

1 *Perspectives in Petrology*

2
3 **Magma Plumbing Systems: A Geophysical Perspective**

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26
27 **Abstract**

28 Over the last few decades, significant advances in using geophysical techniques to image the
29 structure of magma plumbing systems have enabled the identification of zones of melt
30 accumulation, crystal mush development, and magma migration. Combining advanced geophysical
31 observations with petrological and geochemical data has arguably revolutionised our understanding
32 of and afforded exciting new insights into the development of entire magma plumbing systems.
33 However, divisions between the scales and physical settings over which these geophysical,
34 petrological, and geochemical methods are applied still remain. To characterise some of these
35 differences and promote the benefits of further integration between these methodologies, we
36 provide a review of geophysical techniques and discuss how they can be utilised to provide a
37 structural context for and place physical limits on the chemical evolution of magma plumbing
38 systems. For example, we examine how Interferometric Synthetic Aperture Radar (InSAR), coupled
39 with Global Positioning System (GPS) and Global Navigation Satellite System (GNSS) data, and
40 seismicity may be used to track magma migration in near real-time. We also discuss how seismic
41 imaging, gravimetry, and electromagnetic data can image contemporary melt zones, magma
42 reservoirs, and/or crystal mushes. These techniques complement seismic reflection data and rock
43 magnetic analyses that delimit the structure and emplacement of ancient magma plumbing systems.
44 For each of these techniques, with the addition of full-waveform inversion (FWI), the use of
45 Unmanned Aerial Vehicles (UAVs), and the integration of geophysics with numerical modelling,
46 we discuss potential future directions. We show that approaching problems concerning magma
47 plumbing systems from an integrated petrological, geochemical, and geophysical perspective will
48 undoubtedly yield important scientific advances, providing exciting future opportunities for the
49 volcanological community.

50

51 **1. Introduction**

52 Igneous petrology and geochemistry are concerned with the chemical and physical mechanisms
53 governing melt genesis, mobilisation, and segregation, as well as the transport/ascent, storage,

54 evolution, and eruption of magma. The reasons for studying these fundamental processes include
55 understanding volcanic eruptions, modelling the mechanical development of magma conduits and
56 reservoirs, finding magma-related economic ore deposits, exploring for active geothermal energy
57 sources, and determining the impact of magmatism in different plate tectonic settings on the
58 evolution of the lithosphere and crustal growth. However, whilst petrological and geochemical
59 studies over the last century have shaped our understanding of the physical and chemical evolution
60 of magma plumbing systems, assessing the distribution, movement, and accumulation of magma in
61 the Earth's crust from these data remains challenging. A key frontier in igneous petrological and
62 geochemical research thus involves deciphering how and where magma forms, the routes it takes
63 toward the Earth's surface, and where exactly it is stored.

64 This contribution will demonstrate how geophysical data can be used to determine the
65 architecture of magma plumbing systems, providing a structural framework for the interpretation of
66 petrological and geochemical data. To aid the alignment of petrological, geochemical, and
67 geophysical disciplines it is first important to delineate what we mean by 'magma'. We follow
68 Glazner *et al.*, (2016) and define magma as, "naturally occurring, fully or partially molten rock
69 material generated within a planetary body, consisting of melt with or without crystals and gas
70 bubbles and containing a high enough proportion of melt to be capable of intrusion and extrusion".
71 Importantly, this definition specifically considers that magma: (i) forms through the migration and
72 accumulation of partial melt that is initially distributed throughout pore spaces in a rock volume;
73 and (ii) is a suspension of particles (i.e. crystals, xenoliths, and/or bubbles) within melt (see
74 Cashman *et al.*, 2017). As magma starts to solidify, the proportion of suspended crystals and thus
75 the relative viscosity of the magma increases until a relatively immobile, continuous network of
76 crystals and interstitial melt develops; we term this a 'crystal mush' (e.g., Hildreth, 2004; Glazner *et*
77 *al.*, 2016; Cashman *et al.*, 2017). The rheological transition from a magma to a crystal mush is
78 partly dependent on its chemistry, but typically occurs abruptly when the particle volume increases
79 across the 50–65% range (Cashman *et al.*, 2017). Crystal mushes thus exist at or above the solidus

80 and generally cannot be erupted, although they may be partly entrained in eruptible magma as
81 glomerocrysts, cumulate nodules, or restite (Cashman *et al.*, 2017). Migration of interstitial melt
82 within a crystal mush can lead to its accumulation and, thus, formation of a magma. A magma
83 plumbing system therefore consists of interconnected magma conduits and reservoirs, which store
84 magma as it evolves into a crystal mush, ultimately fed from a zone of partial melting (e.g., Fig. 1).
85 These definitions are supported by geophysical imaging and analyses of contemporary reservoirs,
86 which show melt volumes in the mid- to upper crust are typically low (<10%) and likely exist
87 within a crystal mush (e.g., Paulatto *et al.*, 2010; Koulakov *et al.*, 2013; Ward *et al.*, 2013;
88 Hammond, 2014; Comeau *et al.*, 2015; Comeau *et al.*, 2016; Delph *et al.*, 2017). These definitions
89 and geophysical data question the traditional view that magma resides in long-lived, liquid-rich, and
90 volumetrically significant magma chambers. Following this, the emerging paradigm for igneous
91 systems is thus that liquid-rich magma chambers are short-lived, transient phenomena with: (i) melt
92 typically residing in mushes that develop through the incremental injection of small, distinct magma
93 batches; and (ii) magma accumulating in thin lenses (e.g., Hildreth, 2004, Annen *et al.*, 2006;
94 Annen, 2011; Miller *et al.*, 2011; Solano *et al.*, 2012; Cashman & Sparks, 2013; Annen *et al.*, 2015;
95 Cashman *et al.*, 2017). We are now starting to view magmatic systems as a vertically extensive,
96 transcrustal, interconnected networks of magma conduits and magma/mush reservoirs (Fig. 1) (e.g.,
97 Cashman *et al.*, 2017).

98 The current use of geophysical techniques within the igneous community can be separated
99 into two distinct areas focused on either characterising active volcanic domains or investigating the
100 structure and emplacement of ancient magma plumbing systems. For example, in areas of active
101 volcanism, our understanding of magma plumbing system structure principally comes from the
102 application of geophysical techniques that detect sites of magma movement or accumulation (e.g.,
103 Sparks *et al.*, 2012; Cashman & Sparks, 2013). Such geophysical techniques include Interferometric
104 Synthetic Aperture Radar (InSAR; e.g., Biggs *et al.*, 2014), seismicity (e.g., recording of
105 earthquakes associated with magma movement; e.g., White & McCausland, 2016), various seismic

106 imaging methods (e.g., Paulatto *et al.*, 2010; Hammond, 2014), gravimetry (e.g., Battaglia *et al.*,
107 1999; Rymer *et al.*, 2005), and electromagnetic techniques (Desissa *et al.*, 2013; Comeau *et al.*,
108 2015). These techniques allow examination of: (i) the temporal development of magma plumbing
109 systems (e.g., Pritchard & Simons, 2004; Sigmundsson *et al.*, 2010); (ii) vertical and lateral
110 movements of magma (e.g., Keir *et al.*, 2009; Jay *et al.*, 2014); (iii) the relationship between
111 eruption dynamics, volcano deformation, and intrusion (e.g., Sigmundsson *et al.*, 2010;
112 Sigmundsson *et al.*, 2015); and (iv) estimates of melt sources and melt fractions (e.g., Desissa *et al.*,
113 2013; Johnson *et al.*, 2016). However, inversion of these geophysical data typically results in non-
114 unique, relatively low-resolution models of subsurface structures. Furthermore, some methods only
115 capture active processes, which may be short-lived or even instantaneous, potentially providing
116 information on only a small fraction of the magma plumbing system.

117 In contrast to the study of active volcanic domains, the analysis of ancient plumbing systems
118 through field observations, geophysical imaging techniques (e.g., reflection seismology, gravity,
119 and magnetic data), and/or rock magnetic experiments can provide critical insights into magma
120 emplacement, mush evolution, and allow the geometry of entire plumbing systems to be
121 reconstructed (e.g., Cartwright & Hansen, 2006; Stevenson *et al.*, 2007a; Petronis *et al.*, 2013;
122 Muirhead *et al.*, 2014; O'Driscoll *et al.*, 2015; Magee *et al.*, 2016). Whilst such studies of ancient
123 plumbing systems provide a framework for interpreting the structure of active intrusion networks,
124 capturing a snapshot of how magma moved and melt was distributed through the system at any one
125 time is difficult because magmatism has long since ceased.

126 All the techniques employed to define active and ancient plumbing systems, including
127 petrological and chemical analyses, provide information at different spatial and/or temporal
128 resolutions. Answering the major outstanding questions in studies of magma plumbing systems
129 therefore requires the integration of complementary petrological, geochemical, geophysical,
130 geochronological, and structural techniques. Here, we examine active plumbing systems using
131 InSAR, seismicity, seismic imaging, gravimetry, and electromagnetic techniques. To provide a

132 context for the interpretation of data pertaining to the active systems, we also discuss how seismic
133 reflection data and rock magnetic techniques can be used to derive the structure and evolution of
134 ancient intrusion networks. The potential of emerging techniques involving seismic full-waveform
135 inversion (FWI) and unmanned aerial vehicles (UAVs) are also considered, as is the role of
136 numerical modelling in bringing together outputs from different datasets. For each technique
137 described, we briefly discuss the methodology and limitations and provide a summary of the key
138 findings and potential uses, with a focus on integration with petrological and geochemical data. The
139 aim of this review is to facilitate and promote integration between petrologists, geochemists,
140 geochronologists, structural geologists, and geophysicists interested in addressing outstanding
141 problems in studies of magma plumbing systems.

142

143 **2. Understanding magma plumbing system structure**

144 Here, we discuss a range of techniques that can be utilised to define different aspects of magma
145 plumbing system structure and evolution. In particular, we describe how InSAR, seismicity, seismic
146 imaging (e.g., seismic tomography), gravity, and electromagnetic data is used to determine melt
147 fractions and distribution, track movement of magma in near real-time, and/or locate sites and
148 examine the evolution of magma/mush storage. Overall, these geophysical techniques allow the
149 structure of active plumbing systems and their transient evolution to be assessed. We also discuss
150 how seismic reflection data can provide unprecedented images of ancient plumbing systems and
151 associated host rock deformation in three-dimensions at resolutions of 10's of metres. Finally, we
152 examine the application of rock magnetic techniques to assess magma flow and crystallisation
153 processes at a range of scales.

154 Although beyond the scope of this review, it is critical to highlight that interpreting the
155 geophysical response of a rock or magma relies on understanding its physical and chemical
156 properties (e.g., density, temperature, and melt fraction). Laboratory experiments testing how rock
157 or magma properties influence geophysically measured parameters (e.g., seismic velocities and

158 resistivity) thus provide context for interpreting magma plumbing system structure and evolution
159 from geophysical data (e.g., Gaillard, 2004; Pommier *et al.*, 2010; Pommier, 2014).

160

161 **2.1. Insights into magma plumbing systems from ground deformation data**

162 *Technique*

163 Changes in volume within magma plumbing systems can deform the host rock, potentially resulting
164 in displacement of the Earth's surface. Such displacements are a unique source of information for
165 volcanologists and can be modelled to estimate geodetic source depth and, to varying extents, the
166 source geometry and volume change (e.g., Segall, 2010). Measuring the deformation of the Earth's
167 surface can thus provide information about the characteristics and timing of magma movement and
168 accumulation, as well as variations in internal reservoir conditions. Traditionally, deformation
169 measurements are made using levelling, electronic distance meters, tiltmeters, and Global
170 Positioning System (GPS), all of which have proven to be reliable methods and thus are widely
171 used in volcano monitoring (e.g., Dzurisin, 2006). For example, GPS measurements retrieve the
172 relative positions of receivers on Earth's surface from dual frequency carrier phase signals
173 transmitted from GPS or Global Navigation Satellite System (GNSS) satellites with precisely
174 known orbits. Distances between satellites and receivers are assessed from the travel-time, i.e. the
175 measured difference between the transmitted and received times of a unique ranging code, allowing
176 movement of the Earth's surface over time to be monitored (see review by Dixon, 1991).
177 Permanently installed receivers record position data continuously, but receivers can also be
178 deployed for a limited time during GPS campaigns to provide additional measurements, normally
179 made relative to a standard benchmark location (e.g., Dvorak & Dzurisin, 1997). Whilst tiltmeters
180 and GPS can provide continuous measurements, their spatial resolution is limited by logistical
181 constraints such as cost and accessibility, which may be restricted at active volcanoes.

182 The geographic reach of volcano geodesy has been greatly expanded over the past two
183 decades by the application of Interferometric Synthetic Aperture Radar (InSAR), an active remote

184 sensing technique that uses microwave electromagnetic radiation to image the Earth's surface (e.g.,
185 Simons & Rosen, 2007; Pinel *et al.*, 2014). Surface displacements can be measured by constructing
186 interferograms, where the difference in phase between radar echoes from time-separated images
187 appear as 'fringes' of variation in the line of sight distance to the satellite (Fig. 2). The patterns of
188 fringes in individual interferograms are distinctive for different deformation source geometries,
189 such as for horizontal (sill-like) or vertical (dyke-like) opening of intrusions, or the pressurisation of
190 a spheroidal reservoir (i.e. a Mogi source) (e.g., Fig. 2b). However, magma intrusion processes can
191 rarely be uniquely identified from geodetic source geometry alone, and distinguishing between
192 magmatic, hydrothermal, structural (e.g., faulting and compaction), and combinations of elastic and
193 inelastic sources is particularly challenging (e.g., Galland, 2012; Holohan *et al.*, 2017).

194 Whilst a single interferogram only provides displacements in satellite line-of-sight, a
195 pseudo-3D displacement field can be estimated by combining multiple images from polar orbits that
196 are ascending (i.e. satellite moves roughly northward, looking east) and descending (i.e. satellite
197 moves roughly southward, looking west) (Fig. 2a), especially where GNSS measurements can also
198 be incorporated. The lateral spatial resolution of most InSAR data is on the order of metres to tens
199 of metres, whilst vertical movements can be resolved on the order of centimetres and sometimes
200 millimetres. Temporal resolution depends on the satellite revisit time and ranges between days to
201 months depending upon the sensor type and satellite orbit. This means that InSAR can be used to
202 regularly assess ground deformation at virtually any volcano worldwide situated above sea level,
203 with a higher spatial density of measurements than achieved using from ground-based
204 instrumentation.

205 Magmatic processes are only observable by InSAR when either magma movement or
206 internal reservoir processes (e.g., cooling and contraction, phase changes) cause changes in pressure
207 and thereby instigate deformation of the host rock and free surface. The best-fit parameters of a
208 deformation source (e.g., an intruding magma body) are most often assessed by inverting measured
209 displacements using analytical elastic-half space models of simple source geometries, although

210 there are often trade-offs between parameters such as source depth and volume change (e.g.,
211 Pritchard & Simons, 2004). Complex and more realistic deformation source geometries may be
212 retrieved using finite element-based linear inversion of displacement fields (e.g., Ronchin *et al.*,
213 2017). A proportion of any pressure change may be accommodated by magma compressibility,
214 leading to underestimation of volume changes (e.g., Rivalta & Segall, 2008; McCormick-Kilbride *et*
215 *al.*, 2016). Assessing both volume changes and especially the total volume of a magma reservoir
216 from geodetic data therefore remains challenging. Furthermore, host rocks in areas of repeated
217 intrusion that have been heated above the brittle-ductile transition are better described by a
218 viscoelastic rheology (e.g., Newman *et al.*, 2006; Yamasaki *et al.*, 2018), while ductile
219 accommodation of volume changes may occur at greater depth. Where some constraints are
220 available for the structure and rheology of Earth's crust, finite or boundary element models may
221 achieve a more realistic model of the deformation source (e.g., Masterlark, 2007; Hickey *et al.*,
222 2017; Gottsmann *et al.*, 2017).

223

224 ***Observations***

225 Measurements of volcano deformation preceding and/or accompanying eruption have provided
226 insights into the extent and structure of magma plumbing systems and, in some instances, the
227 dynamics of magma movement. For example, InSAR-based observations at Eyjafjallajökull,
228 Iceland have recognised the intrusion of multiple, distinct sills over a decade and their subsequent
229 extraction when tapped during an explosive eruption (e.g., Pedersen & Sigmundsson, 2006;
230 Sigmundsson *et al.*, 2010). Extensive lateral connections via dykes and sills between reservoirs
231 and/or volcanoes have been illuminated by eruptions or unrest accompanied by ground deformation
232 tens of kilometres away, and by the existence of multiple deformation sources (e.g., Alu-Dalafilla
233 shown in Figures 3 and b, Pagli *et al.*, 2012; Korovin, Lu & Dzurisin, 2014; Cordon-Caulle, Jay *et*
234 *al.*, 2014; Kenyan volcanoes, Biggs *et al.*, 2014; global synthesis, Ebmeier *et al.*, 2018). Inter-
235 eruptive deformation at calderas is especially complex and seems to be particularly frequent and

236 high magnitude (e.g., Laguna del Maule; Fournier *et al.*, 2010; Singer *et al.*, 2014; Le Mével *et al.*,
237 2015), with the location of the deformation sources inferred to vary over time (e.g., Campi Flegrei,
238 Trasatti *et al.*, 2004; Yellowstone, Wicks *et al.*, 2006). The geometries of dykes and sills inferred
239 from InSAR data inform our understanding of changing subsurface stress fields (e.g., Afar,
240 Hamling *et al.*, 2010; Fernandina, Bagnardi *et al.*, 2013), as do measurements of displacements
241 caused by moderate earthquakes in close proximity to magma plumbing systems (e.g., Kilauea,
242 Wauthier *et al.*, 2013; Chiles-Cerro Negro, Ebmeier *et al.*, 2016).

243 At a transcrustal scale, deformation measurements have contributed to evidence for temporal
244 variations in magma supply rates (e.g., in Hawaii, Poland *et al.*, 2012). Volume increases in the
245 mid- to lower-crust, notably in the Central Andes, have provided the first observations of deep
246 pluton growth (Pritchard & Simons, 2004). Furthermore, uplift during episodes of unrest that have
247 not (yet) resulted in eruption have been detected at a broad range of volcanoes (e.g., Westdahl,
248 Mount Peulik, Lu & Dzurisin, 2014; Alutu and Corbetti, Biggs *et al.*, 2011) and, in some cases,
249 have been interpreted as evidence for the ‘pulsed’ accumulation of potentially eruptible magma
250 (e.g., Santorini, Parks *et al.*, 2012). In addition to magma movement, volume changes associated
251 with internal reservoir processes can also cause deformation of the host rock and free surface. For
252 example, InSAR measurements have recorded subsidence linked to cooling and crystallisation of
253 sills (Medicine Lake, Parker, 2016; Taupo Volcanic Zone, Hamling *et al.*, 2015). Transient periods
254 of subsidence during inter-eruptive uplift have been attributed to phase transitions in response to the
255 addition of more juvenile magma (e.g., Okmok, Caricchi *et al.*, 2014).

256

257 ***Implications and integration***

258 InSAR has increased the number of volcanoes where measurements of ground deformation have
259 been made, from less than 50 in the late 1990s to over 200 today (Biggs & Pritchard, 2017; Ebmeier
260 *et al.*, 2018). This increase in coverage has been particularly influential in the developing world
261 where monitoring infrastructure is typically poor (Ebmeier *et al.*, 2013; Chaussard *et al.*, 2013),

262 with InSAR often providing the first evidence of magmatic activity at many volcanoes previously
263 considered to be inactive (e.g., Pritchard & Simons, 2004; Biggs *et al.*, 2009; Biggs *et al.*, 2011; Lu
264 & Dzurisin, 2014). A continued increase in the number and range of satellite- and large-scale UAV-
265 based SAR instruments, as well as enhancements to their spatial and temporal resolution, will allow
266 the detection of a greater range of volcanic ground deformation (e.g., Salzer *et al.*, 2014; Schaefer *et*
267 *al.*, 2015; Stephens *et al.*, 2017). Overall, improved InSAR coverage will also increase the number
268 of volcanoes where deformation measurements have been made across multiple cycles of eruption
269 and deformation, increasing its usefulness for both hazard assessment and for characterising the
270 extent, geometry, and changes in magma plumbing systems.

271 Geodetic measurements provide information only about the parts of a plumbing system that
272 are currently active, and do not necessarily reflect the full extent and character of the intrusion
273 network (e.g., Sigmundsson, 2016). However, geodetic analyses of ground deformation provide
274 critical insight into the spatial and temporal development of active plumbing systems. Comparing
275 observations of ancient plumbing systems (e.g., Magee *et al.*, 2013; Schofield *et al.*, 2014),
276 integration of ground deformation measurements with petrological observations (e.g., Caricchi *et*
277 *al.*, 2014; Jay *et al.*, 2014) or thermal models (Parker *et al.*, 2016), as well as tomographic
278 geophysical imaging, will increase the sophistication of models of magmatic systems. Integrating
279 InSAR with gravity or electromagnetic measurements is particularly powerful, as it can allow
280 discrimination between melt, volatiles, and hydrothermal fluids for which deformation signals are
281 similar (see section 2.4) (e.g., Tizzani *et al.*, 2009).

282

283 **2.2. Seismicity and magma plumbing systems**

284 ***Technique***

285 Seismicity (i.e. earthquakes) at volcanoes is primarily caused by the dynamic interaction of magma
286 and hydrothermal fluids with the solid host rock (e.g., Chouet & Matoza, 2013), as well as by
287 fracturing and fragmentation of silicic magma (e.g., Tuffen *et al.*, 2008). There are a number of

288 primary physical mechanisms for causing volcano seismicity (e.g., faulting), each of which
289 typically produces seismic signals of specific frequency content (Chouet & Matoza, 2013).
290 Recording and isolating different volcano seismicity signals therefore allows a variety of plumbing
291 system processes to be assessed. The majority of volcano monitoring agencies have now deployed
292 or aim to use a network of distributed seismic sensors, including broadband seismometers, to
293 monitor volcano activity (Neuberg *et al.*, 1998; Sparks *et al.*, 2012). Furthermore, an increase in
294 computing power and reduction in cost of seismic sensors means that researchers are now
295 developing fast, fully automated detection and real-time location techniques that can locate
296 seismicity to sub-decimetre precision (e.g., Drew *et al.*, 2013; Sigmundsson *et al.*, 2015).

297

298 ***Observations***

299 Volcano-tectonic (VT) seismicity generally produces relatively high frequency (1–20 Hz), short
300 period signals, involving clear primary (P), secondary (S), and surface waves, which are caused by
301 displacement on new or existing faults in the host rock in response to fluid-induced stress changes
302 (e.g., Rubin & Gillard, 1998; Roman & Cashman, 2006; Tolstoy *et al.*, 2008). These earthquakes
303 commonly occur near the propagating edge of intrusions, meaning the space-time evolution of VT
304 earthquake locations can be used to track the horizontal and vertical growth of sills and dykes (e.g.,
305 Keir *et al.*, 2009; Sigmundsson *et al.*, 2010; Sigmundsson *et al.*, 2015). Inflation of a magma or
306 mush body can also induce VT seismicity on any preferentially oriented faults surrounding the
307 intrusion, thereby recording the delivery time and locus of new magma injected into a reservoir
308 (e.g., Roman & Cashman, 2006; Vargas-Bracamontes & Neuberg, 2012).

309 Earthquakes with longer period seismic signals and low-frequencies (0.5–2 Hz) are thought
310 to be generated near the interface between magma and solid rock (Chouet & Matoza, 2013). The
311 earthquake source proximity to the magma causes the seismic signal to resonate in parts of the
312 plumbing system (e.g., conduits, dykes, and cracks), leading to a reduction in its frequency content
313 (Chouet & Matoza, 2013). These earthquakes can potentially be caused by stick-slip motion

314 between the magma and wall-rock or fracturing of cooling magma near the conduit wall (Neuberg
315 *et al.*, 2006; Tuffen *et al.*, 2008). Such earthquakes typically occur at restricted portions of conduits
316 where the magma flow and shear strain rate are highest (Neuberg *et al.*, 2006; Tuffen *et al.*, 2008).

317 Very long period seismicity (VLP) of 10s of seconds to several minutes period are typically
318 attributed to inertial forces associated with perturbations in the flow of magma and gases through
319 conduits (Chouet & Matoza, 2013). These signals can record the response of the host rock to
320 reservoir inflation and deflation and may be used to model conduit shape and size (Chouet *et al.*,
321 2008). To do this requires a better understanding of the links between flow processes and resultant
322 pressure/momentum changes using laboratory experiments and numerical models that include the
323 elastic response to magma flow across multiple signal frequency bands (e.g., Thomas & Neuberg,
324 2012).

325

326 ***Implications and integration***

327 Studies of evolving reservoirs now aim to link episodes of seismicity related to new magma
328 injection to petrological evidence for timing of reservoir recharge events, thereby providing
329 independent constraints on day to year-long time-scales of magma residence and input prior to
330 eruptions. For example, Fe-Mg diffusion chronometry modelling of orthopyroxene crystals from the
331 1980–1986 eruption of Mount St. Helens indicates that compositionally distinct rims grew within
332 12 months prior to eruption (Fig. 4) (Saunders *et al.*, 2012). Peaks in crystal growth correlated
333 extremely well with increased seismicity and SO₂ flux (Fig. 4), confirming the relationship between
334 seismicity and magma movement, as well as demonstrating how a combination of seismicity and
335 petrological information can be used to detect magma injections (Saunders *et al.*, 2012).

336 Petrology and seismicity can also be integrated with other methods, such as GPS and
337 InSAR. Field *et al.*, (2012) analysed volatiles in melt inclusions trapped in phenocrysts within
338 peralkaline lavas from historic eruptions at the Dabbahu Volcano in Afar, Ethiopia. Volatile
339 saturation pressures at typical magmatic temperatures were constrained to be in the range 43–207

340 MPa, consistent with the phenocryst assemblage being stable at 100–150 MPa. The interpreted
341 magma/mush storage depths for these historic eruptions are ~1–5 km, consistent with the depths of
342 earthquakes associated with reservoir inflation following dyke intrusion in 2005–2006 (Fig. 5)
343 (Ebinger *et al.*, 2008; Field *et al.*, 2012). Additionally, the best-fit result for modelling of uplift
344 patterns recorded by InSAR data, which were collected over the same time period as seismicity
345 measurement, suggests the magma/mush reservoir comprises a series of stacked sills over a ~1–5
346 km depth range (Fig. 5) (Ebinger *et al.*, 2008). The consistency of depth estimates based on
347 petrological study of ancient eruptions, along with the seismicity and inflation of the Dabbahu
348 Volcano following axial dyke intrusion in 2005–2006, implies a vertically extensive and potentially
349 long-lived magma/mush storage region. Such multidisciplinary studies demonstrate that joint
350 observations and modelling of seismic signals, petrological data, and other techniques (e.g.,
351 geodesy and gas emissions) significantly strengthen interpretation of the physical structure,
352 emplacement, and evolution of magma plumbing systems.

353

354 **2.3. Identifying melt in plumbing systems using seismic imaging**

355 ***Techniques***

356 Both active and passive source seismological techniques, which utilise man-made seismic events
357 and natural earthquakes respectively, can be used to identify areas where the presence of partial
358 melt or magma causes a local reduction in seismic wavespeed, an increase in anisotropy, or an
359 increase in attenuation (e.g., Berryman, 1980; Hammond & Humphreys, 2000a, b). With the recent
360 availability of dense seismic networks, resolution of the crust and mantle seismic velocity structure
361 has improved to the degree that active source seismic experiments can: (i) use tomographic
362 techniques to image likely storage regions in the upper crust beneath ocean island volcanoes (e.g.,
363 Soufrière Hills Volcano, Montserrat; Fig. 6) (Paulatto *et al.*, 2010; Shalev *et al.*, 2010) and,
364 occasionally, onshore volcanoes (e.g., Mt Erebus, Antarctica, Zandomenighi *et al.*, 2013; Mt. St.
365 Helens, Kiser *et al.*, 2014); and (ii) utilise reflected data to image individual sills beneath mid-ocean

366 ridges (e.g., Kent *et al.*, 2000, Marjanovic *et al.*, 2014). A further example from Katla volcano
367 Iceland, demonstrates how active source seismic experiments can be used to identify S-wave
368 shadow zones (i.e. S-waves cannot travel through fluids) and delays in P-waves, which may be used
369 to infer the location and geometry of shallow-level magma reservoirs (Gudmundsson *et al.*, 1994).
370 However, recent modelling approaches suggest that the upper crust likely represents only a small
371 portion of magma plumbing systems and long-term storage is dominated by mushy zones
372 throughout the lower crust (e.g., Annen *et al.*, 2006). Active source seismic experiments,
373 particularly on land where the crust is thick and coverage less uniform, cannot penetrate to these
374 depths efficiently. Furthermore, whilst seismic tomographic methods using local earthquakes offer
375 3D images of crustal velocity beneath many volcanoes (e.g., Mt. St. Helens, Waite & Moran, 2009;
376 Askja, Iceland, Mitchell *et al.*, 2013), they can only resolve areas directly above the deepest
377 earthquakes. Non-uniform coverage thus makes interpreting tomographic images difficult as
378 resolution varies across the model (see review by Lees, 2007).

379 To illuminate lower crustal regions, seismologists rely on passive seismology. Extending
380 seismic tomographic images of magma plumbing systems to lower crustal depths requires the use of
381 teleseismic body-wave and surface wave data, which emanate far (>1000 km) from the
382 measurement site. However, these data are dominated by longer period signals, meaning their
383 resolution is relatively low. For example, the Fresnel zone (i.e. the region within $\frac{1}{4}$ seismic
384 wavelength and an estimate of the minimum resolvable structure) for active source data at 10 Hz is
385 on the order of 3 km in the upper crust compared to 10–15 km for 1 Hz teleseismic data used in
386 receiver function or tomography studies.

387

388 ***Observations***

389 Active and passive seismological techniques provide crucial insight into transcrustal melt and
390 magma distribution. For example, P-wave seismic travel-time tomography across Monserrat and the
391 Soufrière Hills Volcano images a series of relatively fast seismic velocity zones, which are

392 interpreted as solidified andesitic intrusions, surrounded by regions of slow seismic velocities likely
393 related to either areas of hydrothermal alteration or buried volcanoclastic deposits (Fig. 6) (Paulatto
394 *et al.*, 2010; Shalev *et al.*, 2010). Within the lower crust, inversions using surface wave data
395 generated by ambient seismic noise and receiver function data, which isolates P-wave to S-wave
396 conversions at major discontinuities in the earth, have identified low shear-wave velocities probably
397 related to melt presence beneath several volcanic settings (e.g., New Zealand, Bannister *et al.*, 2007;
398 Toba, Sumatra, Stankiewicz *et al.*, 2010; Ethiopia, Hammond *et al.*, 2011; Jaxybulatov *et al.*, 2014;
399 Costa Rica, Harmon & Rychert, 2015).

400 When trying to determine how much melt or magma is present, numerous studies have
401 shown that seismic velocities are more sensitive to the shapes of melt/magma-filled spaces on a
402 range of scales compared to the melt fraction (e.g., Hammond & Humphreys, 2000a, b; Miller &
403 Savage, 2001; Johnson & Poland, 2013; Hammond & Kendall, 2016). On the grain-scale, melt
404 commonly wets grain boundaries, forming planar pockets (e.g., Takei, 2002; Garapic *et al.*, 2013;
405 Miller *et al.*, 2014), whereas on the larger scale magma may form planar intrusions of either mush
406 (e.g., Annen *et al.*, 2006), or liquid-rich dykes or sills. If these features are preferentially aligned,
407 they will appear as a distributed region of melt to seismic waves and the analyses described will not
408 be able to discriminate between a melt-poor region dominated by aligned melt-pockets on grain
409 boundaries and an elongate melt-rich body such as an intrusion (e.g., Hammond & Kendall, 2016).
410 A further problem is that seismic velocities are affected by variations in temperature (Jackson *et al.*,
411 2002), composition (Karato & Jung, 1998), and attenuation (Goes *et al.*, 2012). Relating seismic
412 velocity anomalies to melt fraction is therefore difficult without some prior knowledge of melt
413 distribution (Hammond & Kendall, 2016).

414 One possible approach to investigate melt distributions further is through measuring seismic
415 anisotropy. If melt has some preferential distribution on a length-scale smaller than the seismic
416 wavelength, such as a stacked network of sills or an anisotropic permeability on the grain scale,
417 then the seismic wavespeed will vary with direction of propagation, i.e. be anisotropic. As a result,

418 measuring the effects of seismic anisotropy allows inferences about sub-seismic wavelength
419 structures, leading to estimates of the preferential orientation of melt distribution. It is common to
420 observe strong anisotropy beneath volcanoes and this has been used to place constraints on melt
421 distribution. For example, high degrees of shear-wave splitting from volcanic earthquakes can either
422 directly map out regions of significant quantities of melt aligned in pockets (Keir *et al.*, 2011), or
423 map out stress changes related to overpressure from injections of magma into the upper crust (Gerst
424 & Savage, 2004; Roman *et al.*, 2011). To image the deeper crustal magmatic system, azimuthal
425 variations in the ratio of P-wave to S-wave speeds (i.e. V_p/V_s) from receiver functions led to the
426 interpretation that a stacked network of sills is present in the lower crust beneath the Afar
427 Depression, Ethiopia (Hammond, 2014). Differences in the velocity of Rayleigh Waves and Love
428 Waves, which are vertically polarised shear-waves and horizontally polarised shear waves
429 respectively, suggest a similar anisotropic melt distribution is present beneath the Toba Caldera,
430 Sumatra (Jaxybulatov *et al.*, 2014) and Costa Rica (Harmon & Rychert, 2015).

431

432 ***Implications and integration***

433 Due to the interference of signals denoting the geometry of melt-filled pockets and the volumetric
434 proportion of that melt, estimating melt fraction remains difficult using seismology alone. Some
435 attempt has been made to directly infer magma/mush reservoir properties from seismic velocities.
436 For example, Paulatto *et al.*, (2012) used thermal modelling to test the range of melt fractions that
437 could account for the low velocity zones imaged in the upper crust beneath Soufrière Hills Volcano
438 (Fig. 6), Montserrat and concluded the melt fraction is between 3 and 10%. However, accounting
439 for resolution of the tomography, together with uncertainties in the distribution and geometry of
440 melt, means >30% melt may be present more locally in the low velocity zones defined beneath
441 Soufrière Hills Volcano (Paulatto *et al.*, 2012). Possible ways forward involve integrating
442 seismological data with: (i) petrological data that can place limits on likely melt fractions and/or
443 emplacement depths (e.g., McKenzie & O’Nions, 1991; Comeau *et al.*, 2016); (ii) geochemical

444 techniques that can help determine timescales of melt and magma evolution (e.g., Hawkesworth *et*
445 *al.*, 2000); and (iii) geodetic or other monitoring data, which helps determine magma movement
446 (Sturkell *et al.*, 2006). Recent efforts applying industry software, such as full waveform inversions
447 (FWI; Warner *et al.*, 2013), which is discussed in section 3.1, are also pushing the potential
448 application of seismological data further and mean that it may be possible to resolve features to sub-
449 kilometre levels, particularly in the upper crust. Together, these techniques may allow us to directly
450 relate seismic velocity anomalies to melt fractions and distributions in the whole crust.

451

452 **2.4. Studying magma plumbing systems using gravimetry**

453 *Techniques*

454 Gravimetry measures the gravitational field and its changes over space and time, which can be
455 related to variations in the subsurface distribution and redistribution of mass (e.g., magma). A
456 variety of gravimeter instruments (e.g., free-fall, superconducting, and spring-based) and techniques
457 (e.g., ground-based, sea-floor, ship-borne, and air-borne instrumentations) are available. Spring
458 gravimeters, where a test mass is suspended on a spring, are mostly used to study magmatic and
459 volcanic processes in ground-based surveys (e.g., Carbone *et al.*, 2017; Van Camp *et al.*, 2017).
460 Changes in the gravitational acceleration across a survey area shorten or lengthen the spring, which
461 is recorded electronically and converted to gravity units. These changes are evaluated across a
462 survey network in relation to a reference and are hence termed ‘relative measurements’. Absolute
463 gravimetry can also be measured, i.e. the value of gravitational acceleration, and serves primarily to
464 create a reference frame into which other geodetic methods (e.g., InSAR, GNSS, levelling, relative
465 gravimetry) can be integrated for joint data evaluation. Recent reviews by Carbone *et al.*, (2017)
466 and Van Camp *et al.*, (2017) provide a broad account of gravimetric instruments, measurement
467 protocols, and data processing relevant for the study of magmatic systems.

468 Static gravimetric techniques obtain a single snap-shot of the subsurface mass distribution.

469 For example, Bouguer anomaly maps are perhaps the best-known products of static gravity surveys

470 and capture spatial variations in gravity over an area of interest, providing insight into anomalous
471 mass distribution in the subsurface. Within magmatic studies, computational modelling and
472 inversion of Bouger anomaly data allows identification of shallow intrusions (e.g., dykes and sills;
473 Rocchi *et al.*, 2007), magma-related ore bodies (Hammer, 1945; Bersi *et al.*, 2016), and plutons
474 (e.g., Figs 7a and b) (e.g., Vigneresse, 1995; Vigneresse *et al.*, 1999; Petford *et al.*, 2000) exhibiting
475 a density contrast with their host rocks.

476 In contrast to static surveys, dynamic gravimetric observations allow spatio-temporal mass
477 changes to be tracked. Dynamic gravimetric studies investigate how the subsurface architecture
478 changes over time and is usually performed by measuring variations in gravity across a network of
479 survey points (e.g., Fig. 7c) or, in a few exceptional cases, by installing a network of continuously
480 operating gravimeters. Dynamic observations demand one-to-two orders of magnitude higher data
481 precision (i.e. to a few μGal where $1 \mu\text{Gal} = 10^{-8} \text{ m/s}^2$) compared to static surveys, making them an
482 elaborate and time-consuming exercise. However, dynamic gravity data yields important insights
483 into the source processes behind non-tectonic volcano and crustal deformation, particularly if
484 combined with surface deformation data (e.g., InSAR and GNSS) as subsurface mass and volume
485 changes can be employed to characterise the density of the material behind the stress changes (Figs
486 7c-f and 8) (e.g., Battaglia & Segall, 2004; Jachens & Roberts, 1985; Poland & Carbone, 2016).
487 There are also cases where volcano unrest, due either to magma intrusion into a ductile host rock or
488 to volatile migration at shallow depths, does not result in resolvable surface deformation; in these
489 scenarios, gravity data have provided vital clues about subsurface processes otherwise hidden from
490 conventional monitoring techniques (e.g., Gottsmann *et al.*, 2006; Gottsmann *et al.*, 2007; Miller *et*
491 *al.*, 2017).

492 Whilst static and dynamic gravimetric observations offer considerable insight into the
493 structure and dynamics of magma plumbing systems, care must be exercised when collecting and
494 interpreting gravity data from active magmatic areas where seasonal variations in hydrothermal
495 systems, aquifers, or the vadose zone can influence subsurface mass distribution (e.g., Hemmings *et*

496 *al.*, 2016). These seasonal changes can, in some cases, result in data aliasing artefacts and inhibit
497 the quantification of deeper-seated magmatic processes (e.g., Gottsmann *et al.*, 2005; Gottsmann *et*
498 *al.*, 2007).

499

500 ***Observations***

501 Gravimetric investigations have been at the heart of studies into the subsurface structure of active
502 and ancient magma plumbing systems for more than 80 years (e.g., Carbone *et al.*, 2017; Van Camp
503 *et al.*, 2017). Using techniques initially designed for imaging salt domes, silicic plutons were the
504 first components of magma plumbing systems to be examined using gravimetry because their low
505 density relative to surrounding rocks produces clear, negative gravity anomalies of ~10 to ~40 mGal
506 amplitude (e.g., Reich, 1932; Bucher, 1944; Bott, 1953). Gravity data have been instrumental in the
507 investigation of upper-crustal, silicic magma plumbing systems, helping to reveal: (i) the 3D
508 geometry of plutons by allowing floor morphologies (e.g., flat-floored or wedge-shaped) to be
509 determined (e.g., Vigneresse *et al.*, 1999; Petford *et al.*, 2000); and (ii) how plutons are constructed,
510 for example, by the amalgamation of multiple intrusions fed from depth by dykes (e.g., Vigneresse,
511 1995). Furthermore, recent high-precision static surveys over active silicic volcanoes have enabled
512 detailed modelling of the sub-volcanic magma plumbing system, commonly demonstrating the
513 occurrence of vertically extensive, transcrustal magma bodies (Figs 7a and b) (e.g., Gottsmann *et*
514 *al.*, 2008; del Potro *et al.*, 2013; Saxby *et al.*, 2016; Miller *et al.*, 2017). In addition to examining
515 silicic magma plumbing systems, negative gravity anomalies with typical amplitudes of up to 60
516 mGal and up to 100 km wavelength can be associated with, and provide insight into, the geometry
517 and size of silicic ash-flow calderas (e.g., Eaton *et al.*, 1975; Masturyono *et al.*, 2001). Positive
518 gravity anomalies with amplitudes of up to 30 mGal and wavelengths of up to 20 km are
519 commonly identified at mafic volcanoes and likely result from dense intrusive complexes (e.g.,
520 Rymer & Brown, 1986).

521 Dynamic gravity observations have provided unprecedented insight into the evolution of
522 magma plumbing systems over timescales of seconds to decades, including: (i) the characterisation
523 of multi-year lava lake dynamics (e.g., Poland & Carbone, 2016); (ii) mass budgets of magma
524 intrusions (e.g., Fig. 8) (e.g., Battaglia *et al.*, 1999; Jousset *et al.*, 2000; Rymer *et al.*, 2005;
525 Bonforte *et al.*, 2007; Tizzani *et al.*, 2009); (iii) shallow hydrothermal fluid flow processes induced
526 by deeper magmatic unrest (e.g., Battaglia *et al.*, 2006; Gottsmann *et al.*, 2007; Miller *et al.*, 2017);
527 and (iv) parameters of magmatic geothermal reservoirs (e.g., Hunt & Bowyer, 2007; Sofyan *et al.*,
528 2011). For example, using data from a network of continuously recording gravimeters, Carbone *et*
529 *al.*, (2013) calculated the density of the Kilauea lava lake as $950 \pm 300 \text{ kg m}^3$, i.e. similar to and
530 potentially less than that of water, suggesting that the magma column within the upper portions of
531 the volcanic edifice is gas-rich. Because density and volatile content are critical controls on magma
532 rheology, identification of a gas-rich magma column and lava lake at Kilauea is crucial to modelling
533 and understanding convection and eruption dynamics (Carbone *et al.*, 2013).

534

535 ***Implications and integration***

536 The advent of data-rich geodetic observations from satellite-remote sensing (e.g., InSAR), in
537 conjunction with spatio-temporal gravity studies, provides unprecedented opportunities to
538 characterise magma plumbing system dynamics and the driving mechanisms behind volcano
539 deformation. At Long Valley caldera, for example, a residual gravity increase of more than $60 \text{ } \mu\text{Gal}$
540 between 1982 and 1999 indicates a mass addition at depth (Battaglia *et al.*, 1999). Joint inversion of
541 InSAR and gravity data from Long Valley derives a best fit-source density of 2509 kg m^3 and is
542 indicative of a magmatic intrusion (Fig. 8) (Tizzani *et al.*, 2009). At the deforming Laguna del
543 Maule volcanic centre, Chile, multi-year InSAR and dynamic gravity records demonstrate that
544 uplift and extension above an inflating sill-like reservoir at $\sim 5 \text{ km}$ depth promoted migration of
545 hydrothermal fluids along a fault to shallow (1–2 km) depths (Miller *et al.*, 2017). Alternatively,
546 although no ground deformation is observed at Tenerife, Spain, deconvolution of dynamic gravity

547 into a shallow and deep gravity field provides evidence of unrest (Prutkin *et al.*, 2014). The gravity
548 data suggest hybrid processes have generated the unrest, whereby fluids were released and migrated
549 upward along deep-rooted faults from an intrusion at ~9 km beneath the summit of Teide Volcano
550 (Prutkin *et al.*, 2014). Overall, combining ground deformation and gravimetric observations has
551 highlighted complex processes both within magma reservoirs (e.g., mass addition by magma input,
552 density decrease by volatile exsolution, or density increase by crystallisation; Figs 7c-f) and in the
553 surrounding host rock (e.g., migration of magmatic fluids, phase changes in hydrothermal systems).
554 Key to a better understanding of the processes governing these magma plumbing system and
555 volcano deformation dynamics is the integration of gravimetric and geodetic data with other
556 geophysical data (e.g., seismicity or magnetotellurics) and petrological data. Coupled with
557 advanced numerical modelling, such multi-parameter studies promise exciting new insights into the
558 inner workings of sub-volcanic magma plumbing systems (e.g., Currenti *et al.*, 2007; Hickey *et al.*,
559 2016; Currenti *et al.*, 2017; Gottsmann *et al.*, 2017; Miller *et al.*, 2017).

560

561 **2.5. Resolving magma plumbing system structure with electromagnetic methods**

562 ***Techniques***

563 Electromagnetic (EM) methods probe subsurface electrical resistivity or its inverse, i.e. electrical
564 conductivity. Spatial variations in resistivity control the position, strength, and geometry of local
565 electrical eddy currents and the magnetic fields they produce. These electrical eddy currents are
566 induced by time-varying, naturally occurring magnetic fields external to Earth, which forms the
567 basis of the magnetotelluric (MT) technique, or by controlled sources. Monitoring these decaying
568 electrical and magnetic fields with passive MT techniques therefore allows the subsurface resistivity
569 distribution to be inferred. Controlled source methods generally probe only the shallow subsurface,
570 but MT has a greater depth range as it uses longer-period signals to penetrate deeper. The signals
571 propagate diffusively, which means EM methods typically have a lower resolution than seismic
572 techniques. However, melt, magma, and magmatic hydrothermal fluids are generally considerably

573 less resistive than solid rock and can thus easily be detected by EM methods, which are sensitive to
574 conductive materials (e.g., Whaler & Hautot, 2006; Wannamaker *et al.*, 2008; Desissa *et al.*, 2013;
575 Comeau *et al.*, 2015). EM methods, particularly MT, have therefore been used extensively to study
576 magmatic systems in various tectonic settings.

577 MT equipment, data acquisition, and processing is described by Simpson & Bahr (2005) and
578 Ferguson (2012). Measured field variations have very low amplitudes, meaning equipment needs to
579 be positioned and installed carefully to reduce vibrational (e.g., from wind, vegetation, or vehicles)
580 and electrical (e.g., from power lines) noise. If data are recorded synchronously at a second, less
581 noisy site, remote reference methods can be used to improve the data quality (e.g., Gamble *et al.*,
582 1979). One further problem is that small-scale resistivity anomalies in the shallow subsurface
583 generate galvanic (non-inductive) effects that distort MT data. The distortion is identified and
584 corrected for, which may involve using controlled source transient electromagnetic data to ensure
585 complete removal (e.g. Sternberg *et al.*, 1988), at the same time as assessing whether the data can
586 be modelled with a one-, two- or three-dimensional resistivity structure (e.g. Jones, 2012). Failure
587 to remove galvanic distortion can result in models having resistivity features at the wrong depth.
588 For example, there has been controversy as to whether a conductor beneath Vesuvius Volcano, Italy
589 is caused by a deep (~8–10 km depth) magma reservoir (Di Maio *et al.*, 1998) or a shallow brine
590 layer (Manzella *et al.*, 2004). All of these factors can be a significant problem when using MT to
591 study magmatic systems, especially on volcanic islands.

592 The relationship between MT data and subsurface resistivity is strongly non-linear meaning
593 that inversion is fundamentally non-unique and computationally expensive (e.g., Bailey, 1970;
594 Parker, 1980; Weaver, 1994). Most practical algorithms for inverting MT data obtain a unique
595 result by minimising a combination of misfit to the data and a measure of model roughness (e.g.,
596 Constable *et al.*, 1987). This approach poorly delimits how magma is distributed in the subsurface,
597 whether it is in sills, dykes, or larger reservoirs (Johnson *et al.*, 2016). Whilst MT data are sensitive
598 to the top surface of a conductor, its base may not be detected because conductive material reduces

599 the penetration depth of the signal. Sensitivity analysis is used to ascertain the model features
600 required to fit the MT data, which allows a conductor to be confined to a certain depth range and
601 thereby constrains its base (e.g., Desissa *et al.*, 2013). Furthermore, if the resistivity of a conductor
602 can be inferred, its conductance (i.e. a product of thickness and conductivity) can be used to
603 determine its thickness (e.g., Comeau *et al.*, 2016).

604

605 ***Observations***

606 EM induction surveys have been conducted on most major sub-aerial volcanoes and magmatic
607 systems; only a few will be mentioned here to illustrate the type information on magma plumbing
608 systems that has been obtained. MT data have been used to image several low resistivity features in
609 the central Andes, particularly beneath the uplifting (10–15 mm/yr) Volcán Uturuncu, Bolivia (Fig.
610 9a) (Comeau *et al.*, 2015; Comeau *et al.*, 2016). The deepest of these bodies has resistivities of <3
611 Ω m, has a top contact at ~15–20 km depth (i.e. it is shallowest beneath Uturuncu), likely has a
612 thickness of >6 km, and extends E-W for ~170 km (Fig. 9) (Comeau *et al.*, 2015; Comeau *et al.*,
613 2016). This large-scale structure is interpreted to be the Altiplano-Puna magma body (APMB),
614 which has been identified in other geophysical datasets (e.g., Fig. 7a) (e.g., gravimetry, del Potro *et*
615 *al.*, 2013), with its low resistivity attributed to the presence of at least 20% andesitic melt and/or
616 magma. Extending from the top of the APMB towards the surface are several vertical, narrow (<10
617 km wide), low resistivity ($<10 \Omega$ m) zones that coincide with areas of seismicity and negative
618 gravity anomalies (Fig. 9). These zones likely reflect a network of dykes and upper crustal magma
619 reservoirs (Jay *et al.*, 2012; del Potro *et al.*, 2013; Comeau *et al.*, 2015; Comeau *et al.*, 2016).

620 Monitoring of magmatic systems can also be undertaken by both time-lapse and continuous
621 EM measurement. For example, MT data collected immediately after the 1977–1978 eruption at
622 Usu volcano, Japan revealed a conductive zone ($<100 \Omega$ m) beneath the summit that probably
623 corresponded to intruded magma. By 2000, MT data revealed that this conductive body had become
624 resistive (500–1000 Ω m) as the intrusion cooled, from 800°C to 50°C, and crystallised

625 (Matsushima *et al.*, 2001). Continuous MT monitoring of Sakurajima volcano, Japan between May
626 2008 and July 2009 revealed temporal changes in resistivity of $\pm 20\%$, some of which correlated to
627 periods of surface deformation and were inferred to reflect mixing between groundwater and
628 volatiles exsolved from an underlying magma body (Aizawa *et al.*, 2011). Continuous MT
629 monitoring at La Fournaise, Réunion Island recorded apparent resistivity decreases associated with
630 the large 1998 eruption, which were attributed to the injection of a N-S striking dyke (Wawrzyniak
631 *et al.*, 2017).

632 Several EM studies have focussed on magma plumbing systems at divergent margins,
633 including mid-ocean ridges and continental rifts. For example, at the fast-spreading East Pacific
634 Rise, a ~ 10 km wide, sub-vertical conductor, slightly displaced from the ridge axis and connected to
635 a deep, broad conductive zone was interpreted as a channel efficiently transporting melt to the base
636 of the crust (Baba *et al.*, 2006; Key *et al.*, 2013). Imaging of a crustal conductor for the first time
637 beneath a slow-spreading ridge, i.e. the Reykjanes ridge in the Atlantic Ocean, suggests that magma
638 injection into crustal reservoirs is intermittent but rapid (MacGregor *et al.*, 1998; Heinson *et al.*,
639 2000). Conversely, slow-spreading continental rifting in the Dabbahu magma segment, Afar,
640 Ethiopia appears to be underlain by a large conductor, either at the top of the mantle or straddling
641 the Moho, containing more melt (>300 km³) than is intruded into the magma plumbing system
642 during a typical rifting episode (Desissa *et al.*, 2013). The volume of this large conductor implies it
643 is a long-lived feature that could source magmatic activity for tens of thousands of years (Desissa *et*
644 *al.*, 2013).

645

646 ***Implications and integration***

647 It is clear from MT studies of the APMB that other geophysical techniques aid and/or corroborate
648 data interpretation (Fig. 9) (e.g., Comeau *et al.*, 2015; Comeau *et al.*, 2016). Over the last two
649 decades, numerous geophysical studies have been applied to examine magma and melt distribution
650 beneath various portions of the East African Rift, providing an excellent opportunity to test how

651 different techniques and data can be integrated. For example, extensive zones of melt beneath the
652 Afar region in Ethiopia inferred from MT data by Desissa *et al.*, (2013) is supported by: (i) the
653 occurrence of coincident, low P-wave velocity (down to 7.2 km s^{-1}) zones identified using from
654 analysis of seismic Pn waves that propagate along the Moho (Stork *et al.*, 2013); (ii) surface wave
655 studies that reveal lower crustal areas in magmatic domains with low S-wave velocities ($\sim 3.2 \text{ km s}^{-1}$) (Guidarelli *et al.*, 2011); and (iii) high anisotropic V_p/V_s ratios and low amplitude receiver
656 functions, which are indicative of the presence of melt (Hammond *et al.*, 2011; Hammond, 2014).
657 Similarly, crustal conductors along the northern flanks of the Main Ethiopian Rift, interpreted to
658 represent melt/magma (Whaler & Hautot, 2006; Samrock *et al.*, 2015; Hübert *et al.*, 2018), coincide
659 with locations where receiver functions either have amplitudes too low to interpret or indicate high
660 V_p/V_s values (Dugda *et al.*, 2005; Stuart *et al.*, 2006). Electrical anisotropy can be inferred directly
661 from MT data consistent with a two-dimensional subsurface resistivity distribution (Padilha *et al.*,
662 2006; Hamilton *et al.*, 2006). Large amounts of electrical anisotropy were found in the lower crust
663 beneath Quaternary magmatic segments in Afar, Ethiopia, where there is also significant crustal
664 seismic anisotropy (see Fig. 11 of Ebinger *et al.*, 2017); oriented melt-filled pockets are the
665 probable cause of both.

667 Although EM methods can image subsurface conductors that are interpreted to represent
668 magma bodies or zones of partial melt (i.e. crystal mushes), additional information is required to
669 determine their composition, volume, and/or melt fraction. However, there are several challenges in
670 inverting measured bulk resistivities to recover this information. Two-phase mixing laws predict
671 bulk resistivity is primarily a function of melt resistivity and geometry in the rock matrix when the
672 fluid phase has low resistivity, as in the case of partial melt. Well-connected melt gives a lower bulk
673 resistivity than isolated melt pockets, for the same melt fraction and resistivity (e.g. Hashin &
674 Shtrikman, 1963; Roberts & Tyburczy, 1999; Schmeling, 1986). Whilst resistivities of basaltic and
675 rhyolitic melts have been measured in laboratory experiments (e.g., Laumonier *et al.*, 2015; Guo *et al.*,
676 2016), they are strongly dependent on temperature, pressure, silica, sodium and water content,

677 making extrapolation uncertain. The web-based SIGMELTS tool can, however, be used to predict
678 melt and bulk resistivities for a wide range of compositions and conditions (Pommier & Le Trong,
679 2011). Importantly, petrological and geochemical characterisation of eruptive products can help
680 inform interpretations of associated, subsurface conductors but it is difficult to ascertain either
681 whether their composition reflects the current magma/melt present in the plumbing system or
682 whether melt pockets are interconnected. These large uncertainties in melt resistivity and the
683 requirement to make assumptions about its geometry make direct inference of melt fraction
684 difficult. Nonetheless, information from laboratory studies, petrology, and geochemistry aids
685 interpreting resistivity anomalies in magmatic regions (see review by Pommier, 2014).

686

687 **2.6. Imaging ancient magma plumbing systems in seismic reflection data**

688 *Techniques*

689 Over the last two decades, major advances have been made in imaging deep crustal melt beneath
690 active volcanic terrains using P- and S-wave tomographic data (e.g., Yellowstone, Husen *et al.*,
691 2004; Mt. St. Helens, Lees, 2007; Hawaii, Okubo *et al.*, 1997). These data image deep (>7 km),
692 often laterally extensive (up to 20 km), sill-like magma reservoirs (e.g., Paulatto *et al.*, 2012).
693 However, like many geophysical and geodetic techniques applied to study active magma plumbing
694 systems, these data typically lack the spatial resolution to resolve the detailed geometry of pathways
695 transporting magma to the Earth's surface. Active source seismic reflection data, which have a
696 spatial resolution of metres-to-decametres down to depths of ~5 km, can provide unprecedented
697 images of and insights into the geometry and dynamics of shallow-level, crystallised, magma
698 plumbing systems (e.g., Fig. 10) (e.g., Planke *et al.*, 2000; Smallwood & Maresh, 2002; Thomson &
699 Hutton, 2004; Cartwright & Hansen, 2006; Jackson *et al.*, 2013; Magee *et al.*, 2016; Schofield *et*
700 *al.*, 2017). Whilst seismic reflection data are traditionally used to find and assist in the production of
701 hydrocarbons in sedimentary basins (Cartwright & Huuse, 2005), we here discuss and support its
702 application to volcanological problems.

703 Acquiring active source seismic reflection data involves firing acoustic energy (i.e. seismic
704 waves) into the subsurface and measuring the surface arrival times (i.e. the travel-time) of reflected
705 energy. Processing of these arrival time data allows reconstruction of the location and geometry of
706 the geological interfaces from which acoustic energy was reflected. Mafic intrusive igneous rocks
707 are generally well-imaged in seismic reflection data because they typically have greater densities
708 ($>2.5 \text{ g/cm}^3$) and acoustic velocities (i.e. $>4000 \text{ m/s}$) than encasing sedimentary strata; these
709 differences result in a high acoustic impedance contrast, causing more seismic energy to be
710 reflected back to the surface compared to low acoustic impedance boundaries (Smallwood &
711 Maresh, 2002; Brown, 2004). In contrast, silicic igneous rocks have similar acoustic properties to
712 encasing sedimentary strata, meaning that felsic intrusions are rarely imaged in seismic reflection
713 data (Mark *et al.*, 2017; Rabbel *et al.*, 2018). Furthermore, because reflection seismology relies on
714 the return of acoustic energy to the surface, seismic reflection data favourably image mafic, sub-
715 horizontal-to-moderately inclined intrusions (e.g., sills, inclined sheets, and laccoliths; Smallwood
716 & Maresh, 2002; Jackson *et al.*, 2013; Magee *et al.*, 2016). Sub-vertical dykes reflect only a limited
717 amount of acoustic energy back to the surface and are thus typically poorly imaged in seismic
718 reflection data (e.g., Smallwood & Maresh, 2002; Planke *et al.*, 2005; Thomson, 2007; Wall *et al.*,
719 2010; Eide *et al.*, 2017a; Phillips *et al.*, 2017).

720

721 ***Observations***

722 Sills and inclined sheets are commonly observed in seismic reflection data as laterally
723 discontinuous, high-amplitude reflections, which may cross-cut the host rock strata (Fig. 10) (e.g.,
724 Symonds *et al.*, 1998, Smallwood & Maresh, 2002; Planke *et al.*, 2005, Magee *et al.*, 2015). Many
725 of the sills and inclined sheets imaged in seismic reflection data are, however, expressed as tuned
726 reflection packages, whereby discrete reflections from the top and base contacts interfere on their
727 return to the surface and cannot be distinguished (e.g., Figs 10 and 11a) (e.g., Smallwood &
728 Maresh, 2002; Peron-Pinvidic *et al.*, 2010; Magee *et al.*, 2015; Eide *et al.*, 2017a; Rabbel *et al.*,

729 2018). It is therefore difficult to assess either intrusion thicknesses, or to detect whether imaged sills
730 are composite bodies made of numerous, stacked, thin sheets. Either way, subtle vertical offsets and
731 corresponding amplitude variations of sill reflections can often be mapped, defining linear
732 structures that radiate out from either the central, deepest portions of sills or areas where underlying
733 intrusions intersect the sill (e.g., Schofield *et al.*, 2012a; Magee *et al.*, 2014; Magee *et al.*, 2016).
734 These structures are interpreted to relate to magma flow indicators such as intrusive steps, broken
735 bridges, and magma fingers (e.g., Schofield *et al.*, 2010; Schofield *et al.*, 2012b; Magee *et al.*,
736 2018).

737 A recurring observation from seismic reflection-based studies of extinct and buried intrusive
738 systems is that complexes of interconnected sills and inclined sheets, which may cover $>3 \times 10^6$
739 km², can dominate magma plumbing systems (e.g., Fig. 10b) (e.g., Svensen *et al.*, 2012, Magee *et*
740 *al.*, 2016). Importantly, where buried volcanic edifices are imaged in seismic reflection data, they
741 rarely appear to be underlain by ‘magma chambers’ (i.e. a spheroidal or ellipsoidal body of now-
742 crystallised magma). Instead, these imaged volcanoes commonly appear laterally offset from
743 genetically related sills and/or laccoliths that are inferred to represent their feeder reservoirs (e.g.,
744 Fig. 10b) (Magee *et al.*, 2013a; McLean *et al.*, 2017). The geometry, location, and connectivity of
745 these intrusions, which can represent magma storage sites and conduits to the surface, are often
746 heavily influenced by both the host rock structure and lithology (see review by Magee *et al.*, 2016).
747 For example, magma may flow along pronounced discontinuities (e.g., bedding) or within specific
748 stratigraphic units (e.g., coal) for considerable distances, occasionally climbing to higher
749 stratigraphic levels by instigating deformation of the host rock or by exploiting pre-existing faults
750 (e.g., Jackson *et al.*, 2013; Magee *et al.*, 2016; Schofield *et al.*, 2017; Eide *et al.*, 2017b). It is clear
751 from seismic reflection data that shallow-level tabular intrusions are commonly accommodated by
752 roof uplift to form a flat-topped or dome-shaped forced fold (e.g., Figs 11a and b) (e.g., Trude *et al.*,
753 2003; Hansen & Cartwright, 2006; Jackson *et al.*, 2013; Magee *et al.*, 2013b). Moreover, if the age
754 of reflections onlapping onto these intrusion-induced forced folds can be ascertained, the timing and

755 to some extent the duration of magmatic activity can be determined (e.g., Trude *et al.*, 2003;
756 Hansen & Cartwright, 2006; Magee *et al.*, 2014; Reeves *et al.*, 2018). Although most seismic-based
757 studies examine intrusions within sedimentary basins, saucer-shaped sills and laterally extensive
758 sill-complexes emplaced into crystalline basement rock are also imaged (e.g., Ivanic *et al.*, 2013;
759 McBride *et al.*, 2018). Lastly, seismic reflection data can also be used to image the internal structure
760 of layered ultramafic-mafic intrusions (e.g., the Bushveld Layered Intrusion, Malehmir *et al.*, 2012)
761 and, in some instances, identify dykes (e.g., Fig. 11c) (e.g., Wall *et al.*, 2010; Abdelmalak *et al.*,
762 2015; Bosworth *et al.*, 2015; Phillips *et al.*, 2017).

763

764 ***Implications and integration***

765 Despite being limited in terms of their spatial resolution (typically a few tens of metres) and ability
766 to image steeply dipping features (i.e. dykes), they provide unprecedented snapshots into the final
767 3D structure of magma plumbing systems. Beyond quantifying the structure and connectivity of
768 magma plumbing systems, seismic-based studies have shown that: (i) magma flow patterns mapped
769 across entire sill-complexes indicate they can transport melt from source to surface over great
770 lateral (>100's km) and vertical distances (10's km), potentially without significant input from
771 dykes (Fig. 10a) (e.g., Thomson & Hutton, 2004; Cartwright & Hansen, 2006; Magee *et al.*, 2014;
772 Magee *et al.*, 2016; Schofield *et al.*, 2017); and (ii) a variety of elastic and inelastic mechanisms can
773 accommodate host rock deformation during magma emplacement, meaning that the location and
774 size of ground deformation does not necessarily equal that of the forcing intrusion (e.g., Jackson *et*
775 *al.*, 2013, Magee *et al.*, 2013b). Importantly, observations from seismic reflection data highlight that
776 the lateral dimension should be considered when modelling the transit of magma in the crust, posing
777 problems for the widely held and simple assumption that magma simply travels vertically from melt
778 source to eruption site.

779 Seismic-based studies have also shown that direct comparison to active deformation
780 structures can be informative. For example, through comparing mapped lava flows and structures

781 associated with the Alu dome to similar features observed in seismic reflection data (see section
782 2.6), Magee *et al.*, (2017) concluded that the shallow-level sill likely has a saucer-shaped, as
783 opposed to the sill-like tabular morphology inferred from an episode of deformation measured using
784 InSAR (Figs 3c and d). Despite its benefits, it is important to remember that seismic reflection data
785 typically reveal only the final geometry of the magma plumbing system. There thus remains a
786 challenge in using these data to understand areas where deformation captures potentially transient,
787 active processes, rather than structures resulting from (multiple) periods of intrusion and cooling
788 (Reeves *et al.*, 2018). One potential and exciting way forward is the development of Virtual
789 Reflection Seismic Profiling, by which microseismicity at active volcanoes may be used to image
790 magma reservoirs and subsurface structure in 4D (Kim *et al.*, 2017). Although challenges exist in
791 dataset integration, the imaging power afforded by modern seismic reflection data thus presents a
792 unique opportunity to further unite field-, petrological-, geochemical-, and other geophysical-based
793 analyses within more realistic structural frameworks (e.g., Figs 3, 11a and b). In our view, however,
794 seismic reflection data are under-utilized in igneous research, remaining an unfamiliar technique to
795 many Earth Scientists in the volcanic and magmatic community.

796

797 **2.7. Rock magnetism**

798 *Technique*

799 Whilst seismic reflection data provide unique 3D images of ancient magma plumbing systems,
800 which can be used to infer magma flow patterns across entire intrusion networks, we commonly
801 lack sufficient data (e.g., boreholes) to test seismic-based hypotheses. It is therefore critical to
802 compare seismic interpretations to field analogues where magma flow patterns, emplacement
803 mechanics, and intrusion evolution can be investigated via other techniques. In this section, we
804 examine how rock magnetic analyses can be used to systematically study magnetic mineralogy and
805 petrofabrics, thereby illuminating the structure and history of igneous intrusions.

806 There are two principal types of rock magnetic study; magnetic remanence and magnetic
807 susceptibility, where the total magnetisation (M) of a rock is the sum of the magnetic remanence
808 (M_{rem}) and the induced magnetisation (M_{ind}), which is a product of the susceptibility (K) and
809 applied field strength (H) (Dunlop & Özdemir, 2001). Remanence carries a geological record of the
810 various magnetisations acquired over time and is central to palaeomagnetic studies. However, we
811 focus on magnetic fabric analysis, which relies on measurements of the anisotropy of magnetic
812 susceptibility (AMS). The AMS signal of a rock carries information from all constituent grains.
813 Although mineral phases that have a paramagnetic behaviour (i.e. they are weakly attracted to
814 externally applied magnetic fields) volumetrically dominate most igneous rocks (e.g., olivine,
815 clinopyroxene, biotite), ferromagnetic mineral phases (e.g., titanomagnetite) are highly susceptible
816 to magnetization and therefore tend to dominate K (e.g., Dunlop & Özdemir, 2001; Biedermann *et*
817 *al.*, 2014). Magnetic fabrics therefore typically reflect the preferential orientation of
818 crystallographic axes (i.e. crystalline anisotropy), the shape-preferred orientation of individual
819 crystals (i.e. shape anisotropy), and/or the alignment of closely spaced crystals (i.e. distribution
820 anisotropy) belonging to Fe-bearing silicate and oxide phases (e.g., Voight & Kinoshita, 1907;
821 Graham, 1954; Hrouda, 1982; Tarling & Hrouda, 1993; Dunlop & Özdemir, 2001). The principal
822 axes of the magnetic fabrics measured by AMS can thus be related to the orientation, shape, and
823 distribution of individual grains (i.e. the petrofabric) (e.g., Fig. 12a).

824 Regardless of whether mineral phases crystallise early or late, whereby their orientation and
825 distribution typically mimics the earlier silicate framework, it is expected that the initial petrofabric
826 developed in intrusive rocks will likely be sensitive to alignment of crystals during primary magma
827 flow. However, it is also critical to recognise that later magmatic processes (e.g., convection and
828 melt extraction) and syn- or post-emplacement tectonic deformation can modify or overprint
829 primary magma flow fabrics during intrusion, solidification (i.e. mush development), or sub-solidus
830 conditions (e.g., Borradaile & Henry, 1997; Bouchez, 1997; O'Driscoll *et al.*, 2015; Kavanagh *et*
831 *al.*, 2018). Whilst anisotropy of magnetic susceptibility (AMS) can thus rapidly and accurately

832 detect weak or subtle mineral alignments within igneous intrusions, which may be attributable to
833 magmatic and/or tectonic processes, evaluating the origin and evolution of petrofabric development
834 requires additional information (e.g., Borradaile & Henry, 1997; Bouchez, 1997). For example,
835 shape-preferred orientation analyses and comparison to visible flow indicators (e.g., intrusive steps
836 and bridge structures) allow magma flow axes and directions that have been inferred from magnetic
837 fabrics to be verified (e.g., Launeau & Cruden, 1998; Callot *et al.*, 2001; Magee *et al.*, 2012a). For a
838 useful précis of AMS-related magnetic theory in igneous rocks, the reader is referred to early works
839 by Balsey & Buddington (1960) and Khan (1962), and more recent summaries provided by Martín-
840 Hernández *et al.*, (2004), O'Driscoll *et al.*, (2008), and O'Driscoll *et al.*, (2015).

841 The principle behind AMS relies on the measurement of the bulk susceptibility (K_m) of a
842 single sample in different orientations to determine the susceptibility anisotropy tensor, which
843 relates the induced magnetisation (M_{ind}) to the applied field (H) in three dimensions (Tarling &
844 Hrouda, 1993). The orientation and magnitude of the eigenvectors and eigenvalues of this tensor
845 define an ellipsoid with three principal axes; the long axis of the ellipsoid, K_1 , defines the magnetic
846 lineation and the short axis, K_3 , defines the normal (i.e. the pole) to the magnetic foliation plane
847 (K_1 – K_2 ; Fig. 12a) (Stacy *et al.*, 1960; Khan, 1962; Tarling & Hrouda, 1993). In order to interpret
848 magnetic fabrics, it is important to determine the mineralogy of the phases carrying the magnetic
849 signal because the composition, grainsize, and distribution of magnetically dominant minerals (e.g.,
850 titanomagnetite) can control fabric orientation (e.g., Hargreaves *et al.*, 1991; Stephenson, 1994;
851 Dunlop & Özdemir, 2001). In addition to primary crystallographic and textural controls on
852 magnetic fabrics, subsequent oxidation of remaining melt and secondary hydrothermal alteration
853 can affect the magnetic mineralogy and, thereby, the AMS signal (e.g., Trindade *et al.*, 2001;
854 Stevenson *et al.*, 2007a). A variety of rock magnetic experiments are thus required to determine the
855 magnetic mineralogy. The most widely used method involves measuring susceptibility, and thereby
856 behaviour of magnetic materials, at varying temperatures ranging from -200°C to 700°C (i.e.
857 thermomagnetic analysis *sensu* Orlický, 1990; Hrouda *et al.*, 1997). For example, paramagnetic

858 materials (e.g., biotite) follow the Curie-Weiss law, whereby their susceptibility drops
859 hyperbolically with increasing temperature. In contrast, the thermomagnetic curve of ferromagnetic
860 materials (e.g., titanomagnetite) displays little change in susceptibility with temperature, apart from
861 when characteristic crystallographic transitions occur (e.g., the Curie point for pure magnetite at
862 ~580°C, Petrovský & Kapička, 2006) temperature. To determine the grain size of ferromagnetic
863 fraction in the magnetic susceptibility signal, the hysteretic property of the magnetisation is
864 important (Dunlop, 2002). Other rock magnetic experiments (e.g., anisotropy of anhysteretic
865 remanent magnetism (AARM) can be conducted to further isolate the relative importance of
866 different paramagnetic and ferromagnetic phases (e.g., McCabe *et al.*, 1985; Richter & van der
867 Pluijm, 1994; Kelso *et al.*, 2002).

868

869 ***Observations***

870 Having established the magnetic mineralogy, AMS fabrics can be interpreted. Even in weakly
871 anisotropic igneous rocks (i.e. visually isotropic), particularly sheet intrusions, it is now accepted
872 that the magnetic lineation and foliation can provide information on magma migration (e.g., flow
873 direction) or regional and local strain (e.g., Hrouda, 1982; Knight & Walker, 1988; Rochette *et al.*,
874 1992; Bouchez, 1997; Tauxe *et al.*, 1998; Callot *et al.*, 2001; Féménias *et al.*, 2004; Magee *et al.*,
875 2012a). For example, comparisons to other indicators of magma flow (e.g., intrusive steps and
876 visible mineral alignments) in sheet intrusions have shown that magnetic lineations commonly
877 parallel the magma flow (e.g., Knight & Walker, 1988; Cruden & Launeau, 1994; Callot *et al.*,
878 2001; Magee *et al.*, 2012a), whilst imbrication of elongate crystals induced by simple shear at
879 intrusion margins define the sense of magma flow (Fig. 12b) (e.g., Knight & Walker, 1988;
880 Hargraves *et al.*, 1991; Stephenson, 1994; Geoffroy *et al.*, 2002; Féménias *et al.*, 2004).

881 Alternatively, contact-parallel magnetic fabrics generated during the formation and inflation of
882 magma lobes can be used to determine flow and emplacement dynamics, even if other evidence for
883 the presence of magma lobes is lacking (e.g., Fig. 12c) (Cruden *et al.*, 1999; Stevenson *et al.*,

884 2007a; Magee *et al.*, 2012b). Identifying changes in fabric orientation within or between individual
885 sheet intrusions is also important because these variations suggest that deformation, imparted by
886 either the emplacement of adjacent magma bodies or tectonic processes, did not significantly
887 modify magma emplacement fabrics (e.g., Clemente *et al.*, 2007).

888 Post solidification textural modification and the possibility of overlap in tectonic and
889 magmatic strain fields during protracted emplacement is a particular complication when studying
890 granitoid and gabbroic plutons (e.g., Mamtani *et al.*, 2013; O'Driscoll *et al.*, 2015; Cheadle *et al.*,
891 2017). In fact, most early studies of granitoid emplacement using AMS, in conjunction with many
892 other structural analysis tools, concluded that tectonic strain was the main source of subtle fabrics
893 (e.g., Brun *et al.*, 1990; Bouchez, 1997; de Saint-Blanquat & Tikoff 1997; Neves *et al.*, 2003;
894 Mamtani *et al.*, 2005). Although primary magma flow fabrics in granitic and gabbroic plutons may
895 thus be overprinted, the magnetic fabrics characterised by AMS can still provide fundamental
896 insights into emplacement mechanics (e.g., Stevenson *et al.*, 2007a; Petronis *et al.*, 2012) and
897 magma/mush evolution (e.g., formation of layering; O'Driscoll *et al.*, 2015).

898

899 ***Implications and integration***

900 Overall, AMS has provided vital magma flow and evolution information that has helped to
901 understand mafic and silicic magma plumbing systems (e.g., Knight & Walker, 1988; Ernst &
902 Baragar, 1992; Glen *et al.*, 1997; Aubourg *et al.*, 2008; Petronis *et al.*, 2013; Petronis *et al.*, 2015).
903 Critical insights emanating from these AMS studies have revealed that: (i) flow trajectories
904 predicted by classic emplacement models (e.g., for ring dykes and cone sheets) are not always
905 consistent with measured AMS fabrics and supporting data, which thereby call into question the
906 application of such models (e.g., Stevenson *et al.*, 2007b; Magee *et al.*, 2012a); (ii) lateral magma
907 flow is recorded in many shallow, planar intrusions associated with volcanic magma plumbing
908 systems (e.g., Ernst & Baragar, 1992; Cruden & Laneau, 1994; Cruden *et al.*, 1999; Herrero-
909 Bervera *et al.*, 2001; Magee *et al.*, 2012a; Petronis *et al.*, 2013; Petronis *et al.*, 2015); and (iii)

910 plutons, particularly those with a granitic composition, commonly consist of incrementally
911 emplaced magma pulses that often develop lobate geometries (e.g., Fig. 12c) (e.g., Stevenson et al.,
912 2007a). Analysing AMS fabrics from layered mafic-ultramafic intrusions can also provide evidence
913 for magma reservoir processes, including crystal settling, or post-cumulus modification of crystal
914 mushes (O’Driscoll *et al.*, 2008; O’Driscoll *et al.*, 2015). Importantly, AMS and related analyses
915 provide robust, testable, and repeatable methods to constrain subtle shape and crystallographic
916 orientations of crystals in igneous rocks. Rock magnetic instrumentation technology continues to
917 advance with better automation of measurement protocols, sensitivity of measurements, and a
918 greater ability to unravel contributors to the AMS signal. The direction and scope of these
919 developments are improving the holistic integration of AMS with other structural, microstructural,
920 geophysical, petrological and geochemical techniques, promising to advance our understanding of
921 magmatism and crustal evolution.

922

923 **3. Future advances**

924 Our understanding of magma plumbing system structure and evolution has been significantly
925 enhanced by the geophysical techniques described above. We have demonstrated that there is scope
926 for advancement within individual methodologies and through the integration of different
927 techniques, particularly involving the synthesis of geophysical, petrological, and geochemical data.
928 In this section, we discuss two new techniques that will potentially revolutionize our understanding
929 of magma plumbing systems. We also briefly discuss how integration of geophysical data with
930 numerical modelling can enhance our knowledge of reservoir construction and evolution.

931

932 **3.1. Full-Waveform Inversion**

933 *Technique*

934 We have demonstrated that seismic reflection data can provide unique insight into the 3D structure
935 of magma plumbing systems (e.g., see review by Magee *et al.*, 2016). In addition to using seismic

936 reflection data to image the subsurface, we can also invert the measured travel-times of reflected
937 acoustic energy to model subsurface P-wave velocities. Full-waveform inversion (FWI) is a rapidly
938 developing technology using active source seismic data to generate models that reproduce both the
939 travel-times and full waveform of the arriving wavefield, thereby matching observed seismic data
940 (Tarantola, 1984). Because FWI considers the full wavefield, as opposed to conventional techniques
941 that only model travel-times, it is a technique capable of recovering high-resolution models of
942 subsurface P-wave velocities and other physical properties (Warner *et al.*, 2013; Routh *et al.*, 2017).
943 The FWI technique begins with a best-guess starting velocity model for the subsurface geology,
944 which is then iteratively updated using a local linearized inversion until the observed seismic data is
945 matched (Virieux & Operto, 2009). FWI is much more computationally expensive than travel-time
946 tomography, as a full-physics implementation of the wave equation is required to generate the
947 predicted seismic data at all energy source and receiver locations for each iteration (Routh *et al.*,
948 2017). FWI, however, has the advantage of being able to resolve much finer-scale structure than
949 conventional techniques.

950

951 ***Observations***

952 To date, 3D FWI has principally been applied within the petroleum sector to obtain high-resolution
953 velocity models that can be used to improve depth-migrated (i.e. travel-time is converted to depth in
954 metres) reflection images of petroleum reservoirs and their overburden (Sirgue *et al.*, 2010; Vigh *et al.*,
955 2010; Warner *et al.*, 2013; Kapoor *et al.*, 2013; Routh *et al.*, 2017). FWI can also produce
956 interpretable, quantitative models of physical properties of rocks in the subsurface that can be
957 related directly to compaction, permeability, and overpressure as measured in subsurface boreholes
958 (Lazaratos *et al.*, 2011; Mancini *et al.*, 2015). Of relevance here is that mafic intrusions, which
959 appear as high-amplitude reflections in seismic reflection data (Figs 10 and 11a), are recovered as
960 high-velocity features in FWI velocity models (Fig. 13) (Mancini *et al.*, 2015; Kalincheva *et al.*,
961 2017). For example, successful application of 3D FWI to a marine ocean bottom seismometer

962 dataset acquired across the Endeavour segment of the Juan de Fuca Ridge led to generation of a
963 velocity model that had a resolution up to four times greater than travel-time tomography (Morgan
964 *et al.*, 2016). Within this new, high-resolution velocity model, several velocity anomalies were
965 identified and interpreted to indicate localized magma recharge of the axial reservoir, induced
966 seismogenic cracking, and increased permeability (Arnoux *et al.*, 2017).

967

968 ***Implications and integration***

969 Active magma plumbing systems comprise a complex network of interconnected conduits and
970 reservoirs with variable geometries and sizes, which likely contain magmatic vapour-rich, liquid-
971 rich, and mush-zones (Christopher *et al.*, 2015). These intrusions will all be associated with reduced
972 P-wave velocities, which could be resolved in high-resolution, 3D FWI datasets as supported by
973 successes in the fine-scale imaging of: (i) low-velocity gas clouds (Warner *et al.*, 2013); (ii) axial
974 reservoirs at an oceanic spreading centre (Arnoux *et al.*, 2017); (iii) relatively narrow, low-velocity
975 fault zones within an antiform (Morgan *et al.*, 2013); and (iv) a subduction zone using 2D FWI
976 (Kamei *et al.*, 2012). A suite of synthetic tests has been performed to investigate whether 3D FWI
977 could be applied to better understand magma plumbing systems (Morgan *et al.*, 2013). These tests
978 indicate that it is possible to recover high-resolution models of P-wave velocity beneath volcanoes,
979 which can then be used to better determine where magma/mush is stored beneath the surface. In
980 particular, these synthetic tests suggest that FWI could be used to: (i) distinguish between
981 continuous zones of mush and individual magma reservoirs; (ii) image sills and conduits of magma
982 and/or fluids that are a few 10s metres across (Fig. 13); and (iii) image the deeper (lower-crustal)
983 part of the magma system. We therefore consider that 3D FWI affords an unprecedented
984 opportunity to obtain high-resolution images of actual magma plumbing systems beneath active
985 volcanoes. To this end, the ongoing PROTEUS (Plumbing Reservoirs Of The Earth Under
986 Santorini) experiment was specifically designed to use 3D FWI to investigate the Santorini magma
987 plumbing system (Hooft *et al.*, 2017).

988

989 **3.2. Unmanned Aerial Vehicle photogrammetry**

990 *Technique*

991 Despite major advances in satellite-based remote sensing systems and aeromagnetic surveys, very
992 high-resolution (i.e., mm–cm scale ground sampling distance) imagery of dykes and other igneous
993 intrusions has been limited to low altitude aerial photography. This in turn has created a critical
994 scale gap in intrusion studies, which range from <1 mm at thin section scale to the metres to 100's
995 of metres scale provided by outcrop analysis, conventional remote sensing, and geophysical data.
996 Fortunately, the emerging capability of unmanned aerial vehicle (UAV) photogrammetry fills this
997 gap (e.g., Eisenbeiss, 2009; Westoby *et al.*, 2012; Bemis *et al.*, 2015; Eide *et al.*, 2017b). It is also
998 noteworthy that several studies have demonstrated that digital photogrammetry can deliver high
999 quality datasets with accuracies similar to more established laser scanning techniques (e.g., Leberl
1000 *et al.*, 2010; Hodgetts, 2013; Thiele *et al.*, 2015).

1001 The basic setup required to carry out UAV (or drone) photogrammetry is commercially
1002 available and relatively inexpensive, comprising a fixed wing or rotary wing UAV, a digital camera,
1003 and access to a suitable digital photogrammetry software package (e.g., Agisoft Photoscan Pro,
1004 Pix4Dmapper Pro, VisualSFM). UAV photogrammetry combines a simple and cost-effective
1005 method to acquire geospatially referenced, overlapping digital aerial images, from which structure-
1006 from-motion algorithms can generate spatial 3D datasets (Bemis *et al.*, 2014; Vollgger & Cruden,
1007 2016). Such an approach can be used for high spatial resolution mapping of all types of well-
1008 exposed igneous intrusions. The resulting data greatly enhance the effectiveness of traditional field
1009 mapping, particularly the characterisation of contact relationships and internal and external structure
1010 (e.g., fractures, fabrics, and phase distributions) of intrusive rocks, complementing AMS and
1011 petrological analyses.

1012

1013 *Observations*

1014 Aa photogrammetric workflow was applied to examine a swarm of 5 cm to 1 m wide Palaeogene
1015 dolerite and dacite dykes exposed on coastal outcrops at Bingie Bingie Point, SE Australia (Fig.
1016 14). The orthophotograph of the entire wave-cut platform shows the distribution of the Palaeogene
1017 dolerite and dacite dykes and their Devonian host rock lithologies, including a prominent
1018 moderately NE-dipping aplite dyke (Fig. 14a). Linear ENE-WSW linear terrain features pick out
1019 the traces of dyke-parallel joints (Fig. 14a). The Palaeogene dykes trend 063° parallel to a major set
1020 of joints in the country rock that likely formed contemporaneously with syn-dyking extension (Fig.
1021 14b). Subsidiary joint sets trend NNW-SSE, sub-perpendicular to the Palaeogene dykes, N-S and E-
1022 W (Fig. 14b). The Palaeogene dykes display considerable structural complexity such as bridge
1023 structures, intrusive steps and apophyses (Fig. 14c). Where present, the steps mostly occur where
1024 dykes cross country rock contacts (e.g., the aplite-tonalite contact in the NE; Fig. 14c).

1025

1026 *Implications and integration*

1027 Data such as the orthophotograph collected at Bingie Bingie Point indicate that high-resolution
1028 structural and lithological mapping and measurement can be carried out much more rapidly than by
1029 traditional survey methods (e.g., plane table or grid mapping). However, the use of conventional
1030 RGB cameras restricts the resulting image data to reflected visible light. Future applications will
1031 include the deployment of multispectral and hyperspectral sensors (infrared to short wave infrared
1032 to thermal infrared) as well as potential field geophysical or geodetic instruments (e.g., Sparks,
1033 2012). A further challenge for UAV applications in many countries concerns the regulatory
1034 framework around the use of drones for research. The global trend is moving to require non-
1035 recreational UAV operators to have remotely piloted aircraft licences and for the associated
1036 organisation to be certified for UAV operations. Innovations in sensor types and design, attachment
1037 of geophysical instruments, machine learning, and integration with complementary techniques such
1038 as AMS will open up new avenues for UAV applications in the study of magma plumbing systems.

1039

1040 **3.4. Numerical modelling of magma reservoir processes constrained by geophysical data**

1041 Geophysical imaging of both active and ancient magma plumbing systems is delivering new
1042 insights into the 3D geometry of reservoirs, the timing and rates of melt and magma transport, the
1043 pathways followed by magmas as they ascend through the crust, and typical stored melt fractions in
1044 mushes. These data can be used to constrain and calibrate numerical models of reservoir processes.
1045 Numerical models are used ubiquitously to understand and predict the behaviour of other
1046 subsurface crustal reservoirs, such as hydrocarbon reservoirs, groundwater resources, and targets for
1047 geological CO₂ storage (e.g., Chen *et al.*, 2003; Class *et al.*, 2009; Dean & Chen, 2011). However,
1048 there has been relatively little focus to date on developing numerical models for magma/mush
1049 reservoirs. Yet such models can integrate across different data sources and types, provide
1050 quantitative estimates of rates, volumes and timescales, and provide a framework for data
1051 interpretation. For example, numerical modelling of heat transfer within the plumbing system at
1052 Okmok Volcano in Alaska, which was informed by analytical models of geodetic data and
1053 estimated magma compositions of erupted material, allowed estimation of the role magma injection,
1054 crystallisation, and degassing processes had on volume changes over time (Caricchi *et al.*, 2014).
1055 Numerical thermal modelling has also helped interpret seismic data from the Soufrière Hills
1056 Volcano, Montserrat, suggesting higher melt fraction in the underlying magma reservoir than was
1057 inferred from seismic data alone (Paulatto *et al.*, 2012). More recent numerical models focus on
1058 crystal mushes, evaluating melt transport and reaction at low melt fractions, and these show that
1059 temperature and melt fraction in mushes can be decoupled; i.e. maximum temperature occurs close
1060 to the centre of the reservoir but maximum melt fraction occurs close to the top (Solano *et al.*,
1061 2014). This decoupling impacts how seismic velocities and electrical conductivities will be
1062 modified within the mush (Solano *et al.*, 2014). Other numerical models show the important role
1063 played by exsolution, crystallisation, and the viscoelastic response of the crust in driving magma
1064 mobilisation in and eruption from shallow reservoirs (e.g., Degruyter & Huber, 2014; Parmigiani *et*
1065 *al.*, 2016), as well as providing insights into the mixing mechanisms of melt and crystals in mushes

1066 (Bergantz *et al.*, 2015). However, most models to date have a lower dimensionality (zero dimension
1067 box models, or one/two dimensions) and capture only a small subset of the key physical and
1068 chemical processes that are likely to occur in crustal magma reservoirs or crystal mushes.
1069 Moreover, few studies have integrated modelling with geophysical data (cf. Gutierrez *et al.*, 2013).
1070 This is in marked contrast to the 3D modelling routinely undertaken of other crustal reservoirs (e.g.,
1071 hydrocarbon reservoirs), which is commonly integrated with and delimited by geophysical data.
1072 There is thus significant scope for improved, and integrated, numerical modelling of crustal magma
1073 reservoirs.

1074

1075 **4. Conclusions**

1076 Determining the structure of magma plumbing systems is critical to understanding where melt and
1077 magma is stored in the crust, which can influence the location of volcanic eruptions and economic
1078 ore deposits, providing an important framework for interpreting the physical and chemical evolution
1079 of magma from petrological and geochemical datasets. Geophysical techniques have revealed
1080 unique insights into the architecture of active and ancient magma plumbing systems, which when
1081 integrated with traditional structural, petrological and geochemical results has yielded exciting
1082 advances in our understanding of magmatic processes. However, divisions between communities
1083 applying these methodologies still exist, contributing to diverging views on the nature of magma
1084 plumbing systems. To help promote collaboration, we have reviewed a range of geophysical
1085 techniques and discussed how they could be integrated with structural, petrological and
1086 geochemical datasets to answer outstanding questions in the volcanological community. In
1087 particular, we demonstrate how a range geophysical techniques can be applied to track melt
1088 migration in near real-time, map entire intrusion networks in 3D, examine magma emplacement
1089 mechanics, and understand the evolution of crystal mushes. For example, Interferometric Synthetic
1090 Aperture Radar (InSAR) allows measurement of the development of active magmatic systems by
1091 successive intrusion, the vertical and lateral movements of magma, and the relationship between

1092 magma plumbing system dynamics and eruption. Seismicity beneath volcanoes can, when the
1093 magma interacts dynamically with the host rock, illuminate in high-resolution the time and spatial
1094 scales of the motion of magma and hydrothermal fluids. Seismic imaging of magma plumbing
1095 systems allows the spatial distribution of melt and magma to be determined whilst the inclusion of
1096 anisotropy within seismic techniques even allows sub-seismic wavelength features to be identified.
1097 Gravimetry can characterise the distribution and redistribution of mass (e.g., magma) in the
1098 subsurface over high spatial and temporal resolutions, helping to reveal the structure and
1099 composition of magma plumbing systems and the source(s) of volcano deformation.

1100 Electromagnetic methods, particularly magnetotellurics, can identify fluids within magmatic
1101 systems (e.g., melt, magma, and hydrothermal fluids). Seismic reflection data provide
1102 unprecedented 3D images of ancient magma plumbing systems and has revealed that laterally
1103 extensive, interconnected networks of sills and inclined sheets can play a pivotal role in transporting
1104 magma through the crust to eruption sites potentially located >100 km away from the melt source.
1105 Rock magnetics can provide fabric data pertaining to magma flow, deformation or crystallisation.

1106 All these methodologies discussed have provided unique insights into the structure of igneous
1107 intrusions and, through integration with petrological and geochemical datasets, are beginning to
1108 help unravel the entire evolution of magma plumbing systems. In addition to the ongoing
1109 application and advancement of these geophysical techniques, emerging methodologies look set to
1110 radically improve our understanding of magma plumbing systems. For example, full-waveform
1111 inversion can image and characterise physical properties across plumbing systems at an
1112 unprecedented resolution, whereas unmanned aerial vehicle photogrammetry provides a tool for
1113 high spatial resolution of outcrop scale intrusions that bridges the scale gap between seismic
1114 reflection data and traditional mapping of magma plumbing systems. The geophysical techniques
1115 discussed also provide critical constraints on input parameters for numerical modelling. Overall, we
1116 consider that the future of magma plumbing system studies will benefit greatly from the synthesis
1117 of geophysics and more traditional petrological and geochemical approaches.

1118

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1126 **6. References**

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1967 coda interferometry. *Journal of Geophysical Research-Solid Earth* **118**, 1067–1078.

1968

1969 **7. Figure captions**

1970

1971 Figure 1: Schematic of a vertically extensive, transcrustal magma plumbing system involving
1972 transient, interconnected, relatively low-volume tabular magma intrusions (e.g., dykes, sills, and
1973 laccoliths) within a crystal mush (based on Cashman *et al.*, 2017; Cruden *et al.*, 2018).

1974

1975 Figure 2: (A) Interferograms showing fringes caused by the pressurisation of a point source directly
1976 beneath a stratovolcano from both ascending and descending satellite lines of sight. Note that the
1977 centre of the fringes are slightly offset from the summit of the volcano (marked by a black triangle).
1978 (B) Typical fringe patterns for analytical deformation sources in an elastic half space from
1979 ascending satellite geometry: (i) Mogi source at 5 km depth; (ii) dyke extending between depths of
1980 3 and 9 km; (iii) rectangular sill; and (iv) a penny-shaped horizontal crack both at 5 km depth.

1981

1982 Figure 3: (A) Ascending line of sight (LOS) co-eruptive interferogram from the 2008 basalt lava
1983 extrusion between the Alu and Alu South domes and the Dalafilla stratovolcano (modified from
1984 Pagli *et al.*, 2012). (B) Inversion of uplift and subsidence patterns, recorded by InSAR during the
1985 2008 basalt lava eruption at the Alu dome in the Danakil Depression, suggested ground deformation
1986 could be attributed to a combination of: (i) deflation of a reservoir, modelled as a Mogi source, at
1987 ~4 km depth; (ii) inflation and deflation of a tabular sill at ~1 km depth; and (iii) opening of a dyke
1988 beneath the eruptive fissure (Figs 3A and B) (Pagli *et al.*, 2012). See Figure 3A for location. (C)
1989 Geological map showing that lava flows radiate out from Alu and originate from the periphery of
1990 the dome, which is cross-cut by an array of randomly oriented faults (modified from Magee *et al.*,
1991 2017). (D) Magee *et al.*, (2017) inferred Alu is underlain by a saucer-shaped sill plumbing system,
1992 based on field observations and comparison to seismic reflection data, not a tabular sill (Fig. 3B).

1993

1994 Figure 4: Example of integrating seismology and petrology to constrain time-scales of magma
1995 storage and recharge (from Saunders *et al.*, 2012). Calculated Fe-Mg diffusion time scales of
1996 orthopyroxene crystals compared to monitoring data for the same eruptive period for Mount St.
1997 Helens. (A) The seismic record of depth against time of the 1980–1986 eruption sequence. (B)
1998 Measured flux of SO₂ gas. (C) Calculated age of orthopyroxene rim growth binned by month for the
1999 entire population. The age recorded is the month in which the orthopyroxene rim growth was
2000 triggered by magmatic perturbation. The black line displays the running average (over five points,

2001 equivalent to the average calculated uncertainty in calculated time scales) of all the data. The peaks
2002 in the diffusion time series correspond to episodes of deep seismicity in 1980 and 1982 and to
2003 elevated SO₂ flux in 1980 and possibly 1982. (D) Running average of the orthopyroxene rim time
2004 scales, displaying reverse zonation (Mg-rich rims) in blue and normal zonation (Fe-rich rims) in
2005 green. There are reverse zonation peaks in the early 1980, probably due to rejuvenation of the
2006 magma system by hotter pulses, whereas Fe-rich rims are more dominant from 1982 on. Vertical
2007 dashed grey lines represent the volcanic eruptions.

2008

2009 Figure 5: Plot of melt inclusion saturation and earthquake hypocentre depths, which suggest magma
2010 storage occurred at 1–5 km depths, beneath the Dabbahu volcanic system in Afar, Ethiopia
2011 (modified from Field *et al.*, 2012). Melt inclusion data obtained from analyses of alkali feldspar,
2012 clinopyroxene, and olivine phenocrysts within Dabbahu lavas <8 Kyr (Field *et al.*, 2012).
2013 Earthquake data recorded during the 2005 dyke event (Ebinger *et al.*, 2008).

2014

2015 Figure 6: (A) P-wave (V_p) tomography beneath Montserrat (black outline), highlighting the location
2016 of fast and slow seismic velocity anomalies (i.e. >6% faster or slower than average) relative to the
2017 location of the Silver Hills (SH), Central Hills (CH), and Soufrière Hills (SHV) volcanoes
2018 (modified from Shalev *et al.*, 2010). The fast velocity anomalies, interpreted to represent solidified
2019 andesitic intrusions underlie the volcanoes (Shalev *et al.*, 2010).

2020

2021 Figure 7: Static and dynamic gravimetric investigations of two active silicic magmatic systems in
2022 the Andes: Uturuncu volcano (Bolivia; A, C, and E) and the Laguna del Maule volcanic field
2023 (Chile; B, D, and F). (A) 3D view of the isosurface corresponding to the -120 kg m³ density contrast
2024 beneath Uturuncu volcano, derived from Bouguer gravity data, interpreted to reflect a large (~750
2025 km³) plumbing system composed of a lower (<10 km) partially molten reservoir and upper,
2026 fractured and fluid-bearing solidified intrusions above sea level (after del Potro *et al.*, 2013). (B) 3D

2027 view of the -600 kg m^3 density contrast isosurface beneath the Laguna del Maule, which is
2028 interpreted to define a magma reservoir ($>50\%$ melt) within a larger region of a crystal mush
2029 system; the 2D planes show slices through the dataset (Miller *et al.*, 2017). Elevation above sea
2030 level (a.s.l.) shown. See Figure 7D for area of data coverage. (C) Map of the 55 km long, dynamic
2031 gravity network (white circles) installed to track changes in gravity over time and space at Uturuncu
2032 volcano between 2010 and 2013 (modified from Gottsmann *et al.*, 2017). (D) Spatio-temporal
2033 residual gravity changes at Laguna del Maule recorded from 2013–2014, after correcting for
2034 deformation effects (modified from Miller *et al.*, 2017). (E) Gravity and deformation data, recorded
2035 from Uturuncu from 2010–2013, plotted against the measured free-air gravity gradient (solid red
2036 line) and associated errors (broken red lines) (modified from Gottsmann *et al.*, 2017). The data
2037 follow the gradient and are indicative of a subsurface density change as a cause of the uplift,
2038 possibly reflecting the release of fluids from a large deep-seated magma reservoir (i.e. the
2039 Altiplano-Puna Magmatic Body; Chmielowski *et al.*, 1999) through the vertically extensive crystal
2040 mush system shown in (A) (Gottsmann *et al.*, 2017). (F) Plot of gravity against horizontal distance
2041 for the source centre at Laguna del Maule (modified from Miller *et al.*, 2017). The increase in
2042 gravity of up to $120 \mu\text{Gal}$ is explained by a hydrothermal fluid injection focused along a fault
2043 system, shown in (D), at 1.5–2 km depth as a result of a deeper seated magma injection, and is best
2044 modelled by a vertical rectangular prism source.

2045

2046 Figure 8: Gravity changes and deformation at the restless Long Valley caldera. (A) Map of the
2047 Long Valley caldera, California, USA, which hosts a resurgent dome (black outline), to highlight
2048 changes in residual gravity between 1982 and 1999 (modified from Tizzani *et al.*, 2009). (B) Plot of
2049 ground uplift and residual gravity changes with radial distance from the centre of the resurgent
2050 dome in (A) (modified from Tizzani *et al.*, 2009). The correlation between uplift and positive
2051 gravity residuals across the resurgent dome indicates ground deformation was instigated by
2052 intrusion of magma (Tizzani *et al.*, 2009).

2053

2054 Figure 9: (A) Map showing MT stations deployed around Volcán Uturuncu (U) and Volcán
2055 Quetena (Q), relative to areas of uplift and subsidence (modified from Comeau *et al.*, 2015). The
2056 white box shows area of modelled 3D MT data (Comeau *et al.*, 2015). (B) Regional 2D
2057 magnetotelluric line through the Altiplano-Puna magma body (APMB) highlighting the position of
2058 Volcán Uturuncu (modified from Comeau *et al.*, 2015). The APMB corresponds to a large,
2059 conductive (i.e. low-resistivity) body (Comeau *et al.*, 2015; Comeau *et al.*, 2016). Above the APMB
2060 are other areas of low-resistivity (e.g., C4) that are likely upper crustal magma reservoirs and dykes
2061 (Comeau *et al.*, 2016). C1–C7 and R1–R2 identify discrete zones of marked conductivity or
2062 resistivity, respectively (see Comeau *et al.*, 2015; Comeau *et al.*, 2016 for details). The white box
2063 shows area of modelled 3D MT data (Comeau *et al.*, 2015). See Figure 9A for location.

2064

2065 Figure 10: (A) Interpreted seismic section and geological map showing the distribution of and
2066 connectivity between sills within the Faroe-Shetland Basin (modified from Schofield *et al.*, 2017).
2067 Mapping of magma flow patterns within individual sills reveals that the sill-complex facilitates
2068 extensive vertical and lateral magma transport. Magma was fed into the sedimentary basin via
2069 basement-involved faults. TWT = two-way travel time. (B) Interpreted seismic section and
2070 geological map describing the spatial relationship between volcanoes/vents and sills, inferred to
2071 represent the magma plumbing system, emplaced at ~42 Ma (modified from Jackson *et al.*, 2013;
2072 Magee *et al.*, 2013a). Sills are laterally offset from the volcanoes/vents summits. No ‘magma
2073 chambers’ are observed in the seismic data, which images down to ~8 s TWT (i.e. ~>10 km)
2074 (Magee *et al.*, 2013a).

2075

2076 Figure 11: (A) Interpreted seismic section from the Exmouth Sub-basin offshore NW Australia,
2077 which images a saucer-shaped sill that is overlain by a forced fold and feeds a small vent from its
2078 inclined limb (modified from Magee *et al.*, 2013b). See Figure 11B for line location). (B) Time-

2079 structure map of the folded horizon (thick black line) in (A), highlighting fault traces and vent
2080 locations and thicknesses (modified from Magee *et al.*, 2013b). (C) Seismic section from the
2081 Farsund Basin, offshore southern Norway, which images part of a dyke-swarm that has been rotated
2082 by basin flexure post-emplacement (modified from Phillips *et al.*, 2017).

2083

2084 Figure 12: (A) At the sample scale, all magnetic grains create a magnetic fabric. (i) Dominantly
2085 prolate fabric, where K_2 and K_3 are least certain and form a girdle. Only the magnetic lineation (K_1)
2086 can be confidently determined. (ii) When $K_1 > K_2 > K_3$, both a foliation (K_1 – K_2) and a lineation (K_1)
2087 may be discerned, defining a triaxial fabric. (iii) When K_1 and K_2 are equally uncertain and form a
2088 girdle, K_3 is perpendicular to a foliation. (B) Schematic representation of how magma flow within a
2089 planar sheet intrusion can produce imbricated magnetic fabrics at its margins, the closure of which
2090 define the magma flow direction (after Féménias *et al.*, 2004). (C) AMS data and interpretations
2091 from part of the Trawenagh Bay Granite, NW Ireland (adapted from Stevenson *et al.*, 2007a). (i)
2092 AMS foliation traces are shown in blue and lineation traces in red. Lobes were defined in this
2093 intrusion based on foliations curving around a lineation axis. In some lobes, the magnetic lineation
2094 trend was parallel to this axis, whilst in others they tended to splay or converge down flow. (ii) 3D
2095 sketch showing the geometry of three of the lobes (numbered in part i).

2096

2097 Figure 13: (A) Starting model derived from smoothed, pre-stack, time-migrated (PSTM) stacking
2098 velocities. (B) Final 2D FWI-derived velocity model obtained using 10 km streamer data and
2099 inversion frequencies of between 2.5 and 24 Hz. (C) FWI velocity model overlain by the 2D pre-
2100 stack, depth-migrated (PSDM) section. Strong irregular reflections in the lower half of the section
2101 are from basaltic intrusions, which appear as high-velocity anomalies in the FWI velocity model.
2102 Both the FWI velocity model and the PSDM pick out a major unconformity, and show shallow
2103 channels in the upper parts of the section (redrawn from Kalincheva *et al.*, 2017).

2104

2105 Figure 14: (A) UAV orthophotograph of the wave cut platform at Bingie Point, NSW, Australia
2106 showing the distribution of Palaeogene dolerite (Dol) and dacite (Dac) dykes within Devonian
2107 tonalite (Ton), diorite (Di), and aplite (Ap) host rocks. (B) Circular histogram of joint sets measured
2108 in the Devonian rocks from the orthophotograph; the dominant (purple) set is parallel to and likely
2109 contemporaneous with the Palaeogene dykes. (C) Annotated close-up image highlighting dykes and
2110 structural features. The northern dacite dyke shows two broken bridge (BB) structures, whilst the
2111 central dolerite dyke displays prominent step structures (S). Narrow apophyses are also associated
2112 with the broken bridges and steps.

Figure 1

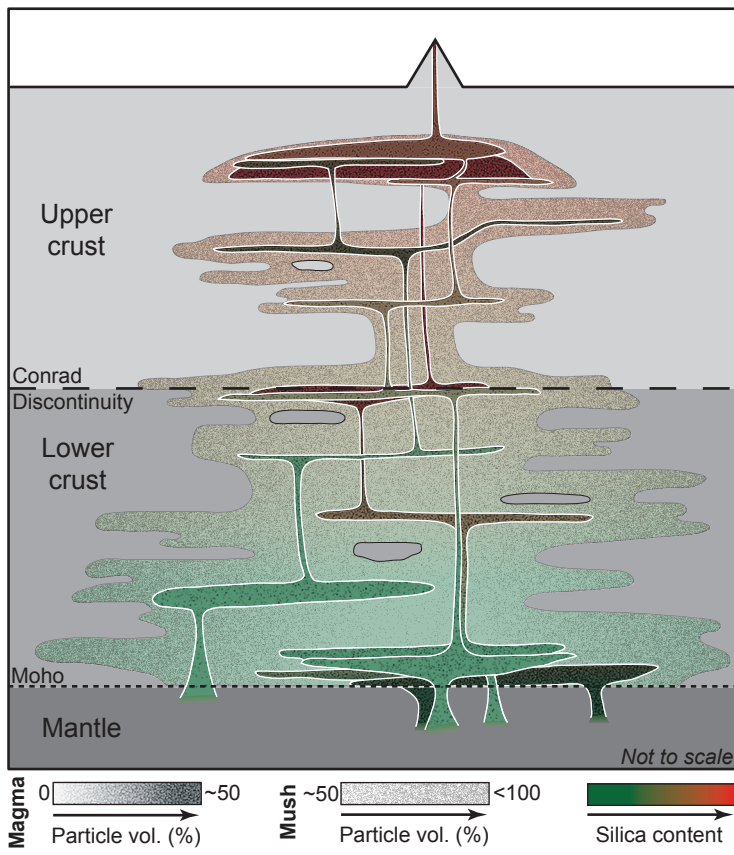


Figure 2

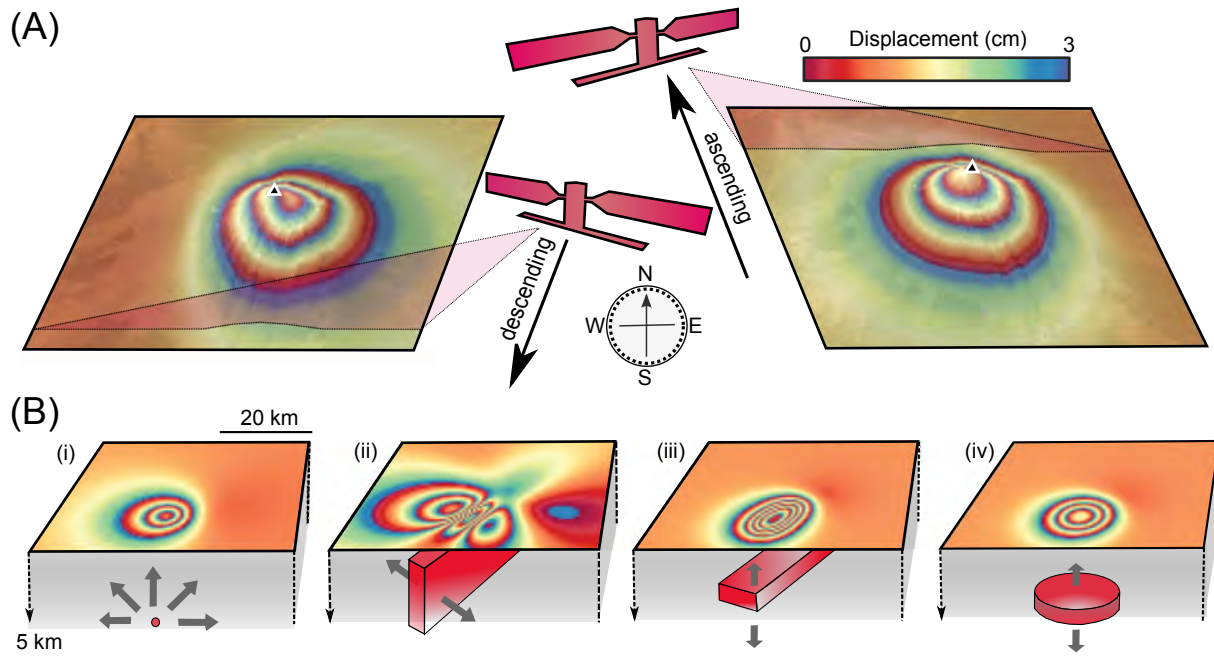


Figure 3

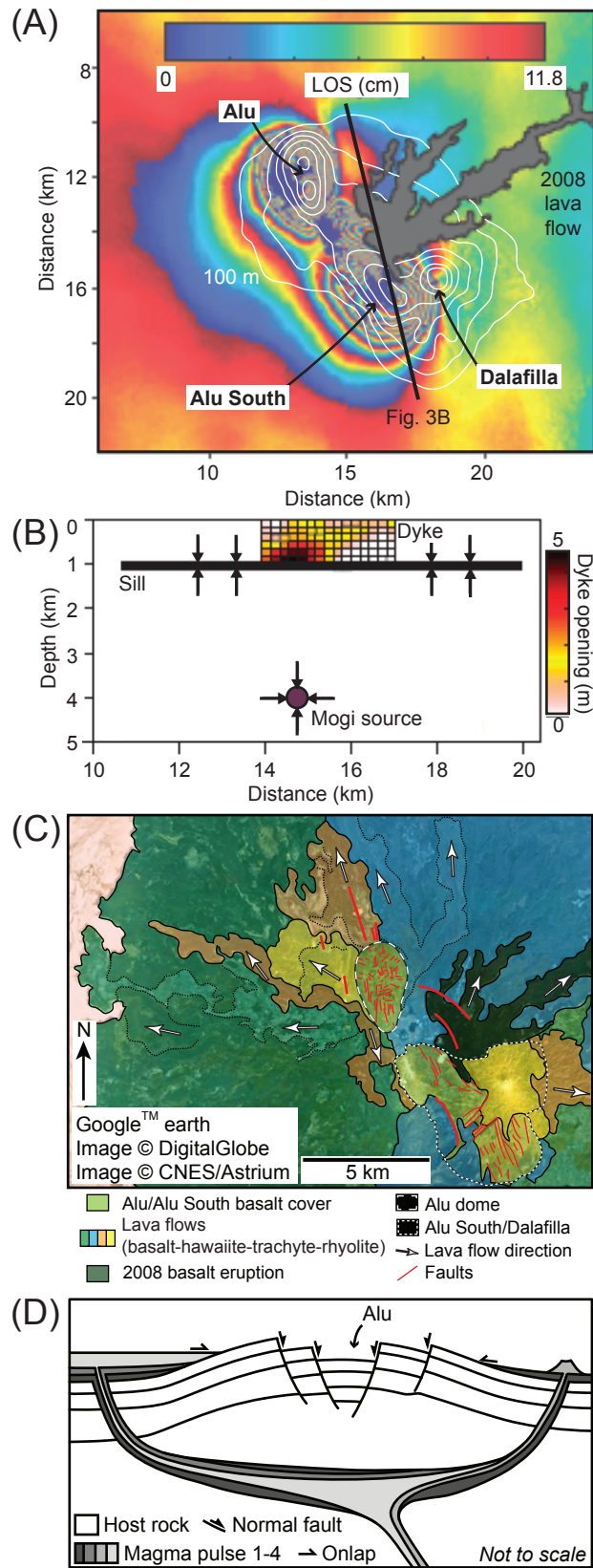


Figure 4

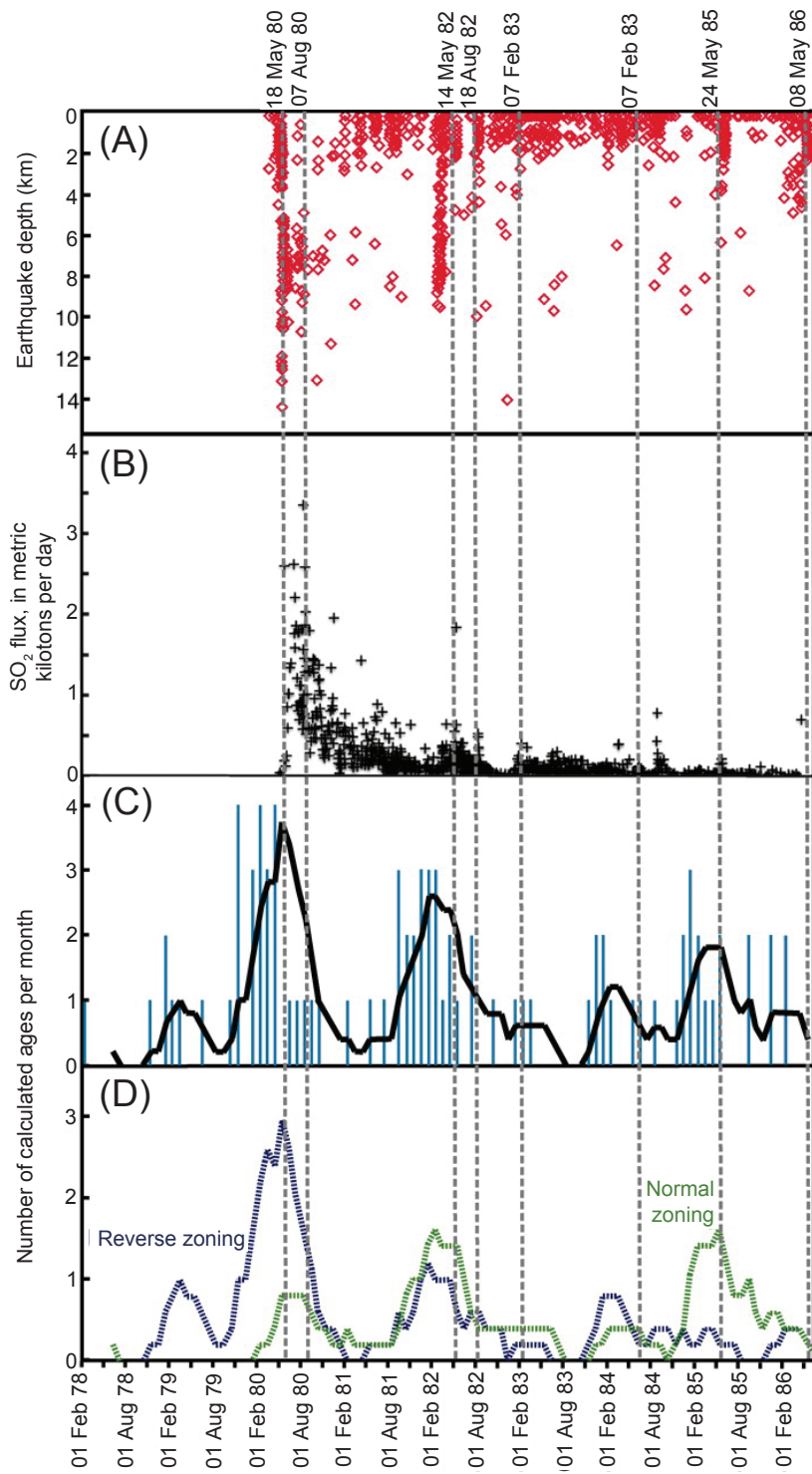


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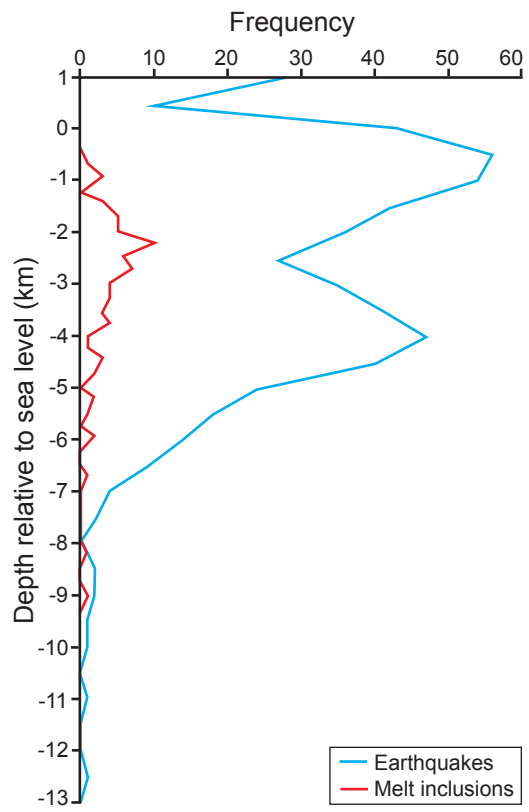


Figure 6

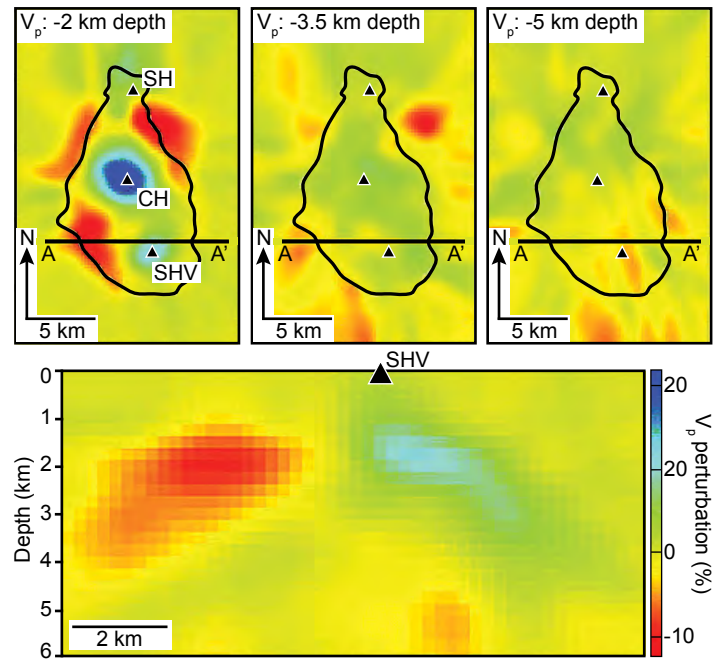
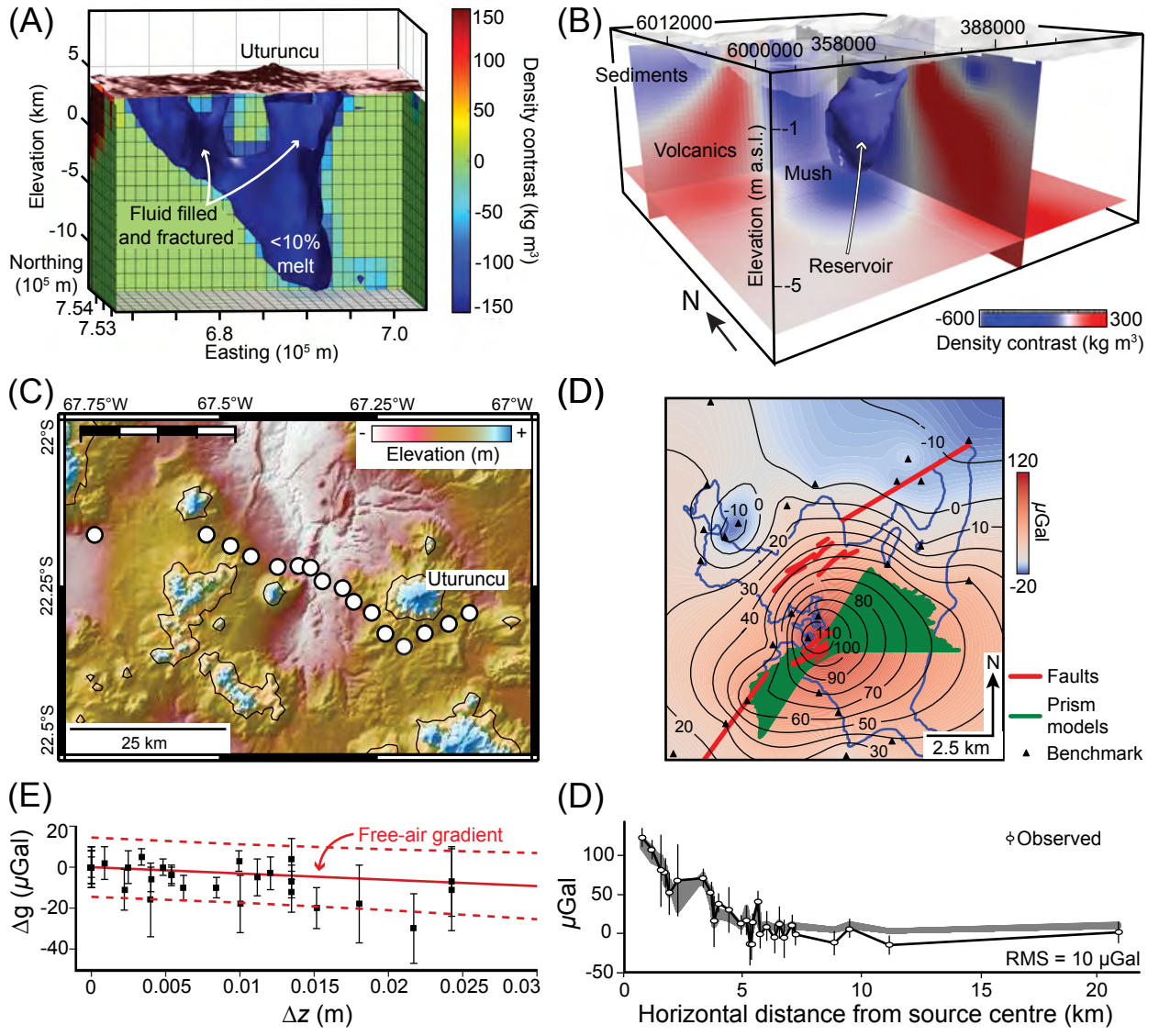


Figure 7



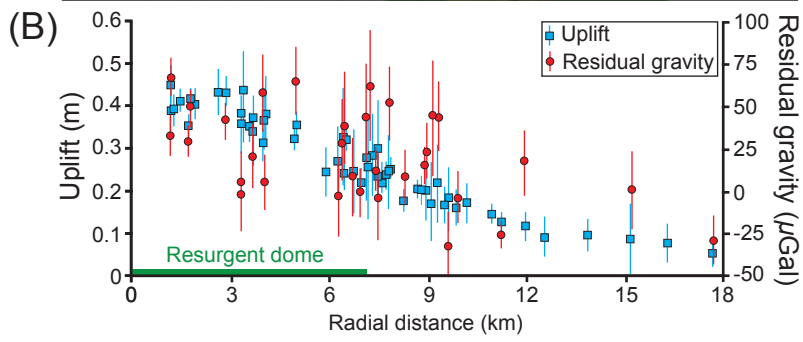
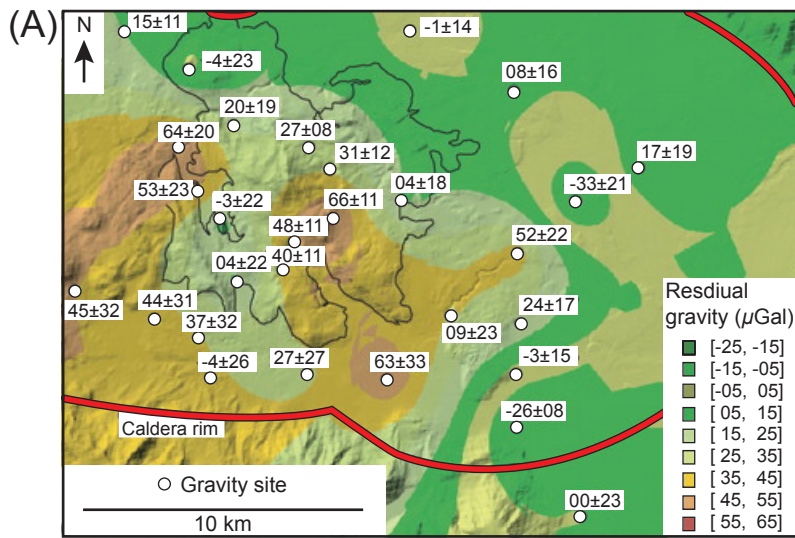


Figure 9

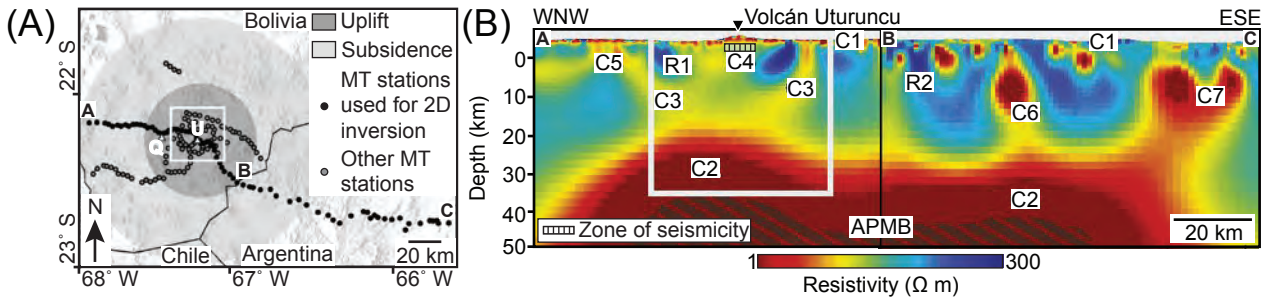


Figure 9

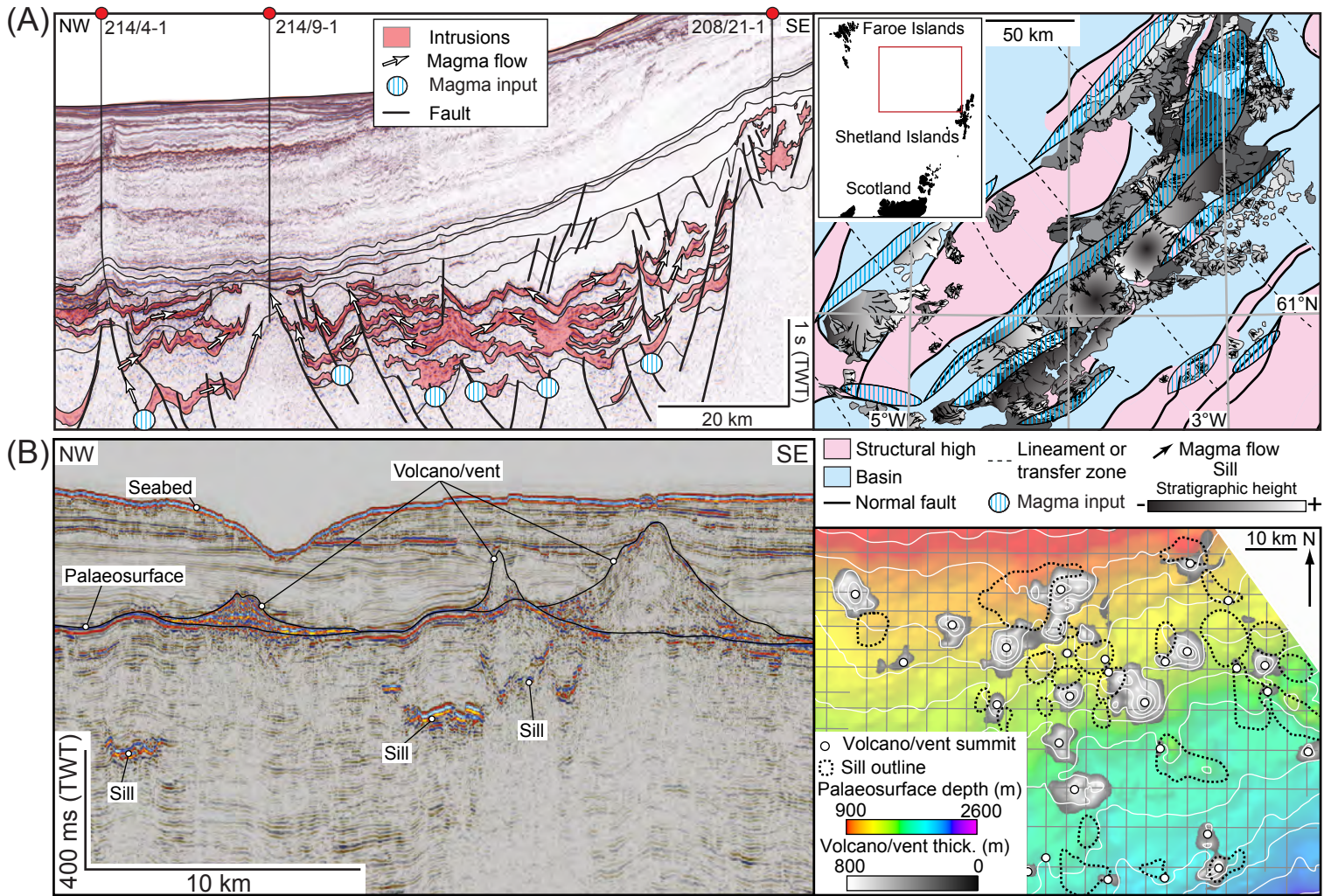


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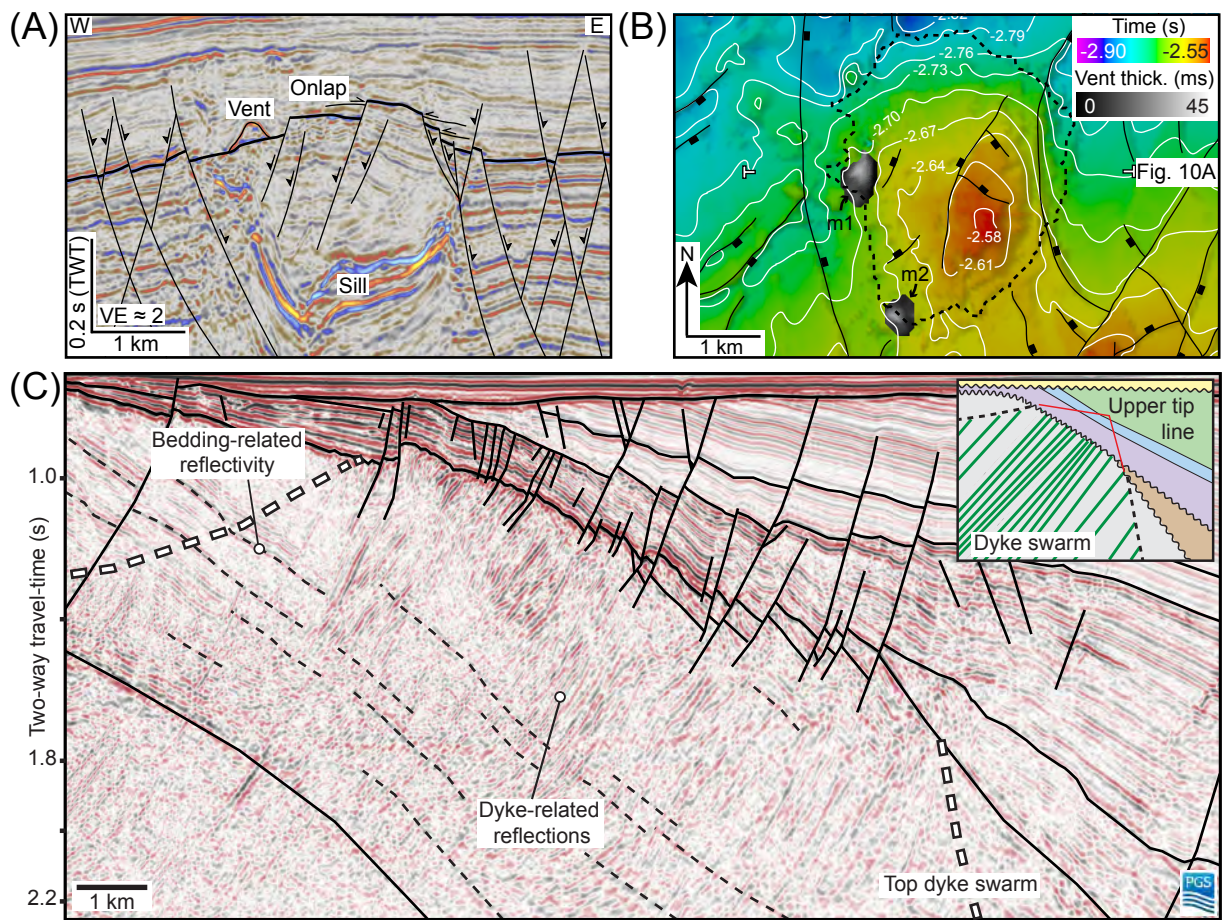


Figure 11

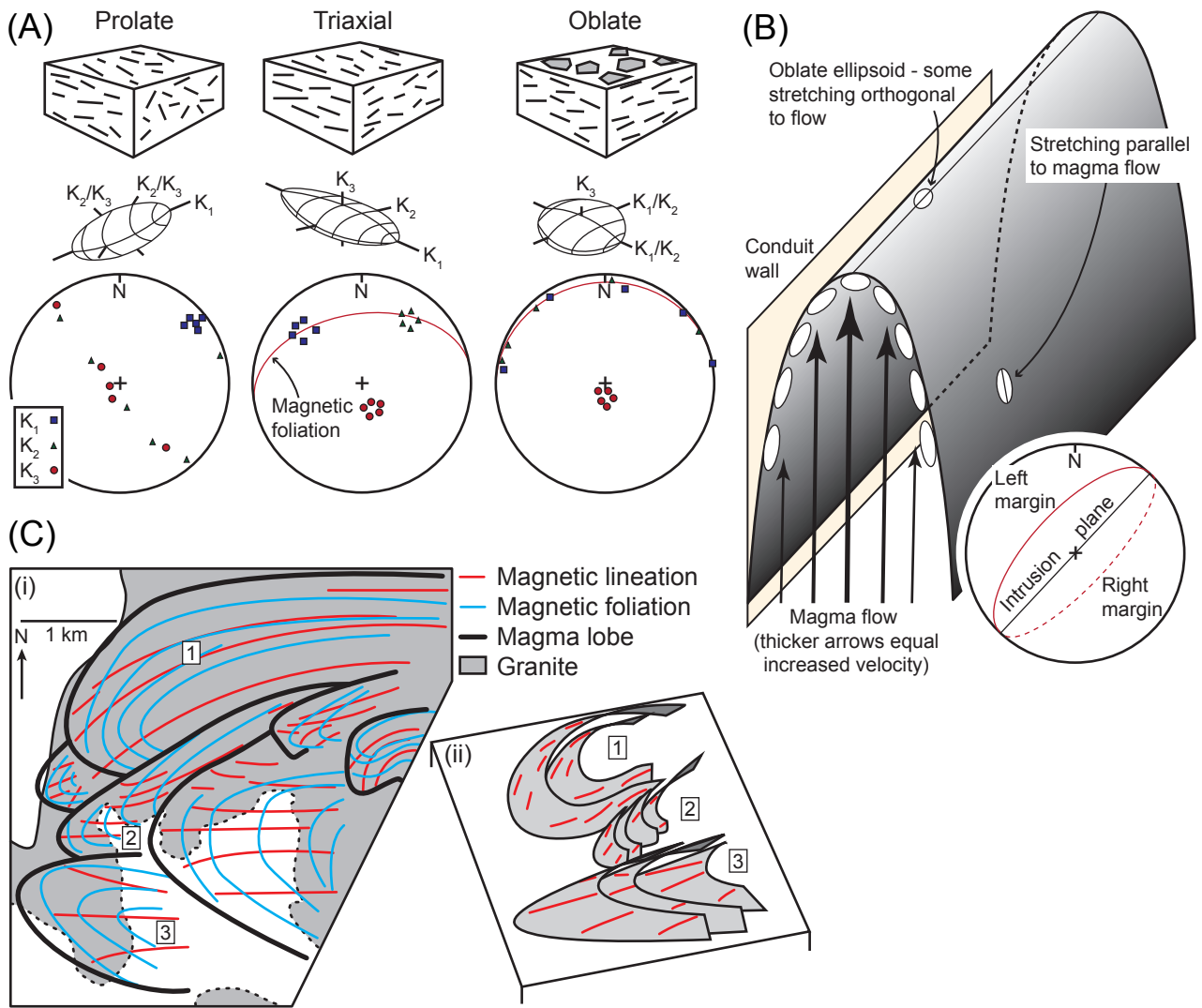


Figure 12

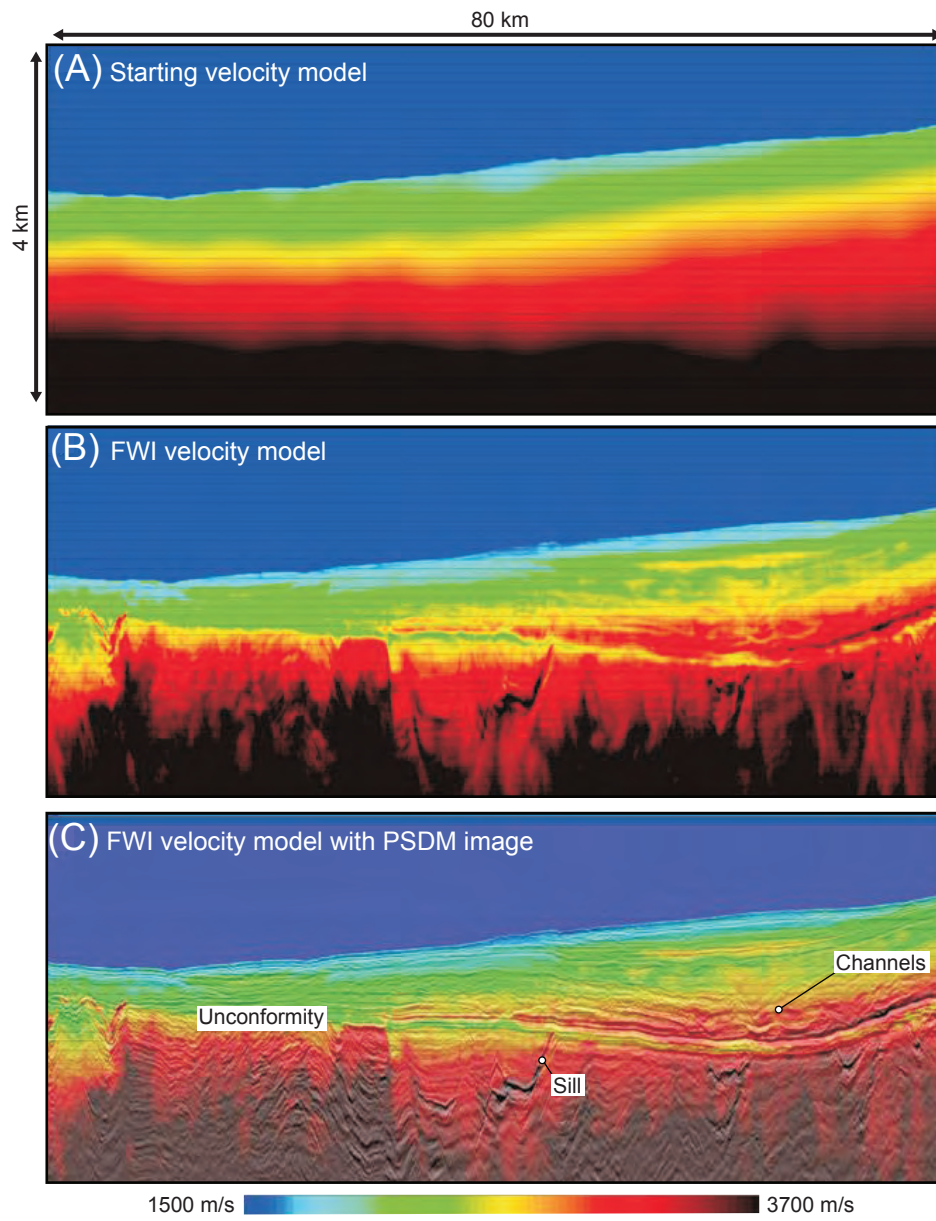


Figure 13

