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Special Section:

Merging geophysical, petro-chronologic and modeling perspectives to understand large silicic magma systems

Key Points:

- Seismic and gravity data are jointly inverted to produce models of v_p and density beneath an active volcano including the magma reservoir
- Temperature and melt distribution are estimated jointly from v_p and are consistent with a mush zone with 6–17% melt fraction
- Analysis of v_p /density cross-plot indicates low melt fraction and low-aspect ratio melt inclusion geometry

Supporting Information:

- Supporting Information S1

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Vertically Extensive Magma Reservoir Revealed From Joint Inversion and Quantitative Interpretation of Seismic and Gravity Data

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Abstract Recent advances in our understanding of arc magmatic systems indicate that melt is stored for long periods in low-melt fraction crystal mushes and that eruptible magma reservoirs are short-lived and are assembled rapidly by buoyancy-induced instabilities and draining of the crystal mush. Many aspects of their architecture remain unclear, particularly in relation to their geometry and shallow melt distribution. We investigate the storage of melt below the active Soufrière Hills Volcano (SHV), Montserrat, using joint geophysical inversion combined with a quantitative interpretation approach based on rock physics. We jointly inverted active-source P -wave traveltimes and gravity anomalies to derive coincident 3-D models of P -wave velocity and density to a depth of 8 km. Comparative analysis of the active SHV and extinct Centre Hills volcano and effective elastic medium computations allow us to constrain temperature, melt fraction, and melt geometry. A continuous column of partial melt is inferred beneath SHV, at 4–8 km depth. Melt fraction is ~6% (ranging from 3 to 13% depending on melt geometry) and is maximum at 5–6 km depth. When under-recovery of the low- v_p volume is taken into account, the melt fraction is revised to ~17% (ranging from 11 to 28%). Analysis of v_p /density cross plots indicates that the melt distribution is best represented by low-aspect ratio geometries. These likely span a multiscale spectrum ranging from grain-scale inclusions and fractures to 100-m-scale dykes and sills. Our results confirm the concept of vertically extensive crystal mush including one or multiple more melt-rich layers.

Plain Language Summary We use data from the Soufrière Hills Volcano on Montserrat to study the storage of magma in the plumbing system of an active volcano. By measuring the time it takes for seismic waves to travel from artificial sound sources to recording instruments located on the island and on the seabed around it, we can determine the speed of sound in the rocks. By measuring small variations in the gravitational attraction, we can determine the density distribution. We then jointly analyze the speed of sound and density and calculate that the degree of melting of the rocks is between 6% and 17% and the total volume of molten rocks is 1.5 to 4.5 km³. The partially molten region extends from 4 to at least 8 km depth. Some more thoroughly molten layers are likely present but are too small to be imaged with our method. Our findings support a model of the magma storage system comprised mostly of solid crystals with a relatively small amount of molten rocks.

1. Introduction

1.1. Magma Reservoirs

The magmatic systems of arc volcanoes extend from the primary melt source in the mantle wedge to the eruptive vent. The details of the ascent history and the thermal, chemical, and mechanical interactions of the melt with the arc lithosphere determine the composition of the magma and the eruptive style and are therefore of importance for understanding past eruptions and the potential for future eruptions and their associated hazards. Recent years have seen a paradigm shift from the concept of a low-crystallinity magma chamber in which silicic magmas are generated by fractional crystallization and crystal settling (e.g., Bateman & Chappell, 1979) to the concept of a mostly solid crystal mush, governed by multiphase porous flow (Cashman et al., 2017; Cooper & Kent, 2014; Jackson et al., 2018), and in which eruptions are driven by inherent buoyancy-induced instabilities (Bergantz et al., 2015; Seropian et al., 2018). Furthermore, it

has been hypothesized that eruptible magma reservoirs with high melt fraction ($> \sim 30\%$) are likely to be quickly assembled and short-lived (Parks et al., 2012; Yu & Lee, 2016). The development of this new paradigm has been driven by geochemical and petrological observations and numerical modeling of melt migration (e.g., Cashman & Blundy, 2013; Costa & Morgan, 2011; Nakamura, 1995; Solano et al., 2012, 2014).

1.2. Volcano Tomography

Geophysical studies at active volcanoes have recovered evidence for low melt fraction ($< 15\%$) reservoirs (Heath et al., 2015; Huang et al., 2015; Kiser et al., 2018; Paulatto et al., 2012) but efforts to quantify magma in the subsurface by geophysicists have faced some interpretative issues and large uncertainties. The elastic properties of rocks are affected by partial melt and can theoretically be used to constrain melt fraction in the field (Hammond & Kendall, 2016; Seccia et al., 2011; Ward et al., 2014). However, the strength of the P -wave seismic velocity (v_p) reduction associated with the presence of melt depends on the microscopic distribution of the melt, leading to a trade-off between melt inclusion shape (represented by the aspect ratio of spheroidal inclusions) and melt fraction (Mainprice, 1997; Paulatto et al., 2012). Other elastic properties, including v_p/v_s ratio, seismic anisotropy, and seismic attenuation, are more sensitive to melt fraction compared to v_p (De Siena et al., 2014; Hammond & Humphreys, 2000; Hammond & Kendall, 2016) but are often harder to measure at high spatial resolution.

One of the best targets for understanding crustal magmatic system is the Soufrière Hills volcanic system, Montserrat, because of the abundance of petrological and geological constraints, including analysis of crustal xenoliths (Kiddle et al., 2010), the existing high-quality geophysical dataset (Hautmann, Camacho, et al., 2013; Voight et al., 2014), and the relatively simple structure and history of the volcano (Wadge et al., 2014), which helps the interpretation of geophysical imaging.

1.3. Study Area

The island of Montserrat is part of the Lesser Antilles intraoceanic volcanic arc (Bouysse, et al., 1990) and is composed of four volcanic edifices that young towards the south (Harford et al., 2002): the Silver Hills (SH, 2.6–1.2 Ma), the Centre Hills (CH, 950–550 ka), the Soufrière Hills Volcano (SHV, 170 ka–present), and the South Soufrière Hills (SSH, 135–125 ka). SHV started a new eruption sequence in 1995 characterized by several pulses alternating with quiescent periods lasting several months to years with the last eruption episode in 2010 (Wadge et al., 2014). The eruptive products (1 km³) consist of a high-crystallinity andesite ($\text{SiO}_2 = \sim 59\text{--}61\text{ wt\%}$; melt content $\approx 35\%$) containing mafic inclusions (Barclay et al., 2010; Devine et al., 1998). The magmatic plumbing system beneath SHV is thought to consist of a vertically extensive mush column with multiple high-melt fraction magma lenses or chambers (Christopher et al., 2015; Edmonds et al., 2016). Thermobarometric studies of erupted lavas indicate that they were stored prior to eruption in a reservoir at > 5 km depth (Devine et al., 1998) and were reheated by injection of a mafic magma (e.g., Murphy et al., 2000; Plail et al., 2014). Inversion of ground deformation data (Elsworth et al., 2008; Hautmann et al., 2010, 2014; Voight et al., 2006) together with petrological data (Edmonds et al., 2016) supports the concept of a multilevel magmatic system comprising a shallow magma reservoir at 6 km depth and a deeper connected reservoir at a depth of at least 12 km. Seismic reflection imaging using local earthquakes found evidence for numerous horizontal reflectors at 6 to 19 km depth, which are interpreted as sill intrusions by Byerly et al. (2010) but may alternatively be caused by compositional layering. P -wave travel-time inversion imaged a low- v_p volume at 5.5–7.5 km depth (Paulatto et al., 2012), which was interpreted as the shallow magma reservoir that fed the recent eruptive episodes. Direct melt estimates from rock physical interpretation of the v_p model range from 1 to 15%. This wide uncertainty range is due to the strong sensitivity of v_p to melt inclusion geometry, which is a poorly constrained parameter. Integration of the tomography model with numerical modeling of the formation of the shallow magma reservoir by incremental intrusion of sills indicates that the observed v_p anomaly is consistent with a combination of a thermal anomaly, a noneruptible mush and a relatively small (~ 13 km³) reservoir with $> 30\%$ melt fraction (Annen et al., 2014; Paulatto et al., 2012). However, seismic traveltime tomography has a tendency to underestimate low- v_p anomalies because of wavefront healing (Malcolm & Trampert, 2011). This fact, combined with a seismic resolution length in the target area of ~ 3 km (Paulatto et al., 2012), means that the maximum strength and size of the low- v_p anomaly are not well determined, and therefore, a wide range of melt volumes and melt fractions are compatible with the v_p model of Paulatto et al. (2012).

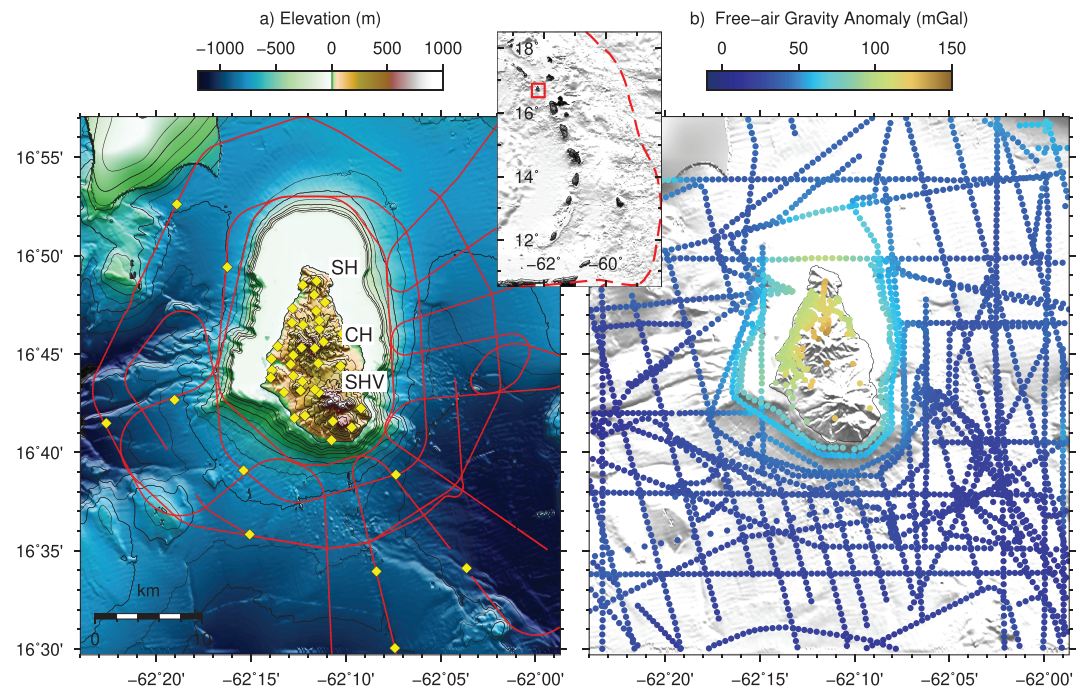


Figure 1. Experiment area and data distribution. (a) Topography map of survey area with seismic acquisition geometry. The yellow diamonds mark station locations; the red dots mark shot locations. (b) Gravity measurements locations and free air anomaly. Inset shows location of Montserrat along the Lesser Antilles arc.

In the following analysis we reprocess the seismic data presented in Paulatto et al. (2012) together with combined land and marine gravity data (Hautmann, Camacho, et al., 2013) and we develop a quantitative method to estimate melt content that addresses and mitigates some of the uncertainties discussed above by integrating multiple physical parameters and incorporating petrological constraints. We use joint geophysical inversion to obtain coincident models of P -wave velocities and density beneath SHV. By analyzing v_P/ρ density cross plots with the help of self-consistent effective medium calculations, we obtain robust constraints on melt fraction and first-order constraints on melt geometry.

2. Data

2.1. Seismic Data

We used seismic data collected as part of the SEA-CALIPSO experiment in December 2007 (Voight et al., 2014) during a quiescent period in the final stages of the last eruption sequence. The dataset includes the recordings of 4414 airgun shots fired along a series of radial lines and circles around the island (Figure 1a). The temporary recording array included 10 ocean bottom seismometers (OBS) equipped with three-component short-period (SP) geophones and a hydrophone, 29 three-component SP seismometers, and 204 SP vertical geophones (Shalev et al., 2010). The shots were also recorded by the permanent seismometer array managed by the Montserrat Volcano Observatory (MVO) (Luckett et al., 2007). For this study we use a subset of the array containing the 10 OBS, 15 three-component SP seismometers, 15 vertical SP geophones, and 5 MVO broad-band seismometers. The land stations were selected among the full set of deployed stations based on data quality and location, to be as widely and regularly distributed as possible. The inverted seismic dataset includes 118,622 first arrival P -wave traveltimes at offsets of up to 45 km. Uncertainties were manually estimated based on signal-to-noise ratio and range from 15 to 100 ms (Paulatto, Minshull, Baptie, et al., 2010; Paulatto et al., 2012). S -waves were not considered as they are not generated directly by the airgun sources and conversions at the seabed and basement interfaces did not provide reliable phases with acceptable signal to noise ratio.

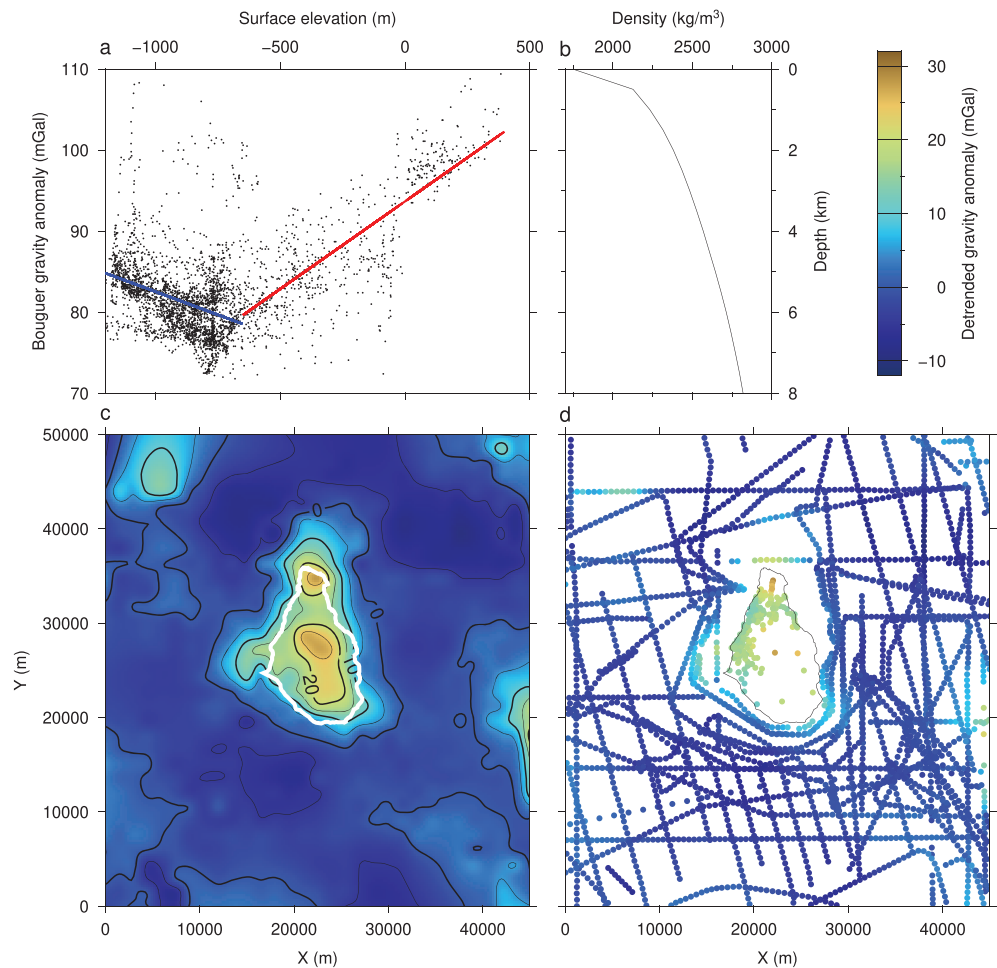


Figure 2. Preprocessing of the gravity data. (a) Correlation of surface elevation and Bouguer gravity anomaly with terrain correction. The blue and red lines show linear regression lines for data points where elevation is < -700 and > -700 m, respectively. (b) Reference density profile. (c) Gridded gravity anomaly. (d) Distribution of gravity anomaly values as used in inversion.

2.2. Gravity Data

We assembled marine and land gravity data from the survey area. The marine data come primarily from two French expeditions: the 1998 Aguadomar cruise [doi: 10.17600/98010120] and the 2002 Caraval cruise [doi: 10.17600/2010030, Le Friant et al., 2004] and a number of other expeditions transiting in the area between 1964 and 1987. The marine gravity data were corrected with static shifts to minimize inconsistencies at cross-over points and subsampled to a minimum inline spacing of 500 m, resulting in 2,975 data points. The land gravity data include 157 measurements collected between April 2011 and July 2012 (Hautmann, Camacho, et al., 2013) (Figure 1b). Uncertainties in the gravity data are 1 mGal for the marine data and 0.3 mGal for the land data (Hautmann, Camacho, et al., 2013). We applied a terrain correction calculated using a land/sea topography grid and the multiscale tessellation implementation of Cella (2015). We found a minimum correlation between the resulting gravity anomaly and topography for a terrain density of 2,450 kg/m³. However, we notice that the land and marine data would require different terrain corrections, with the marine data having a minimum correlation for a significantly lower terrain density than the land data (Figure 2a). This difference is due to the lithological difference between the on-land and submarine subsurface. The high topography on land corresponds to hard-wearing and dense andesite lavas and intrusive igneous rocks and therefore would require a higher terrain density than areas of low topography offshore, which are covered with softer and lower-density volcanoclastic deposits and marine sediments. To account partially for this difference, we chose to calculate a terrain correction using a multilayer density model

based on the average vertical density profile derived from the v_p model of Paulatto et al. (2012) and the Nafe-Drake v_p /density relationship (Brocher, 2005) (Figure 2b). The final gravity anomaly used in the inversion was calculated by removing the average and long-wavelength trend derived by fitting a first order 2-D polynomial to the terrain corrected gravity anomaly (Figures 2c and 2d).

3. Methods

We simultaneously invert the traveltimes and gravity anomalies using the joint inversion framework developed by Moorkamp et al. (2011). Joint inversion with seismic traveltimes and a suitable regularization can help shrink the null space of the gravity inversion (e.g., Heincke et al., 2014) and produce robust density models. We couple the seismic and gravity data into a single objective function and add a minimum cross gradient constraint (Gallardo & Meju, 2004) that imposes structural similarity between v_p and density (ρ). The cross-gradient is defined as the cross product of the spatial gradients of the two properties and can be zero when one of the two properties is zero or when the gradients are parallel (or anti-parallel):

$$\Phi_x = \nabla v_p \times \nabla \rho \quad (1)$$

The full objective function contains two data misfit terms, two model regularization terms, and the cross-gradient term:

$$\Phi = \lambda_t \Phi_{d,t} + \lambda_g \Phi_{d,g} + \lambda_s \Phi_{r,s} + \lambda_\rho \Phi_{r,\rho} + \lambda_x \Phi_x \quad (2)$$

$\Phi_{d,t}$ and $\Phi_{d,g}$ are the data misfits for the seismic traveltimes and gravity anomalies, respectively, and are defined as the l^2 norms of the data residuals weighted by their respective uncertainties. $\Phi_{r,s}$ and $\Phi_{r,\rho}$ are the model norms for v_p and ρ , respectively, and in our case are defined as the second spatial derivatives of the two models weighted by individual a-priori model covariance matrices. The model covariance matrices are used to keep the water layer fixed and to boost the model update at depth. For v_p we use an a-priori model covariance that increases linearly with depth below the sea floor to contrast the decay in ray coverage. For ρ we use an a-priori model covariance that increases as the square of depth to balance the decrease in sensitivity of the gravity measurements with the distance from the causative body (Li & Oldenburg, 1998). The parameters λ_t , λ_g , λ_s , λ_ρ , and λ_x are scalar weights that control the balance between the different terms of the objective function.

The seismic forward problem is solved using a finite difference discretization of the eikonal equation on a regular rectangular grid (Podvin & Lecomte, 1991), which allows the accurate calculation of traveltimes in models with strong velocity contrasts. The gravity forward problem is based on a parallelized analytical solution to the gravitational attraction of a rectangular prism (Moorkamp et al., 2010). The objective function is minimized iteratively using the L-BFGS optimization algorithm (Liu & Nocedal, 1989).

4. Inversion and Results

4.1. Inversion in Practice

The tomographic inversion was run on a regular Cartesian grid with 500-m cell size and extending over $45 \times 50 \times 12.5$ km in the X, Y, and Z directions, respectively. The starting v_p model was generated by taking the average vertical v_p profile from Paulatto et al. (2012) and hanging it from the seabed. v_p in the water layer was determined based on temperature measurements from expendable bathythermograph deployments and salinity measurements from a hull-mounted conductivity sensor. The starting density anomaly model was taken to be zero everywhere. Both the seismic and gravity inversions were seen to be robust with respect to changes in the starting model.

The choice of regularization parameters was guided by the analysis of trade-off curves or L-curves (Hansen, 1992). We first ran a series of independent seismic traveltime inversions to determine the optimal ratio between the seismic data weight λ_t and the v_p regularization weight λ_s and a series of independent gravity inversions to determine the gravity data weight λ_g and the density regularization weight λ_ρ . L-curves are usually plotted for single-iteration inversions to save computation time. We chose instead to run full inversions for each test. Each inversion is run until one of the following conditions are met: (i) the normalized

RMS residual is equal or lower than 1, (ii) a solution with lower total value of the objective function than the previous one cannot be found, and (iii) we reached 100 iterations. About 75% of the data norm reduction is accomplished within the first 10 iterations and 95% within the first 50 iterations. Continuing the inversions beyond 100 iterations leads to negligible data norm improvements in the stand-alone inversions. We found that the optimal parameterization for the seismic inversion corresponds to $\frac{\lambda_s}{\lambda_t} = 10^4 - 10^{4.5}$ (Figure S1 in the supporting information). Weaker regularization led to the inversion terminating early near a local minimum of the objective function. Stronger regularization lead to oversmoothed models. We found a broader trade-off for the gravity inversion and a less clear kink in the L-curve close to $\frac{\lambda_p}{\lambda_g} = 10^2 - 10^{2.5}$ (Figure S2). The kink in the gravity L-curve is likely less pronounced because of the inherent regularization resulting from the rapid decay of the gravity sensitivity with distance from the causative body.

Combining the seismic and gravity data required some trial and error testing to ensure that the different terms of the objective function were reduced at the same rate. We inverted the two datasets together, minimizing a joint objective function, initially without a cross-gradient term ($\lambda_x = 0$). We started by setting λ_s and λ_p to the same value 10^4 and determined λ_t and λ_g based on the ratios obtained from the analysis of the L-curves. We then varied these parameters by up to two orders of magnitude and chose a parameterization that resulted in the largest reduction in both data misfit terms and a comparable amount of detail in the v_p and density models ($\lambda_t = 10^0$, $\lambda_g = 5 \cdot 10^1$, $\lambda_s = 10^4$, $\lambda_d = 10^3$). The last step consisted in determining the optimal weighting for the cross-gradient term. We tested increasing values of the cross-gradient weight λ_x over several orders of magnitude and monitored the total misfit and convergence rate. The optimal λ_x was chosen at the inflection point where the misfit starts to increase significantly ($\lambda_x = 10^{12}$). Variations of up to 1 order of magnitude in any of the weights do not change the recovered structure significantly. The inversion termination criteria were the same as for the individual inversions except that we ran the joint inversion for up to 500 iterations since the minimization of the joint objective function converges more slowly than the individual inversions and requires a larger number of iterations to reach a comparable level of fit (Figure S3). This is due to the nonlinearity of the cross-gradient and the potential conflict of the two data types in some parts of the model.

4.2. Resolution Analysis

We carried out extensive tests to determine the minimum resolution length. We tested the resolvability of sinusoidal checkerboard anomalies with 5 and 8 km cell size (Figures S4 and S5 in the supporting information). For the 5 km checkerboard we tested the recovery of two layers of anomalies, while for the 8 km checkerboard we tested the recovery of a single layer of anomalies. The v_p and density anomalies are spatially correlated but have independent amplitudes. The v_p anomalies have a 10% maximum amplitude, while the density anomalies have a 300 kg/m³ maximum amplitude. Both the 5 and 8 km checkerboards are well recovered in the standalone seismic traveltime inversion (Figures S4 and S5), with some visible smearing. The gravity inversion is very effective in recovering the first layer of checkerboards (both 5 and 8 km cell size) but cannot recover the deeper layer in the 5 km cell size case (Figures S4 and S5). In the joint inversion the v_p resolution is slightly improved and smearing is reduced. The density resolution is improved, but the deeper layer of checkerboards is still unresolved. We also run a series of boxcar spike tests showing that the average depth of isolated cubic anomalies with 4 km side is well recovered, but the density structure is smoothed more than the v_p structure (Figure S6). The cross-gradient constraint significantly reduces vertical smearing and improves the resolvability of deep density anomalies compared to the independent gravity inversion when the assumption of structural similarity is valid. In summary the seismic resolution length is smaller than 5 km within the target area beneath the island of Montserrat. Horizontal gravity resolution length is comparable. The vertical gravity resolution is poorer but is improved with the introduction of the cross-gradient constraint.

4.3. V_p and Density Structure

At shallow depth (<3 km), we observe three main high- v_p bodies (4,500–5,500 m/s) with roughly circular footprint (Figure 3), each corresponding to one of the three volcanic centers that make up the island as previously described by Paulatto, Minshull, Baptie, et al., 2010, Paulatto et al., 2010, Paulatto, Minshull, & Henstock, 2010]. The high v_p beneath SHV is slightly less prominent than beneath CH and SH. Two

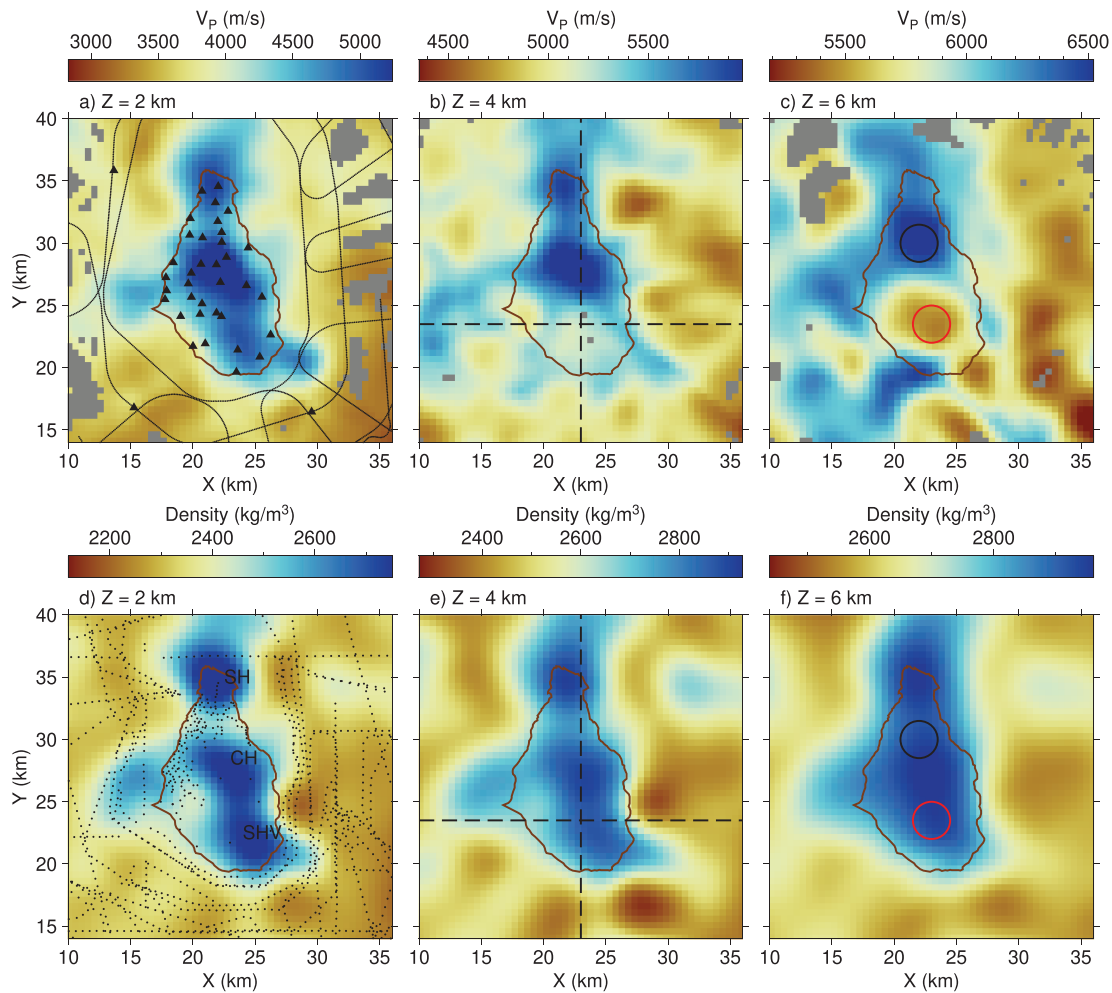


Figure 3. v_p and density at 2, 4, and 6 km depth in the final model of the joint inversion. The black triangles mark the seismic station locations. The black lines mark the shooting lines. The small black dots mark the gravity measurement locations. The dashed lines mark the location of the profiles shown in Figures 4 and 5. The red and blue circles mark the low- v_p volume and the “pluton” discussed in Figure 9. The gray areas have no ray coverage.

smaller high- v_p bodies extend west of CH (Bransby Point) and east of SHV (Roche’s Buff) likely related to the Belham valley fault zone (Kenedi et al., 2010). Low v_p is observed at 2–4 km depth beneath the minor volcanic cones west of SHV (at $X = 18$ km in Figure 5a). Beneath the surrounding seafloor v_p is lower (3,000–4,000 m/s) and consistent with the thick sedimentary cover observed in multichannel seismic reflection sections (Kenedi et al., 2010; Watt et al., 2012).

At depth greater than 3 km beneath CH and SH a region of high v_p extends continuously to at least 7–8 km depth, where the model resolution drops (Figures 4 and 5). v_p increases rapidly, reaching 6,300–6,600 m/s at 6 km depth, and then increases more gradually or remains stable. The maximum in v_p is located between CH and SH. Beneath SHV the high- v_p core transitions into a low- v_p anomaly that extends from 4 to at least 7.5 km depth (Figures 5) and likely continues deeper. There is a weak velocity inversion at 3.25 to 4.25 km depth that brings v_p down to a local minimum v_p of 5200 m/s (Figure 8b). The low- v_p anomaly is pear-shaped (Figure 6a), with a roughly circular footprint with a diameter of ~6 km at 7 km depth, narrowing to 4 km at 4 km depth. It is surrounded by a ring of elevated v_p (Figure 3). The maximum v_p difference between the low- v_p anomaly beneath SHV and the high- v_p core beneath CH and SH is of -1200 m/s at 6 km depth (Figure 8b), considerably higher than the anomaly of about -700 m/s estimated by Paulatto et al. (2012).

The v_p structure retrieved in our models is broadly similar to that described in Shalev et al. (2010) and Paulatto et al. (2012). Our model extends deeper than that of Shalev et al. (2010) (8 km compared to 5

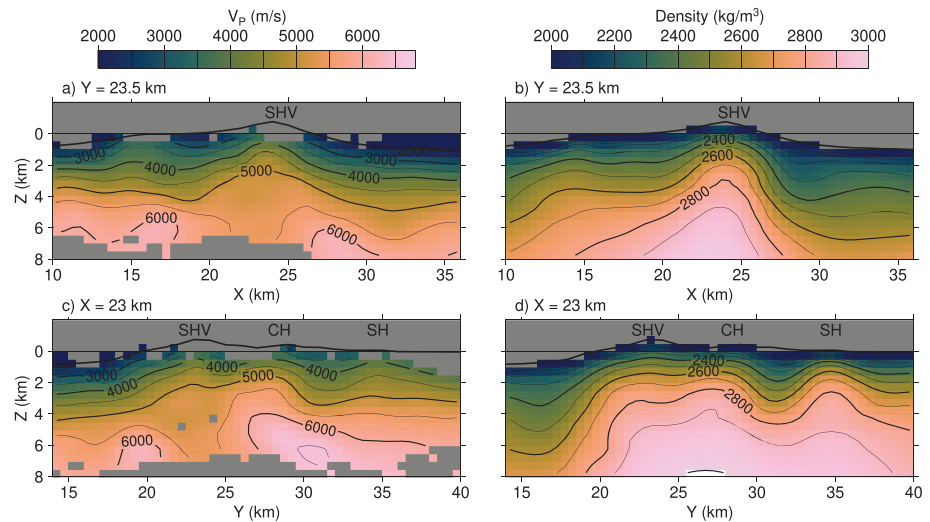


Figure 4. v_P and density along vertical cross-sections at $Y = 23.5$ km and $X = 23$ km. See Figure 3 for the location of the profiles. The gray areas have no ray coverage.

km) because we include the longer offset OBS data. Our model is also more detailed than that of Paulatto et al. (2012) due to our more robust ray tracing algorithm. Paulatto et al. (2012) used a ray shooting approach that is unable to predict traveltimes in the presence of strongly irregular structure and their model was therefore more strongly regularized than strictly required by the data. The additional constraints on v_P from the gravity data introduced through the cross-gradient term only weakly influences the v_P structure and has the positive effect of weakening the artifacts generated by the acquisition geometry.

The density anomaly distribution broadly resembles the v_P anomaly. The main features are high density at the cores of the three volcanic centers and in coincidence with the smaller high- v_P anomalies west of CH and east of SHV (Figure 3). The high-density cores extend vertically to the bottom of the model and are stronger

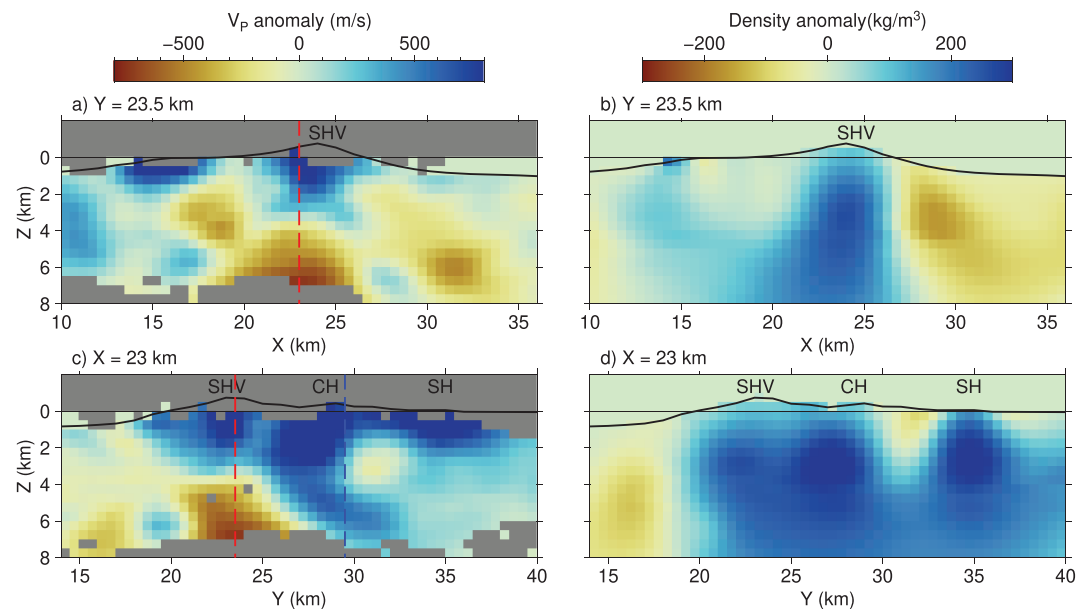


Figure 5. v_P and density anomaly along vertical cross-sections at $Y = 23.5$ km and $X = 23$ km. See Figure 3 for the location of the profiles. The red and blue dashed lines mark the vertical profiles shown in Figure 8 and correspond to the locations of SHV and CH respectively. The gray areas have no ray coverage.

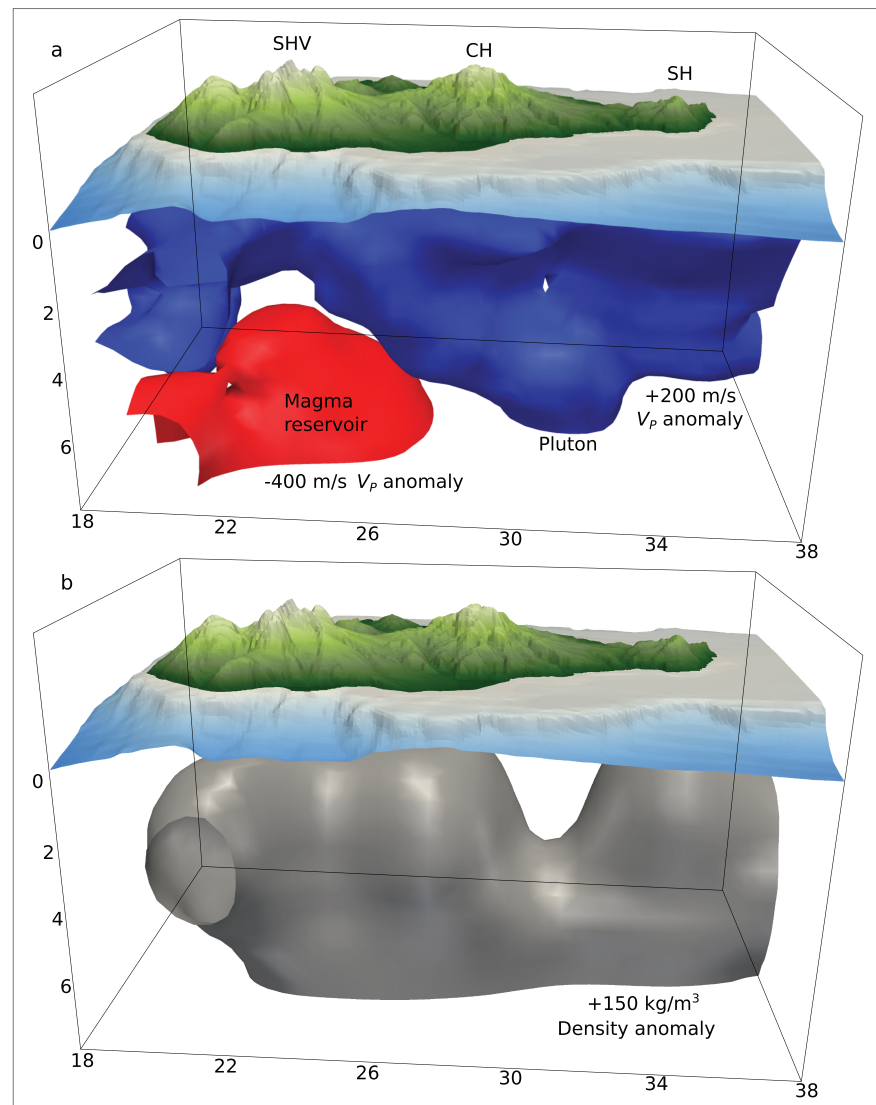


Figure 6. Three-dimensional view of the (a) v_p anomaly and (b) density anomaly. The surface topography is plotted above the model with a 2:1 vertical exaggeration.

under CH, where they reach $+400 \text{ kg/m}^3$ (Figure 5). We generated a model of absolute density by adding the reference density model used for the terrain correction to the final density anomaly model. Absolute density inside the high-density cores reaches $2,800\text{--}3,000 \text{ kg/m}^3$ (Figure 4), while the surrounding sediment and arc crust have lower densities ($2,300\text{--}2,700 \text{ kg/m}^3$). There is no evidence of a significant reduction coincident with the low- v_p anomaly beneath SHV.

5. 5 Discussion

The structure at shallow depth ($<3 \text{ km}$) can be summarized as comprising two domains: the island's edifice with high v_p and density and the surrounding region with low v_p and density. This is consistent with the island having formed on top of a thick sedimentary cover that accumulated in the back-arc of the older island arc (Bouysse, 1986) and with the development of extensive subaerial and submarine volcanoclastic fans that make up the flanks of the island (Kenedi et al., 2010). The cores of the volcanic edifices are composed of intrusive igneous rocks and extrusive lavas of intermediate composition that are faster and denser than the volcanoclastic fans and marine sediments that surround them. The low v_p observed west of SHV (Figure 4a) is likely attributable to high porosity and metasomatism associated with the hydrothermal

system, as previously suggested by Shalev et al. (2010). At greater depth (>3 km) we observe high v_P and density beneath the extinct volcanic centers (CH and SH) and low v_P and high density beneath SHV (Figure 3). The high- v_P -high-density cores become less distinct at depth and merge into a single feature elongated in the N-S direction (Figure 6). The measured v_P and density are broadly consistent with the presence of intrusive igneous rocks of intermediate and mafic composition (Kiddle et al., 2010). The low v_P and relatively high density beneath SHV can be explained by high temperatures and partial melt and are analyzed in detailed in sections 5.2–5.4 after a description of the rock physical relations adopted.

5.1. Physical Property Calculation

To link the measured seismic properties to melt fraction, we use an implementation of effective elastic medium theory. Partially molten rocks are multiphase materials comprised of solid crystal grains, a viscous liquid melt, and likely a compressible magmatic volatile fluid phase (e.g., Wallace, 2005). The macroscopic elastic properties (the effective medium properties) of the crystal/melt/volatile composite depend on the properties of the components, on their respective abundances, and on their geometry and connectivity. For simplicity we consider a two-phase material composed of solid crystal grains and melt, without a volatile phase. The presence of volatile bubbles in the melt has the effect of reducing its bulk modulus and v_P (Kieffer, 1977); therefore, by ignoring the volatiles we are likely to overestimate melt content (Figure 7b). We use the self-consistent scheme (SCS) developed by Berryman (1980a, 1980b) to predict the elastic properties of a multiphase material with spheroidal components with any given aspect ratio ($a < 1$: oblate, $a = 1$: spheres, $a > 1$: prolate). The effective elastic properties are calculated by considering the effect on stress and strain caused by the introduction of inclusions of each constituent. None of the constituents is singled out to act as a background matrix; instead, the background medium is the effective medium with properties to be determined. The advantage of this approach is that all the constituents are treated on the same level and are assigned an inclusion geometry and volume fraction. The interaction between different inclusions is implicitly taken into account, leading to a behavior that includes a critical melt fraction (ϕ_c) where the rigidity of the medium drops rapidly and there is a transition from crystal supported to fluid supported stress (Figure 7a). This transition in elastic properties has been observed in laboratory experiments (Caricchi et al., 2008) and also corresponds to a strong decrease in viscosity (e.g., Lejeune & Richet, 1995; Suckale et al., 2016).

If all the components are homogeneous and isotropic and the inclusions are randomly orientated, the mixing relationships are as follows:

$$\begin{aligned} \sum_{i=1}^N c_i (K_i - K^*) P^{*i} &= 0 \\ \sum_{i=1}^N c_i (\mu_i - \mu^*) Q^{*i} &= 0 \end{aligned} \quad (3)$$

K^* and μ^* are the effective moduli to be determined, K_i and μ_i are the bulk and shear moduli of the constituents and c_i are their volume fractions. The factors P^{*i} and Q^{*i} contain all the information on the inclusion geometry, represented by the aspect ratio a , and are called the *polarization coefficients* as they are calculated from the stress polarization tensor introduced by Eshelby (1957) and named by Kröner (1958) because of its analogy with the polarization vector in electrostatics. The stress polarization tensor relates the far field stress to the stress inside the inclusion. Expressions for P^{*i} and Q^{*i} for oblate and prolate spheroids can be found in Berryman (1980b). The above equations have to be solved by iteration because the polarization coefficients also depend K^* and μ^* . A suitable initial guess for the elastic moduli is the Voigt-Reuss-Hill average (Chung, 1963).

Equation 3 provides the unrelaxed elastic moduli valid in the high-frequency approximation, assuming that the melt inclusions are isolated, and the pore fluid has no time to flow and equilibrate stress imbalances. To calculate the relaxed moduli, one needs to first calculate the moduli of the dry matrix using equation 3 and then use Gassmann's fluid substitution equation (Korringa et al., 1979) to derive the saturated moduli (Mavko, 1980). The frequency band of our seismic experiment (3–60 Hz) is likely to be lower than the characteristic frequencies of the dominant relaxation mechanisms in partially molten rocks, including melt squirt, viscous shear, and Biot's flow (Schmeling, 1985); therefore, in our calculations we use the relaxed moduli and seismic velocities.

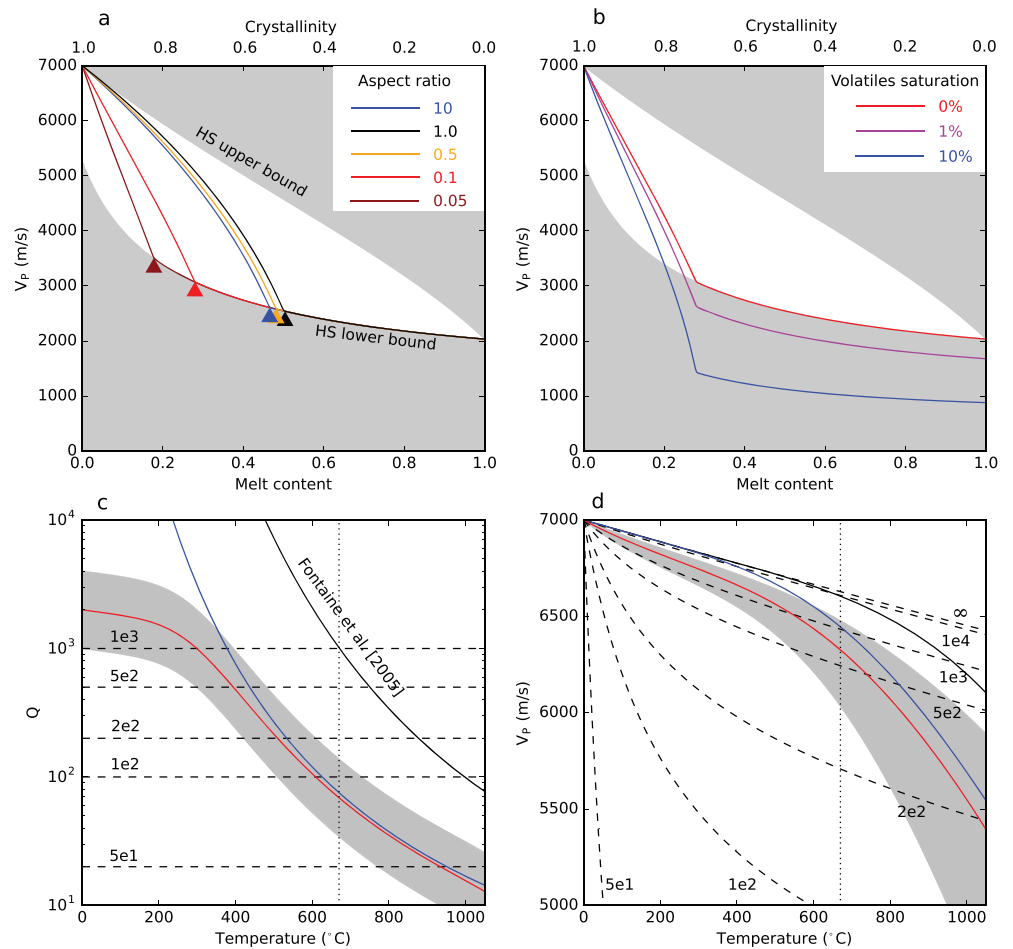


Figure 7. Physical properties. (a) Variation of P -wave velocity for a crystal/melt composite with increasing melt content calculated using the SCS method in the low-frequency limit. Different colors correspond to different aspect ratios of melt inclusions. The triangles mark the critical melt content above which the composite behaves as a suspension of particles in a fluid. The HS bounds stand for the Hashin-Shtrikman bounds (Hashin & Shtrikman, 1961). (b) Variation of P -wave velocity for a crystal-melt composite with the addition of volatile bubbles (aspect ratio = 1.0) within the melt. The melt pockets are assumed to have an aspect ratio of 0.1. The volatile phase is assumed to be supercritical water. (c) Seismic attenuation model used for the temperature calculation. The black curve corresponds to the model by Fontaine et al., (2005); the blue curve corresponds to the same model with parameters tuned for arc intrusive rocks; the red curve corresponds to the model with Q clipped at low temperature. The shaded gray area represents a factor of 2 uncertainty. (d) v_p dependence on temperature for the Q models shown in (c), compared to predictions for constant Q (dashed curves). The constant Q models produce unrealistic results.

The SCS predicts a faster decrease in v_p with melt fraction for melt in low-aspect ratio inclusions ($a = 0.05$) compared to spherical inclusions ($a = 1.0$) and high-aspect ratio inclusions ($a = 10$) (Figure 7a). The point where the curves reach the lower Hashin-Shtrikman bound corresponds to the critical melt fraction, ϕ_c . In the low frequency approximation, ϕ_c for a partially molten gabbro is predicted to be 18%, 28%, and 48% for $a = 0.05$, 0.1, and 0.5 respectively. The cases with $a = 1$ and $a = 10$ give similar results to $a = 0.5$. Analysis of Mn zonation in garnets from dacitic rocks suggests that the critical melt fraction in crustal systems is $\sim 30\%$ (Yu & Lee, 2016). The erupted SHV lavas contain 50–65% phenocrysts and microphenocrysts (Sparks et al., 2000), suggesting a critical melt fraction close to $\sim 35\%$. Comparing these observations with Figure 7a, we conclude that an inclusion aspect ratio of 0.1–0.2 would be suitable for our modeling; however, the seismic waves are sensitive to melt distribution at multiple scales, from the pore-scale to the scale of dykes and sill; therefore, the chosen melt geometry for seismic interpretation does not necessarily reflect the microscopic pore geometry. In addition, melt geometry is dependent on melt fraction, on the crystal

and melt composition, and on volatile content; therefore, a single aspect ratio can only represent a crude approximation of reality. At low melt fraction melt is likely to be present in thin melt tubes ($a \gg 1$) and films ($a \ll 1$) along crystal edges and faces. As the melt fraction increases, they become thicker. A detailed analysis of melt inclusion geometry for the specific composition of interest, supported by laboratory measurements, would be needed to justify the choice of a more complicated melt geometry model. In our analysis we choose $a = 0.1$ as our preferred model but given the uncertainties discussed above, we show a range of melt fraction predictions for a between 0.05 and 0.5 representing a minimum and maximum melt fraction consistent with the measured v_P .

The effect of increasing temperature (T) on v_P of the solid matrix is calculated based on the approach of Karato (1993):

$$\frac{\partial \ln v_P}{\partial T} = \frac{\partial \ln v_P}{\partial T} \bigg|_a - F(\alpha) \frac{Q(\omega, T)^{-1} H}{\pi R T^2} \quad (4)$$

where Q is the seismic attenuation or quality factor, H is the activation enthalpy, R is the universal gas constant, and $F(\alpha)$ is a function of α , the exponent describing the frequency dependence of Q and is close to 1 when α is small. The first term on the right-hand side of the above equation is the derivative due to anharmonic effects and can be taken to be a constant C . We use $C = -8.1 \cdot 10^{-5} \text{K}^{-1}$ based on experimental data for gabbro (Christensen, 1979). The second term accounts for anelastic effects and can be simplified by taking $F(\alpha) = 1$ (no frequency dependence).

When solving equation (4), most authors pick a constant value of Q (Dunn et al., 2000; Goes & van der Lee, 2002). This is acceptable when calculating relatively small temperature anomalies around a central temperature; however, in our case we wish to correct v_P from ambient temperature to temperatures above the solidus. For this reason, we take into account the dependence of Q on temperature, using the attenuation model of Fontaine et al. (2005):

$$Q^{-1} = A \cdot \left[\omega^{-1} \cdot d^{-1} \cdot e^{-\frac{E}{RT}} \right]^\alpha \quad (5)$$

where C is a preexponential constant, ω is the seismic wave frequency, d is the grain size in micrometers, and E is the activation energy of creep. In the shallow crust the activation energy and activation enthalpy are approximately the same, since the activation volume is negligible. This exponential expression predicts a large Q ($>10^6$) at temperatures close to $T = 0^\circ\text{C}$ and a rapidly decreasing Q as the solidus temperature is approached (black curve in Figure 7c). The values of A , E , C , and α are generally poorly constrained and need to be determined experimentally. Fontaine et al. (2005) determined the following values for the parameters in equation 5 based on measurements of shear wave attenuation on gabbro samples from the Oman ophiolite: $\alpha = 0.08$, $E = 8.73 \cdot 10^5 \text{ J/mol}$, $C = 34.72 \text{ s}^{-\alpha} \mu\text{m}^{-\alpha}$, $d = 1,000 \mu\text{m}$. Since we are interested in P -wave attenuation we start by dividing the expression in equation 5 by a factor of 2.25 (e.g., Romanowicz & Mitchell, 2015). The gabbro samples are water-poor and have a solidus temperature of $\sim 1050^\circ\text{C}$. In contrast, the middle crust at Montserrat is likely to be hydrous (Edmonds et al., 2016) and therefore has a much lower solidus temperature ($\sim 670^\circ\text{C}$) and a lower activation energy (Rybacki et al., 2006). In the absence of accurate data, we choose to keep the above values for C and d , and we set $\omega = 10 \text{ s}^{-1}$, $H = E = 3 \cdot 10^5 \text{ J/mol}$, which is close to the activation enthalpy of diabase (Caristan, 1982), and $\alpha = 0.15$. This combination of parameters results in a curve that has similar properties to the one determined by Fontaine et al. (2005) but is shifted to lower temperatures (blue curve in Figure 7c) to account for the lower solidus and earlier onset of creep in hydrous conditions. We add an additional term to the expression in equation 5 to limit Q at low temperatures to below a value $Q_0 = 5 \cdot 10^3$ (red curve in Figure 7c); this allows us to fit the v_P predicted from analysis of crustal xenoliths (Kiddle et al., 2010). Putting together our Q model with equation 4, we predict a decrease of v_P with increasing T that is initially approximately linear and accelerates as the solidus temperature is approached (Figure 7d), as observed in laboratory measurements on crustal rocks (e.g., Kern & Schenk, 1985). A v_P decrease of $\sim 10\%$ is predicted for a temperature increase from 0°C to the solidus at 670°C . A more typical scenario would be an increase from mid-crustal temperatures of 400 to 670°C , which would result in a $\sim 5\%$ v_P decrease.

The effect of temperature on density is modeled using the thermal expansion equation:

$$\alpha_V = \frac{1}{V} \frac{\partial V}{\partial T} = -\frac{1}{\rho} \frac{\partial \rho}{\partial T} \quad (6)$$

assuming a constant volumetric thermal expansion $\alpha_V = 10^{-7}/^\circ\text{C}$ (Bauer & Handin, 1983). The above equation can be easily solved analytically. Given ρ at a reference temperature T_0 , we have

$$\rho(T) = \rho(T_0)e^{-\alpha_V(T-T_0)} \quad (7)$$

5.2. Composition of the Crust

We compared the vertical profiles at CH with predicted v_p and density for intrusive igneous rock samples brought to the surface as xenoliths embedded in erupted lavas (Kiddle et al., 2010) (Figure 8a). The most common xenoliths recovered in the SHV lavas are Hornblende gabbro (Hb-gabbro), noritic and gabbroic anorthosite (N/G-anorthosite) and quartz diorite (Qtz-diorite). We also compared the observed profiles to the properties of the mafic enclaves. Based on the major element composition the Hb-gabbro has a predicted v_p of 6,880 m/s and density of 2,940 kg/m³ at atmospheric pressure and temperature (Hacker & Abers, 2004). When corrected for pressure and temperature found in arc crust at 4–8 km depth, v_p decreases to 6,500–6,700 m/s (Figure 8a), while density is only slightly decreased to 2,930 kg/m³. For the v_p correction due to temperature we use equations 5 and 6 and a reference geothermal gradient of about 50°C/km (T_{CH} in Figure 8c). This geothermal gradient is estimated assuming a surface heat flux of 101 mW/m² (Manga et al., 2012); a radiogenic heat production of 0.48 mW/m³, based on the average Pb, Th, and K content (Kiddle et al., 2010; Rybach, 1988; Zellmer et al., 2003); and a thermal conductivity increasing from 1.8 Wm^{−1} · K^{−1} at the surface to 2.4 Wm^{−1} · K^{−1} at 3 km depth. The thermal conductivity was estimated based on the Voigt-Reuss-Hill average of the conductivities of the main mineral phases (Čermák & Rybach, 1982) and assuming a porosity decrease from 20% at the surface to 5% at 3 km depth. We obtained a very close match between the predicted gabbro v_p and density and the CH profiles at depth greater than 6 km (Figure 8a), indicating that the inferred temperature gradient of 50°C/km and the chosen Q model are appropriate for the arc crust away from active volcanoes. We can also conclude that porosity is decreased to negligible amounts beneath 5–6 km depth. We interpret the high v_p and density region at 5–8 km depth beneath CH and SH as an old intrusive complex of mafic composition, likely similar to the recovered xenoliths. The lower v_p and density in the shallow crust are primarily governed by porosity so we cannot make detailed inferences on rock composition from physical properties. It is highly likely however, that the composition of the upper 5 km is intermediate, based on the composition of the eruptive products (e.g., Devine et al., 1998).

5.3. Temperature and Melt Estimate

The low- v_p anomaly at 4–8 km depth beneath SHV indicates the presence of partial melt as previously shown by Paulatto et al. (2012), but the lack of evidence for a corresponding significant density reduction suggests that the average melt fraction is relatively low. The relatively large uncertainties and low resolution of the density model at the target depth mean that a direct melt estimate from the density model is not meaningful. This was expected, given the relatively small effect of melting on density. Assuming a matrix density of 2,900 kg/m³ and a melt density of 2,250 kg/m³, 20% melt fraction would correspond to a <5% density decrease. We will consider density jointly with v_p later in this section, but first we reevaluate the melt fraction necessary to explain our observations using only v_p . We loosely follow the approach of Paulatto et al. (2012), with several notable improvements including a more realistic thermal model based on available geological data and consistent with the melt calculation, a realistic Q model, a thorough discussion of uncertainties, and integration of available petrological data.

A major pillar of our analysis is the observation that the eruptive products of CH are very similar to those of SHV (Zellmer et al., 2003); therefore, the structure and composition of their crustal magmatic systems and intrusions can be expected to be similar. We can treat CH as cooled and solidified analog of SHV. Any differences in v_p may be attributed to increased temperature and melt beneath the active SHV and at shallow depth to differences in porosity. Since we are interested in the maximum melt fraction within the low- v_p anomaly, for this comparison we considered the minimum v_p profile within a 3 km diameter circle

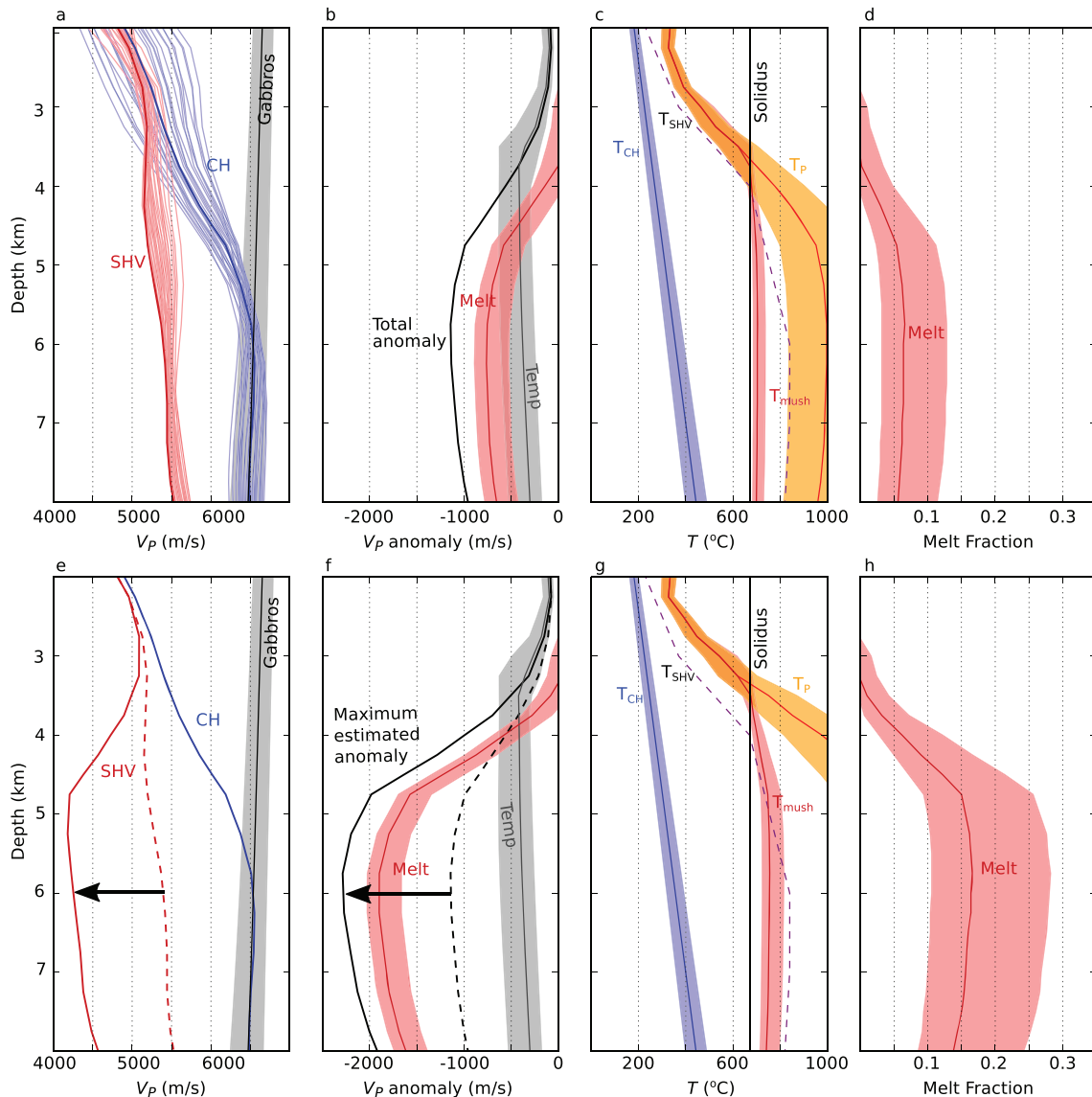


Figure 8. Temperature and melt estimates. (a) Blue: vertical v_P profiles within a 3 km diameter circle centered on CH, the thick blue curve represents the median curve; red: v_P profiles within a 3 km diameter circles centered on SHV, the thick red curve represents the minimum curve; gray band: range of v_P expected for gabbroic rocks based on mineral composition measured by Kiddle et al. (2010), corrected for temperature and pressure. (b) v_P anomaly. Black curve: recovered v_P anomaly; gray: anomaly due to hotter geotherm at SHV; red: portion of recovered anomaly due to partial melt. The width of the shaded area represents uncertainties in the thermal calculation, mostly due to uncertainties in Q and in the reference geotherm. (c) Temperature estimates. Blue (T_{CH}): background arc geotherm, width of blue shaded area represents a 10% uncertainty in geothermal gradient; black: solidus of hydrous arc intrusive rocks; orange (T_P): predicted temperature assuming entirety of v_P anomaly is due to raised temperatures; red (T_{mush}): temperatures beneath SHV calculated from v_P consistently with melt content; dashed gray (T_{SHV}): temperature beneath SHV based on geological data. (d) Melt prediction based on recovered v_P anomaly. The width of the shaded area represents uncertainties in thermal calculation and in melt inclusions aspect ratio. The minimum melt fraction estimate corresponds to the lower Q and aspect ratio = 0.05. The maximum melt fraction estimate corresponds to the larger Q and aspect ratio = 0.5. (e–h) Same as a–d, taking into account the estimated 50% recovery of the low- v_P anomaly.

(Figure 3) centered on SHV. The v_P anomaly was calculated with respect to a reference v_P profile representative of CH, calculated as the median vertical profile of the cylindrical volume within a 3 km diameter circle centered on CH (Figure 8a).

First, we calculated the temperature profile that would be expected at SHV if the v_P anomaly was entirely due to increased temperatures by inverting equation 5 using a simple grid search. The predicted temperature T_P (Figure 8c) crosses the solidus of arc intrusive rocks at about 4 km depth; therefore, it is an unphysical result beneath this depth, and it indicates that some melt is required to explain the observations. We can compare

T_P with a temperature profile for SHV (T_{SHV} in Figure 8c) based on geological data: a temperature of 230°C at 2 km depth from temperature measurements in geothermal boreholes (Brophy et al., 2014), a temperature of 370°C at 3 km depth corresponding to the basal depth of the hydrothermal system, the solidus temperature (670°C) at 4 km depth, and a temperature of 850°C at 6 km depth. The two profiles differ slightly at 2–4 km depth, but for the melt calculation they are equivalent as what matters is the solidus temperature and the depth at which it is reached in the crust; therefore, for the following calculation we use the predicted temperature profile T_P obtained by inverting v_P . Melt inclusion data from erupted lava samples indicate that melt in the upper crust at SHV contains 1–9 wt% water (Edmonds et al., 2016); therefore, for the solidus we used a temperature of 670°C (Johannes, 1984). A difference in the solidus temperature of $\pm 50^\circ\text{C}$ would not significantly affect our results. With the temperature clipped at the solidus we calculated the temperature anomaly with respect to the reference arc geotherm discussed in the previous section and we estimated the portion of the v_P anomaly due to the higher temperatures at SHV to be between -300 and -600 m/s (Figure 8b), corresponding to a 4–8% v_P anomaly. The uncertainty range in v_P anomaly was estimated assuming a factor of 2 uncertainty in the seismic attenuation factor Q and a 10% uncertainty in the reference geotherm. At depth between 3 and 4.5 km the increased temperature beneath SHV is responsible for most of the observed v_P anomaly. Beneath 4.5 km depth the increase in temperature is responsible for $\sim 40\%$ of the v_P anomaly.

The remaining v_P anomaly can be explained by the presence of partial melt. Melt fraction was estimated using the SCS assuming the pore fluid to be a silicate melt with 71 wt% SiO_2 (75 wt% of the nonvolatile components) (Devine & Rutherford, 2014) and 6 wt% H_2O (Humphreys et al., 2010; Edmonds et al., 2016), a density of 2173 kg/m³, a v_P of 2,035 m/s, and a bulk modulus of 9.0 GPa (Ueki & Iwamori, 2016). Temperatures above the solidus were estimated jointly and consistently with melt fraction assuming a melt fraction of 0% at the solidus and 35% at 850°C (Murphy et al., 2000). With these assumptions, the predicted melt fraction ranges from 3 to 13%, depending on the melt geometry. Our preferred estimate is 6% for the case with 0.1 aspect ratio inclusions. We obtained the lowest melt fraction assuming a representative aspect ratio of the melt inclusions of 0.05, corresponding to either grain boundary melt films or mesoscopic (larger than the grain size, but smaller than the seismic resolution length) high-melt fraction veins and channels. The higher limit was calculated assuming an aspect ratio of 0.5, corresponding to subspherical microscopic melt pockets or mesoscopic melt bodies. Aside from the aspect ratio, the other main sources of uncertainty are the reference thermal structure and the sensitivity of v_P to temperature.

The greatest source of uncertainty, however, is the tendency of traveltime tomography to underestimate low- v_P anomalies due to wavefront healing, which means that first arrival ray paths avoid the strongest low- v_P areas (Malcolm & Trampert, 2011). It follows that the smoothing of seismic tomography does not necessarily conserve the effect of partial melt and provides instead a minimum melt fraction estimate. Anomaly recovery test with synthetic models containing idealized 5 km wide mush zones with 10%, 20%, and 30% melt fraction and a melt inclusion aspect ratio of 0.1 show that with the given experiment geometry only 90–42% of the low- v_P anomaly amplitude can be recovered (Figure S7). We can take this observation into account by recalculating the melt fraction assuming that the measured v_P anomaly magnitude is 50% of the real anomaly (Figure 8f) leading to a revised melt fraction estimate of 11%–28%, with a central value of 17% (Figure 8h). This melt fraction estimate should be understood as an average over the seismic resolution length, which for depths of 4–8 km is 3–5 km (see resolution tests and the discussion in Paulatto et al., 2012). The estimated melt fraction is maximum at 5–6 km depth (Figures 8d and 8h), decreasing slowly below this depth, and is consistent with estimates based on mechanical modeling of the plumbing system (Chen et al., 2018). The maximum in melt content coincides with the depth of the pressure source identified by analysis of ground deformation data collected during the 2003 lava dome collapse (Chen et al., 2018; Voight et al., 2006) and during a lava extrusion episode in December 2008/January 2009 (Hautmann, Hidayat, et al., 2013). The shallow magma reservoir corresponding to the pressure source was estimated to have a volume of 4–8 km³ and is therefore smaller than the seismic resolution length, so we cannot expect to fully recover it and image it in detail.

To narrow the range of melt fraction compatible with the observed v_P anomaly, we could incorporate additional petrological information, for example, constraints on the dihedral angles or detailed melt topology information expected for the given composition. Based on recovered xenoliths in the erupted lavas, the

main mineral phases expected in the upper crust beneath SHV are plagioclase and hornblende (Kiddle et al., 2010), which have dihedral angles with liquid phases of $25 \pm 11^\circ$ and $28 \pm 14^\circ$, respectively (Holness, 2005). Takei (2002) calculated that melt in equilibrium with solid grains with dihedral angles of $25\text{--}30^\circ$ has an equivalent softening effect to a spheroidal melt geometry with inclusion aspect ratio of 0.1–0.15, which would lead to a melt fraction just above the central value in Figure 8 ($\sim 17\%$). However, the applicability of dihedral angles is debated, and high-resolution scanning electron microscope images suggest that dihedral angles alone may not be sufficient to determine bulk rock properties since the grain wetness varies with melt fraction and may have a disproportionate effect on seismic velocities and attenuation (Mu & Faul, 2016). In addition, dihedral angles are only related to the geometry of microscopic melt inclusions, while for the understanding of averaged seismically derived properties mesoscopic melt distribution may be more relevant. Accurate constraints on the thermal structure and Q distribution and the dependence of Q on temperature and seismic frequency would also narrow the uncertainty range of the melt fraction estimate.

5.4. Multiparameter Interpretation

A complementary approach to reducing the nonuniqueness of the melt estimation is to jointly interpret our v_p and density models. Figure 9a shows a v_p /density cross-plot for all model cells that are well resolved. The first unsurprising observation is that the points cluster around the Nafe-Drake curve. This is expected since the Nafe-Drake relation was used to generate the average background density model. The Nafe-Drake curve is an experimentally derived v_p /density relationship that is valid for most crystalline rocks, including igneous rocks with a wide range of silica content. Deviations from this curve can be attributed mainly to differences in porosity, melt fraction, and temperature. The standard deviation from the curve is 546 m/s in v_p and 104 kg/m³ in density. In comparison, the cross-plot produced by individual inversions without the cross-gradient constraint is more widely spread out. We also notice that there are systematic variations in the v_p /density relation. Points in Figure 9a are colored based on the elevation of the seabed or land surface above the point. Offshore points (negative elevation) tend to lie closer to the Nafe-Drake curve, while onshore points tend to lie below the Nafe-Drake curve, except for the deepest parts of the model. In other words, onshore parts of the model have relatively lower v_p for the same density, possibly a consequence of lower porosity and higher temperature.

We have highlighted two regions of particular interest: in red the cells lying inside the low- v_p volume beneath SHV, corresponding to the present day magma reservoir, and in blue the cells lying inside the high- v_p volume beneath CH at depth between 5 and 8 km, which we interpreted to correspond to a solidified pluton, once representing the upper part of the CH volcano's magma reservoir (hereafter called “the pluton”). The locations of the two volumes are shown in Figure 3. The pluton has the highest v_p and density found throughout the model and lies 200–500 m/s below the Nafe-Drake curve, with $v_p = 6,550 \pm 180$ m/s and density = $2,950 \pm 60$ kg/m³. The magma reservoir has comparable density ($2,940 \pm 50$ kg/m³) but significantly lower v_p ($5,470 \pm 130$). This low- v_p anomaly is distinct from other low- v_p anomalies in having a relatively high density and therefore it lies considerably below the Nafe-Drake curve on the cross plot.

To understand how temperature and melt fraction affect the v_p /density relationship and interpret the differences between the magma reservoir and the pluton, we calculated how hypothetical rock samples on the Nafe-Drake curve would be affected by a change in either temperature or melt content. We disregarded water-saturated porosity variations as they are unlikely to play an important role at >5 km depth in the arc crust where porosity is very small as demonstrated previously. First, the effect of temperature on v_p was modeled using equation 5 and the effect of temperature on density was modeled using the thermal expansion equation and assuming a constant thermal expansion coefficient (equation 7). Raising the temperature by 200°C causes v_p to decrease by ~ 300 m/s and density to decrease by less than 20 kg/m³, resulting in a downward shift from the Nafe-Drake curve in the cross-plot (Figure 9b). Second, we used the SCS to model the effect on v_p resulting from changes in melt fraction assuming idealized spheroidal melt inclusion geometry. The reduction in v_p once again depends on the inclusion aspect ratio, with low-aspect ratio inclusions resulting in a stronger reduction in v_p (greater softening) than sub-spherical inclusions and high-aspect ratio inclusions. The effect on density was modeled as a simple weighted average and is independent of the inclusion geometry. An increase in melt fraction results in a more rapid decrease in v_p and a smaller decrease in density, and thus a shift below the Nafe-Drake curve (Figure 9c). In contrast, similar calculations carried

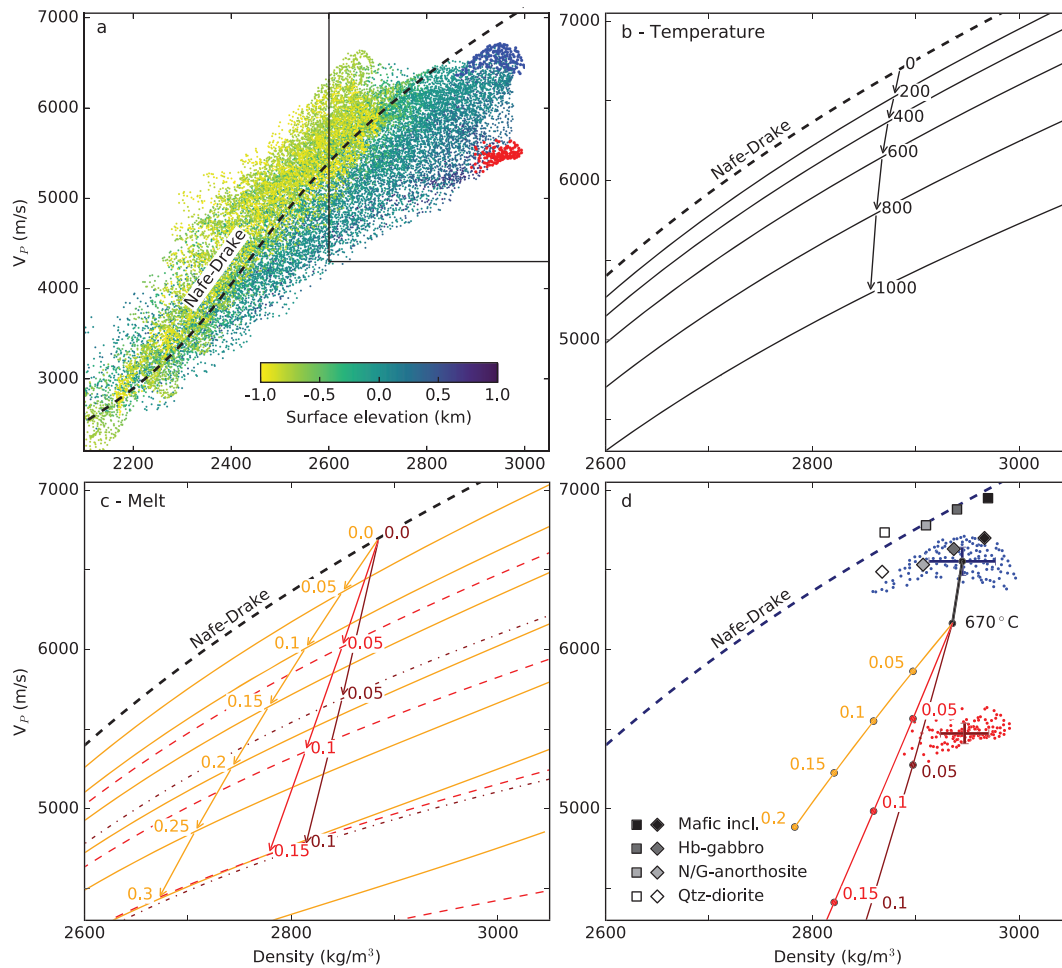


Figure 9. v_p /density cross-plots. (a) Point cloud including all well-resolved model cells. Points are colored based on the surface elevation at the corresponding (X , Y) location. Dashed line: Nafe-Drake relation as parameterized by Brocher (2005); blue dots: pluton; red dots: low- v_p volume beneath SHV. The rectangle outlines the area of particular interest shown in b, c and d. (b) Effect of temperature on v_p /density relation. Thin black curves represent predicted v_p and density for increasing T from 0 to 1000°C. Arrows represent the trajectory followed by a rock sample starting with $v_p = 6700$ m/s, $\rho = 2,880$ kg/m³, and $T_0 = 0^\circ\text{C}$. (c) Effect of melt on v_p /density relation for different aspect ratios of inclusions. Orange: aspect ratio = 0.5; Red: aspect ratio = 0.1; Dark red: aspect ratio = 0.05. Number annotations represent melt fraction. (d) Trajectory in v_p /density space from pluton to magma reservoir. The blue and red dots represent the points inside the pluton and magma reservoir respectively as in (a). The black, gray and white squares correspond to the properties of the xenoliths and mafic inclusions at ambient temperature and pressure. The black, gray, and white diamonds are the same but for a temperature of 345°C and pressure of 0.2 GPa. The dark blue and dark red crosses represent the average properties of the pluton and magma reservoir. The black line represents the change in properties of the pluton when the temperature increases from $T_0 = 325^\circ\text{C}$ to the solidus at 670°C. The orange, red and dark red lines represent an increase in partial melt assuming an inclusion aspect ratio = 0.5, 0.1, and 0.05 respectively.

out for water-saturated porosity predict a reduction in v_p and density of roughly equal amplitude, resulting in a shift that does not deviate drastically from the Nafe-Drake curve.

To reproduce the v_p /density of the pluton, we can start with the properties of the gabbroic xenoliths. At atmospheric temperature and pressure, they lie close to the Nafe-Drake (squares in Figure 9d). After we apply a correction to a temperature of 345°C and pressure of 0.2 GPa (diamonds in Figure 9d), they shift downward over the point cloud of the pluton. Compared to the average properties of the pluton (dark blue cross in Figure 9d), a further increase in temperature and some partial melt are required to reach the v_p and density of the present-day magma reservoir (dark red cross in Figure 9d). As already indicated by the independent v_p analysis and shown in Figure 8, we estimate that roughly 50% of the difference is attributable to a 300–500°C increase in temperature from a background temperature of 350–400°C at the pluton to slightly above the solidus (700–800°C) within the present-day magma reservoir. The remaining v_p and density differences can be attributed to 3–12% melt fraction, depending on the aspect ratio of the melt inclusions. The path

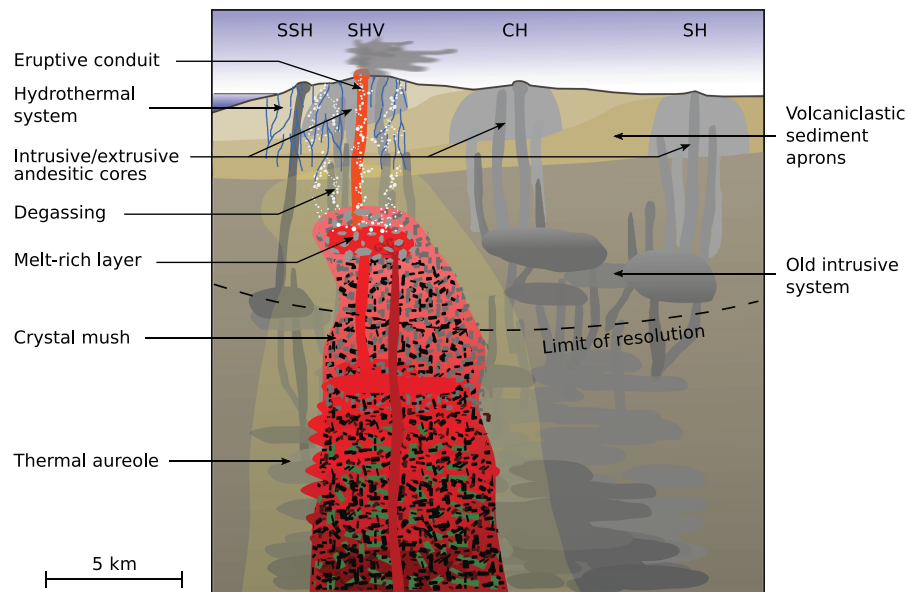


Figure 10. Schematic representation of the magmatic system of Montserrat. The imaged low- v_p volume is interpreted to represent the top of a vertically extensive mush zone including multiple melt-rich layers. A cooled and solidified system is observed adjacent to the active system beneath the extinct volcanoes of SH and CH.

in the v_p /density space linking the pluton and magma reservoir is closer to the predicted curve for low-aspect ratio inclusions ($a = 0.05$) and therefore is more consistent with the lower end of the melt fraction estimate. However, the separation between the low-aspect ratio and high-aspect ratio curves is too small compared to the density uncertainties to allow strong inferences on the melt geometry to be made based on the cross-plot. The magma reservoir density may also be overestimated since it is situated beneath a high-density body and would suffer from vertical smearing as shown in the resolution analysis. This may explain why the predicted densities for all three aspect ratios considered are lower than the measured density of the magma reservoir.

5.5. Interpretation

Taking the preferred melt inclusion aspect ratio of 0.1 and assuming a cylindrical mush volume with 5 km diameter and 4 km height and melt fraction varying linearly between 6 and 17% at the center of the anomaly and 0% at the edge, we estimate a total melt volume of 1.5 and 4.5 km³ distributed over a volume of ~80 km³ at 4 and 8 km depth beneath SHV. If the melt is contained within pockets of eruptible magma (35% melt fraction), this would equate to a maximum eruptible volume in the range of 4 to 13 km³. However, a significant proportion of this volume is likely to be distributed in noneruptible parts of the reservoir. The prediction of a continuous column of partial melt between 4 and 8 km depth supports the concept of a vertically extensive low-melt fraction crystal mush surrounding one or multiple melt-rich layers (Figure 10), with the top one at 5–6 km depth. This model is intermediate between the simple model of two discrete high-melt-fraction stacked reservoirs (Elsworth et al., 2008) and that of a vertically elongated low-melt-fraction reservoir (Voight et al., 2010). Any melt-rich bodies (>35% melt) must be smaller than the seismic resolution length (3–5 km diameter) and cannot be fully imaged with our method. If their size is greater than ~500 m, as is predicted based on ground deformation data (Hautmann, Hidayat, et al., 2013; Voight et al., 2006), they should be resolvable with higher resolution seismic imaging techniques such as full waveform inversion (Morgan et al., 2013). Such techniques have strict requirements on spatial density and frequency bandwidth of the recording array and seismic sources and are unsuitable for application to the data presently available at Montserrat. Based on our analysis, we recommend a two-pronged approach to further improve the robustness of melt fraction estimates from seismological and other geophysical data: (i) to employ, when possible, waveform inversion methods that are less affected by wavefront healing and are better able to fully recover low- v_p anomalies compared to traveltime based methods, and (ii) to measure multiple parameters, including v_p/v_s ratio, seismic attenuation and electrical resistivity, which can be interpreted jointly and thus overcome the melt fraction/geometry trade-off.

6. Conclusions

Joint inversion of first arrival seismic traveltimes and gravity data with a cross-gradient constraint provides the first multiparameter model of the Montserrat magmatic system. The magma reservoir of the active Soufrière Hills Volcano is characterized by low v_P and relatively high density. Temperature and partial melt calculations using effective medium theory and detailed analysis of the v_P /density cross-plot show that a temperature anomaly of 300–500°C and a melt fraction of 3–13% are required to explain the observed v_P and density beneath SHV. If the effect of wavefront healing is taken into account, the melt estimate is revised to 11–28%. The maximum predicted melt fraction is found at 5–6 km depth and is likely to correspond a high-melt fraction (>35%) layer smaller than the seismic resolution length. This high-melt fraction layer lies at the top of a partial melt region that extends to at least 8 km depth, supporting the concept of a vertically extensive mush zone (Cashman et al., 2017; Edmonds et al., 2016). With the available data, the joint seismic/density inversion and interpretation cannot fully resolve the aspect ratio/melt fraction trade-off. However, the method presented can be applied to a larger range of geophysical data and parameters, including S-wave velocity, v_P/v_S ratio, seismic attenuation, and electrical resistivity and is a clear way forward to accurately image in-situ melt fraction and geometry.

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