical displacement for the Model I mode gravity solution modelled as a dyke (red dashed line) using Okada (1985) for a tensile crack. C) Combined vertical displacement of the sill and dyke for gravity solution Model I. D) Vertical displacement for Model II mode gravity solution modelled as a dyke using Okada (1985) for a tensile crack. E) Combined vertical displacement of the sill and dyke for gravity solution Model II.
Figure B.4: Deformation models of the 2015–2014 gravity change source. Note color scale changes between images. A) InSAR sill solution (red rectangle) from Feigl et al. (2014) that best fits the observed vertical displacement between 2014 and 2015. B) Vertical displacement for the mode gravity solution modelled as a dyke (red dashed line) using Okada (1985) for a tensile crack. C) Combined vertical displacement of the sill and dyke.
Figure B.5: Deformation models of the 2016–2015 gravity change source. Note color scale changes between images. A) InSAR sill solution (red rectangle) from Feigl et al. (2014) that best fits the observed vertical displacement between 2015 and 2016. B) Vertical displacement for the mode gravity solution modelled as a dyke (red dashed line) using Okada (1985) for a tensile crack. C) Combined vertical displacement of the sill and dyke.
calculated (Figure B.3C), compared to the observed displacement of ~0.25 m Figure (B.3A). Model II of the gravity solution produces a smaller vertical displacement (Figures B.3D and E), but still tens of cm greater than the observed vertical displacement.

In 2015–2014 the gravity solution is centered mostly on land and the vertical displacement of the equivalent dyke is 0.8 m when combined with the observed sill inflation model (Figure B.4C). The characteristic subsidence lobe is not present in the observed deformation field (A).

In 2016–2015 the gravity source is centered beneath the lake, and it is possible that the equivalent dyke displacement solution of the gravity source could go undetected, although the subsidence lobe in above the dyke should be detected by MAU2 CGPS station.

These displacement models show that a dyking mechanism for generating gravity changes is unlikely as it produces deformation that should be clearly detectable in all time periods. In addition, studies by Le Mével et al. (2016), and Miller et al. (2017) both conclude that the magma system pressure is not sufficient to generate dyking in the roof of the chamber. For these and the other reasons presented we consider dyke injection an unlikely mechanism and prefer a method of passive fluid infiltration.

### B.5 References


Appendix C

Magnetic Survey on Laguna del Maule

C.1 Introduction

I collected a magnetic dataset as a compliment to the Bouguer gravity, and to provide some geophysical data coverage over the lake. A lake borne seismic reflection dataset was collected independently by Katie Keranen and Dana Peterson from Cornell University. The models and interpretation of the combined seismic and magnetic data are in preparation for publication in a separate journal article. This appendix provides extra details on the data acquisition, processing, modelling and interpretation.

C.2 Data Acquisition and Processing

Over 2nd and 3rd March 2016, I collected 200 line km of magnetic data over Lake Maule. I collected data along 20 east-west lines spaced at 400 m with four, 2 km spaced N-S tie lines. Navigation was by preset GPS waypoints, allowing straight lines to be maintained.

I mounted an overhauser magnetometer (GEM systems GSM-19) in a non magnetic inflatable zodiac, towed 30 m behind a larger fibreglass hulled inflatable boat (Figure C.1). Magnetic data sampling was at 1 Hz and acquisition speeds varied from 10 to 30 km/h resulting in an average sample spacing of ~5 m. Data are positioned using a Garmin GPS 60 sampling at 2 seconds. Horizontal positioning accuracy is estimated at ± 10 m (2 samples), whilst vertical positioning is taken from the precisely measured (using Differential GPS) lake level (+ 0.1 m) plus a fixed offset of the sensor mounted on a pole (approximately 0.75 m) (Figure C.2). GPS data were interpolated to a 1 second data set and merged with the magnetic data, to provide a positioned magnetic value every one second.

To control diurnal variations, a 3 component flux gate magnetometer (Sensys FGM3D) was installed at our accommodation on the lake shore, sampling at 1 Hz. The 3 component data were summed (root mean square) to a single value, and a 151 point (2.5 minute)
Figure C.1: Red inflatable boat towed 30 m behind fibreglass hulled tow boat. Magnetometer and GPS mounted in the inflatable, as in Figure C.2

Figure C.2: Magnetometer mounted in the inflatable boat. Floor is made of non magnetic aluminium. The magnetometer sensor is mounted on pole inside black plastic bag. Blue dry bag contains GPS and magnetometer recording unit.
hanning smoothing filter (Figure C.3) applied to remove spikes. The base station data were subtracted from the rover data to correct for diurnal variations (max 25 nT) (Figure C.4). Analysis of cross over points of lines and tie lines shows repeatability to $+\sim 60$ nT with 66% (1 standard deviation) of points being within 14 nT (Figure C.5).

We subtracted the IGRF model for the acquisition date, using a single value of 24340 nT (for the latitude, longitude of the lake), as the IGRF varies less than 4 nT over the study area. Inducing field inclination is $-36.8^\circ$, and declination $3.3^\circ$. This inducing field should result in anomalies with the positive pole oriented directly north of the negative pole.

### C.3 Image Analysis

In the Total Magnetic Intensity (TMI) image (Figure C.6A), shallow, magnetic lava flows dominate the response. On the east and north shores of the lake it appears some anomalies are remanently magnetised, with strong negative anomalies north of positive anomalies. In the south center of the lake low frequency, low to moderate amplitude anomalies exist. The first vertical derivative image (Figure C.6B) highlights NE-SW trending features, particularly north of the peninsula.

### C.4 Inversion Method

I use the open source SimPEG (http://www.simpeg.xyz) python code (Cockett et al., 2015) to invert the data. I construct a 50 x 50 m spaced horizontal mesh with the vertical dimen-
Figure C.4: Rover raw and diurnal corrected (red and black, plotted on top of each other), and base station (green) timeseries.

Figure C.5: Histogram of cross-over point variation. 66\% of cross over points are within 14 nT.
Figure C.6: Boat magnetic data acquired within Laguna del Maule, and depth models. A) Total magnetic anomaly corrected for diurnal variations and removal of the IGRF field; B) First vertical derivative of the magnetic anomaly; C) Depth slice from the modelled data at 1475 m a.s.l. (∼700 m depth); D) Depth slice from modelled data at 1875 m a.s.l (∼300 m depth). Lava flows on the lake floor dominate the total magnetic anomaly. The first vertical derivative and in depth slices highlight northeast-southwest-trending faults in the northern portion of the lake.

The inversion increasing from 50 m to 350 m at the base of the model. The model contains ∼1.9 million cells and the inversion was run on a cluster with 128 Gb ram, so that the entire model can be inverted without tiling. The possibility of remanent magnetisation requires a three step inversion process (Fournier et al, 2016). First, magnetic data are inverted for an equivalent-source layer, which is used to derive magnetic amplitude data. Next, amplitude data are inverted for an effective susceptibility model, providing information about the geometry and distribution of magnetized objects. Finally, the effective susceptibility model is used to constrain the Magnetic Vector Inversion (MVI), recovering the orientation and magnitude of magnetization. All three algorithms are formulated as regularized least-squares problems solved by the Gauss-Newton method. I further constrain the solution by imposing sparsity...
constraints on the model and model gradients via an approximated \( l_p \)-norm penalty function. I elaborate a Scaled Iterative Re-weighted Least-Squares (S-IRLS) method, allowing for a stable and robust convergence of the algorithm while combining different \( l_p \)-norms on the range \( 0 \leq p \leq 2 \). The goal is to reduce the complexity of magnetization models while also imposing geometrical constraints on the solution. I test models where the objective function is rotated parallel to the dominant NE trend of the TMI map, in order to provide better continuity of anomalies oriented in this direction. Additionally I apply a compact norm to make the boundaries of magnetic features sharper. I apply a \( l_p = 0 \) ‘compact’ norm to the amplitude and \( y \) and \( z \) gradients of the amplitude, while the \( x \) gradient has a ‘smooth’ \( l_2 \) norm. Combined with the rotated objective function this works to sharpen the east-west boundaries of the NE/SW trending features.

C.5 Interpretation

The reversely magnetised linear features are interpreted as basaltic dykes intruding beneath the lake. Below around 1000 m depth the model loses resolution. Comparison of the dykes with seismic reflection data shows that the dykes are probably intruded along faults. The remanent magnetisation suggests that the dykes are older than 780 Ka, and were intruded in the Matuyama magnetic epoch, and possibly relate to activity after the formation of the Bobadilla Caldera at 950 Ka. Further paleomagnetic and dating studies would be useful to better constrain the age of the dykes. More details of interpretation will be provided in the paper currently in preparation.

C.6 References


Appendix D

Supplementary Material for ‘Internal Structure and Volcanic Hazard Potential of Mt Tongariro, New Zealand, from 3D Geologically Constrained Gravity and Magnetic Models’

D.1 Geophysical Model

The geophysical model is supplied as a Geoscience Analyst project file for visualisation. The model file can be downloaded from http://tinyurl.com/ntw5tel.


D.2 Gravity Data Correction Scheme

We corrected data to the ellipsoid height datum as our heights are largely derived from GNSS (WGS84 datum) measurements. Orthometric heights from the older surveys were converted to ellipsoid heights using a geoid model (NZVGD09) from Land Information New Zealand, so that they could be processed in a uniform manner with the new stations. We corrected all data for latitude variations using the theoretical gravity value from 1980 GRS Geodetic Reference System (GRS80) formula. As our survey covers a large range of elevations (600 to 2300 m.a.s.l.) we applied an atmospheric correction to account for the changing density of atmosphere. We applied the free air correction using the full equation with second order terms and applied a Bouguer correction, with the curvature correction included. Finally
we applied terrain corrections in a 3 step process. Firstly, inner terrain corrections out to Hammer Zone D were estimated in the field and then calculated using Hammer’s formula (Hammer, 1939). We then applied outer corrections using an 8 m DEM for Zones E and F and used a 15 m DEM for corrections in zones G to M. The terrain corrections represent one of the larger sources of error. Comparison of the manual inner (B–D) corrections with those from the 8 m DEM show a mean variation of 0.04 mGal. To estimate errors in the outer zones (E–M) we computed terrain corrections with an 8 m DEM and a 15 m DEM and calculate an RMS error in the differences between the two of 0.045 mGal. The DEMs used for terrain corrections do not take into account the bathymetry of lakes and as such the terrain corrections may be over-estimated due to low density water being corrected as denser rock (Hasegawa et al., 2009). Lake Taupo is New Zealand’s largest lake (616 km²) and is located 13 km north of the model area, but is within the broader area used to determine the regional field. To determine the gravity effect of the lake, we constructed a forward model using a slab with the approximate outline of the lake and a thickness of 100 m (the average lake depth). The maximum gravity effect at a station directly on the lake shore is 6 mGal, reducing to 0.02 mGal for the nearest station within our modelling area. We subtracted the gravity signal of Lake Taupo from our terrain correction value to ensure our stations used for regional field determination are properly corrected. The smaller Lake Rotoaira (13 km²), directly on the north edge of the model area, is on average 10 m deep and has a maximum gravity effect at the closest station of 0.023 mGal. This is within our noise envelope so we did not apply a further correction. No corrections were applied for the many small lakes on Mt Tongariro as they are too small to produce a measurable gravity effect.

D.3 Bulk Density from Gravity Measurements

The Nettleton (Nettleton, 1939) and Parasnis (Parasnis, 1966) methods provide means of estimating bulk density from minimisation of the correlation of Bouguer and terrain corrected gravity with topography and variants of these methods have been applied in several volcano studies (Gottsmann et al., 2008; Hautmann et al., 2013). We compare calculated bulk density estimations to rock physical property measurements to determine the optimal gravity correction density. An important and often overlooked caveat for both methods is that there is no direct causation of the topography by the geology, i.e., that dense rocks do not form high peaks or vice versa. This assumption can be difficult to verify in volcanic terranes where high standing ridges are often composed of denser, more erosion resistant material such as lava flows, while lower density pyroclastics fill valley floors. Both methods also rely on there being no regional affect in the gravity data which could bias the analysis. To minimise these restrictions, we selected data from 156 stations located between the Waihi and Poutu fault zones where Cassidy et al. (2009) showed that the regional field is relatively flat. These data should be mostly free of long wavelength variations caused by large scale faulting and regional gravity changes, and best represent the bulk density of the volcano superstructure from 1300 to 2300 m elevation.

For Nettleton’s method we calculated the Pearson correlation coefficient between the Bouguer gravity and elevation for a range of Bouguer and terrain correction densities. The correlation coefficient closest to zero, corresponds to the bulk terrain density of 1900 kg/m³ (Supplementary figure D.1). We used the same data set for Parasnis method and ran an ordinary
Figure D.1: Plot of result of Nettleton’s method, showing density at the minimum Pearson correlation coefficient between elevation and Bouguer corrected gravity for each density.

Figure D.2: Plot of result of Parasis method, showing density derived from the slope of the free air anomaly vs elevation.
least squares regression on the free air anomaly against the elevation data. The density is retrieved by the relation: density = regression slope / \(2\pi G\), where \(G\) is the universal gravity constant (Supplementary figure D.2).

All except two points plot within the 95% confidence interval of the best fit regression line. These points are located on the summit of Ngauruhoe and hand specimen densities from the 1954 summit cone are the lowest in the physical property dataset, between 498–731 kg/m\(^3\). As these points are outside the 95% confidence bounds, they are excluded from the regression analysis. The retrieved density is \(1904 \pm 40\) kg/m\(^3\) and agrees very well with the density retrieved from Nettleton’s method. The bulk density values retrieved by Nettleton’s and Parasnis’s methods are close to the mean wet density of pyroclastic materials sampled \((1931 \pm 267\) kg/m\(^3\)). However, this density appears to be too low, as andesite lava (wet density 2535 \(\pm 205\) kg/m\(^3\)), makes up a sizeable proportion of the massif (Supplementary Figure D.3). We interpret this discrepancy as due to a violation of the assumptions of Nettleton’s and Paranis’s method, implying that even with careful data selection there is still an implicit correlation between the topography and density. As such we use our physical property measurements for the correction density (Supplementary figure D.3).

### D.4 Inversion Method

VPmg solves the inversion using the steepest descent method, where no matrix inversion is required (Fullagar and Pears, 2007). For the conventional least squares inversion, the objective at each iteration is the smallest parameter perturbation required to halve the \(L2\)-norm data misfit. Stochastic inversion is an option for heterogeneous property inversion of either basement or other units. In this case, individual model cells are subjected to random property perturbations; the perturbation is accepted if it produces a reduction in misfit and if it is compatible with the expected property distribution within the geological unit. Maximum perturbation size is defined in terms of absolute property change, for property inversion, or in terms of fractional change in depth, for geometry inversion. Degree of fit is determined by the magnitude of the chi-squared data norm, \(L2\), and the \(L1\) data norm, defined by

\[
L2 = \frac{1}{N} \sum_{n=1}^{N} \left( \frac{o_n - c_n}{\varepsilon_n} \right)^2 \tag{D.1}
\]

\[
L1 = \frac{1}{N} \sqrt{\pi/2} \sum_{n=1}^{N} \left| \frac{o_n - c_n}{\varepsilon_n} \right| \tag{D.2}
\]

where \(N\) is the number of data, \(O_n\) is the measured data, \(C_n\) the calculated model response and \(\varepsilon_n\) the uncertainty (standard deviation) applied to the nth data point. If the data uncertainties are controlled by normal random variables with zero mean, then both \(L2\) and \(L1\) have expected values of unity. Therefore, the model is deemed acceptable if \(L2 < 1\) and/or if \(L1 < 1\). For a starting \(\varepsilon_n\), we used a value of 10% of the standard deviation of the range of observed data. If a model converges with this \(\varepsilon_n\) we successively lower the \(\varepsilon_n\) until the inversion stalls, at which point we have achieved the best fit possible.

The \(L1\) and \(L2\) misfits are dimensionless, as they are normalised by the uncertainties, \(\varepsilon_n\).
Figure D.3: Histograms of physical property measurements for a range of rock types.
A depth weighting is applied to lower the sensitivity of the inversion to deeper cells, with the aim of creating 'smaller' bodies and limiting the smearing of bodies with depth.

D.5 Gravity Results

D.5.1 Apparent density inversion

We begin our exploration of the gravity data by performing an ‘apparent density’ (AD) inversion. An AD model is a voxet with only 1 cell per vertical prism (i.e. the cell vertical extent is the same as the whole voxet) and the inversion adjusts the density in each prism until the calculated response of the model matches the input data. Apparent density models are useful for looking at lateral density variations in the dataset. The result of the apparent density inversion is seen in Supplementary Figure D.4A. We use a diverging colour scheme to highlight areas of higher and lower density, relative to the Bouguer correction density of 2300 kg/m$^3$. The model has an RMS misfit of 1.0 mGal which is relatively high and indicates that a more detailed model which incorporates vertical density variations is required.

The residual gravity (observed - calculated) map (Supplementary Figure D.4B) shows the highest areas of misfit are under the main part of the volcanic edifice, where the geology is likely to be highly three dimensional. Areas of high positive misfit are also seen in this area, and also in some areas on basement rock. The basement areas contain data from stations which may have errors up to 1 mGal associated with them, and this could be reflected in the misfit in these areas.

The AD model consists of a central low density zone flanked by high densities on either side. The low density area is mostly confined between the Waihi and Poutu faults and the high density region correlates with outcropping or shallow basement. Within the central low density zone, several areas of higher or lower density exist. Two areas of high density occur, one to the south of Pukekaikiore and the other to the east of Blue Lake. An area of low density is associated with Mt Ngauruhoe and another low density area occurs around Red Crater.

D.5.2 3D unconstrained density inversion

To explore the depth extent of lateral density variations identified in the apparent density inversion we construct a model using a 3D unconstrained inversion. In this model we use the same voxet as the apparent density inversion, however we sub divide the vertical dimension into smaller cells. The top 2000 m of cells are constant thickness of 100 m and those deeper increase in thickness via an expansion factor of 1.1 times the previous cell thickness. The expanding cell size results in cells at -5000 m.b.s.l. that are ~400 m thick reflecting the lower resolution of the data at depth. The voxet is filled initially with cells of 0 kg/m$^3$ density contrast, equivalent to the bulk density of 2300 kg/m$^3$. We set the target RMS misfit to be 0.1 mGal (similar to our observation errors), to ensure we extract maximum information from our high quality data. The final model successfully converged with an RMS of 0.15 mGal.
Figure D.4: A) Apparent density model. Location of vents in white triangles and active faults in black lines. B) Residual gravity distribution and histogram.
Figure D.5: A) Slice of unconstrained density inversion model at a 350 m.a.s.l. Location of vents in white triangles and active faults in black lines. B) Residual anomaly map and histogram of residuals, RMS = 0.15 mGal, from the unconstrained density inversion.
A representative depth slice of the resulting model voxet at 350 m.a.s.l. is shown in Supplementary Figure D.5A, along with the residual misfit anomaly map and histogram in Supplementary Figure D.5B. In this depth slice we see significantly more detail than from the AD inversion. High densities in the northwest and east correspond to outcropping basement. The broad low density zone between the Waihi and Poutu faults persists, as do the smaller high density zones to the east of Blue Lake and along the Waihi fault south of Puakeaikire. Outside the central fault zone a more complex pattern of high and low densities occur. High density west of the Puakeonake vents correspond to the lava field from these cones and high density along the north part of the Waihi fault corresponds to lava flows from North Crater. High density east of the Poutu fault may relate to shallower basement east of this fault, however some small high and low density anomalies east of the Poutu fault are also associated with high residuals, so should be treated with caution. These areas are also associated with gravity stations that have higher errors (up to 1 mGal) than the rest of the dataset.

The 3D unconstrained inversion model may have overestimated the thickness of many of the features in the model. For instance it is unlikely that the lavas from Puakeonake cones extend from $\sim 1100$ m.a.s.l. at surface to 350 m.a.s.l.

### D.6 Magnetic Results

#### D.6.1 Apparent susceptibility inversion

We initially invert the magnetic data for an apparent susceptibility model to show the broad lateral distribution of magnetic rocks. The result of this inversion is shown in Supplementary Figure D.6A, along with a map and histogram showing the distribution of residuals from the inversion, (Supplementary Figure D.6B). The RMS of the apparent susceptibility inversion is 8 nT, <1 % of the maximum anomaly amplitude. The main features are a pronounced area of very low to no susceptibility to the north of Mt Ngauruhoe, extending to Upper Te Maari Crater. This low is ringed by a series of magnetic highs with a maximum susceptibility on the north flank of Mt Ngauruhoe. A small area of low susceptibility is coincident with the Ketetahi hot springs and a broad area of moderate magnetic susceptibility is located around the Tama Lakes.

#### D.6.2 3D unconstrained susceptibility inversion

We next performed a 3D unconstrained susceptibility inversion to define the lateral and vertical extent of both magnetic and non magnetic bodies within the volcano. From this model we extract a representative depth slice at 750 m.a.s.l. to show the main magnetic and non-magnetic features (Supplementary Figure D.7A). The RMS of this inversion is 4 nT and shows a normal distribution of residuals indicating no bias in the model (Supplementary Figure D.7B).

To constrain the model to geologically realistic values the range of susceptibilities was allowed to vary between 0 and 0.2 SI. This is greater than our limited range of physical
Figure D.6: A) Result of apparent susceptibility inversion. Location of vents in white triangles and active faults in black lines. B) Residual anomaly map and histogram of residuals, RMS = 8 nT, from the apparent susceptibility inversion.
Figure D.7: A) Result of unconstrained susceptibility inversion at a 750 m.a.s.l. Location of vents in white triangles and active faults in black lines. B) Residual anomaly map and histogram of residuals, RMS = 4 nT, from the unconstrained susceptibility inversion.
property measurements however magnetic susceptibility measurements on hand specimens are commonly underestimated compared to bulk rock composition, and are within ranges reported for similar sized andesite stratovolcanoes (e.g. Finn et al., 2007).

The dominant feature of this model is a very low to no magnetisation zone under the central part of the massif, extending to the north side of Ngauruhoe and to Upper Te Maari Crater. This zone is broadly coincident with the low density anomaly resolved from the gravity data. Surrounding this zone is a narrow irregular ring of higher magnetised rock, coincident with flank lava flows, extending out to the Waihi and Poutu faults. Within this ring an area of low magnetisation is coincident with the Ketetahi hot springs. The area of highest magnetisation occurs at Mt Ngauruhoe and around the Upper Tama Lakes. An area of high magnetisation on the extreme south of the model relates to lava flows from Mt Ruapehu. West of the Waihi fault an area of low magnetisation occurs to the north of the Pukeonake cones, coincident with a gravity high, and may be related to lava flows from theses vents.

D.7 Limitations and Sensitivity of Geophysical Models

A way of testing the sensitivity of our geologically constrained models is to compare them to the unconstrained 3D models. If features introduced in our geologically constrained models are discernible in the unconstrained models then we can have a higher degree of confidence in their validity. Likewise within the geologically constrained models we can test a range of starting models and see how closely they converge towards a single model through the inversion process. In each case, the RMS misfit of the model indicates all models are mathematically acceptable representations of the observed data.

The sensitivity of features to a constrained basement depth is judged by if they extend below basement depths in the unconstrained model, or terminate at or above the basement in the geologically constrained model. As an example we compare the iso-surfaces of the low density root imaged beneath Ngauruhoe from the 3D unconstrained inversion and the geologically constrained model (Supplementary Figure D.8A). In the unconstrained inversion the low density root from Mt Ngauruhoe appears to extend into the basement to a depth of -300 m. In the geologically constrained model this body is terminated at the basement, however the low pass filtered data used to construct the basement surface may have removed the signal of a small low density body close under the basement interface so it is not unreasonable to imagine the extension of the low density body into the basement in the unconstrained 3D model using the full wavelength dataset. This is also feasible when we consider that repeat magmatic intrusion is likely to leave the basement highly fractured, effectively lowering its density. Similarly we compare the depth extent of the low magnetic susceptibility body, associated with the hydrothermal system from unconstrained and geologically constrained models in Supplementary Figure D.8B. In the geologically constrained model the low susceptibility zone terminates a few hundred metres above the basement, while in the unconstrained model a larger volume of very low susceptibility material extends into the basement. This example also illustrates the geophysical equivalence of a smaller higher susceptibility body and a larger, lower susceptibility body. The two examples may be taken as end members of the set of likely models.
Figure D.8: Perspective views from south-east showing: A) Density iso-surface \((2250 \text{ kg/m}^3)\) of low density root under Ngauruhoe from unconstrained inversion (red) and geologically constrained inversion (blue). B) magnetic susceptibility iso-surfaces of the hydrothermal system extent from unconstrained inversion (yellow - 0.001 SI) and geologically constrained inversion (red - 0.025 SI). In both sections topography is shown as dark grey surface and the gravity derived basement as light grey surface. No vertical exaggeration.
Finally we test the sensitivity of our geologically constrained inversion to the initial starting model used to seed the inversion. In the first model we constrained the basement surface to follow the outcrop topography in the Kaimanawa Ranges and under the TgVM made a flat lying surface at 300 m elevation. In the second model we used the mapped faults and prior information about their offsets to construct a geologically feasible starting surface. Geometry inversions on both surfaces using a homogeneous basement density converged on similar results with the same RMS (1.7 mGal) misfit. While the inversion using the flat basement surface starting model did not explicitly recreate faulted offsets, the inverted surface does intersect the locations of the fault traces seen in the inversion of the faulted starting model. The final basement depths between the Poutu and Waihi fault zones agree within 100 m (∼10% of the total depth) in both models, showing that the retrieved shape of the basement surface is mostly insensitive to the initial starting model.

D.8 References


