Density Structure of Kīlauea volcano: Implications for magma storage and transport

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Short title: Density structure of Kīlauea volcano
Section: Heat flow and volcanology
Abstract

A Bayesian linear regression to determine the bias in the Nafe-Drake relationship between compressional velocity and density provides an improved model for the density structure of Kīlauea volcano, Hawai‘i. In previous work, we combined the results of seismic tomography with the Nafe-Drake relationship between compressional velocity and density to explain the large values of gravity disturbances overlying the summits and rift zones of the island’s volcanoes. These results were used to determine mechanisms for gravitational instability of the island flanks. Here we use laboratory measurements of the relationship of velocity and density for a wide range of Hawai‘i island rocks as a prior in a Bayesian regression, with seismic tomography, to refine the 3D density structure for Kīlauea volcano. This refined structure shows dense bodies (3220 kg/m$^3$) between 5 and 8 km below sea level that underly regions of magma storage, found from geodetic and geophysical studies, beneath the summit and East Rift Zone of Kīlauea volcano. Above these bodies, density iso-surfaces surround and cradle sources of pressure change determined from geodetic models, both at the summit and along the East Rift Zone. Continued subsidence of the summit following the 2018 eruption is aligned with a bowl-shaped density structure, formed primarily by density isosurfaces between 2800 and 2900 kg/m$^3$ at 4 to 6 km depth. These surfaces underly the ~3 km depth at which dike injection initiates, are largely aseismic, and from their density values are inferred to contain high concentrations of olivine. Taken together, these density structures are consistent with an olivine-rich mush with variable porosity that increases in density with depth and provides a mechanism to form olivine cumulates both at the summit and along the rift zones. This structural framework for Kīlauea volcano is consistent with melt and mush transport occurring over a large range of depths to accommodate the growth and spreading of the volcano.
Keywords: Gravity anomalies and Earth Structure
Volcano seismology
Magma migration and fragmentation
Physics and chemistry of magma bodies
Introduction

The surface structure of Kīlauea volcano is evident from surface faulting and topography (Denlinger and Okubo, 1995) (Figure 1), from observations of dikes exposed in the summit caldera (Casadevall and Dzurisin, 1987), and from dikes along the East Rift Zone (ERZ) (Montgomery-Brown et al., 2010). The internal structure of Kīlauea volcano is much less evident, and is interpreted from geologic and geophysical measurements (Anderson et al., 2015; Brooks et al., 2006; Flinders et al., 2010; Flinders et al., 2013; Ge et al., 2019; Hildenbrand et al., 1993; Kauahikaua et al., 2000; Lerner et al., 2024; Montgomery-Brown et al., 2015; Montgomery-Brown et al., 2009; Montgomery-Brown et al., 2010; Ryan, 1988; Segall et al., 2019; Shelly and Thelen, 2019; Zucca and Hill, 1980). The summit and ERZ are connected by nearly continuous gravity and magnetic anomalies (Hildenbrand et al., 1993; Kauahikaua et al., 2000). The magnetic anomaly of the summit and ERZ is related to dike structure (Hildenbrand et al., 1993), and the axis formed by dikes overlies a deep zone of magma storage along the ERZ, as shown from the subsidence during the 2018 eruption (Figure 2). Gravity anomalies at the summit and along the rift zone are associated with olivine accumulation within a magma system at depth (Clague and Denlinger, 1994; Flinders et al., 2010; Kauahikaua et al., 2000).

Previously, Denlinger and Flinders (2022) showed that the Nafe-Drake velocity – density relationship, when combined with the seismic tomography results of Lin et al. (2014), produced a density structure that approximated the observed gravity disturbance field on the island. This correspondence was then used to determine the relative importance of gravitational loading and magma pressure on the instability of the island flanks.
Though the modelled gravity field was previously found to be within 90% of the observed field for elevations below 1 km (Denlinger and Flinders, 2022), the densities estimated using the Nafe-Drake relationship are too low. High, mantle-like, compressional velocities of more than 7.2 km/s, beneath the summits of Kīlauea and Mauna Loa volcanoes were associated with a density of only 3027 kg/m³. Yet many field data indicated that higher densities are common: picrites have been emplaced in the summit (Casadevall and Dzurisin, 1987) and also comprised some lava erupted from Kīlauea Iki in 1959 (Helz, 1987). These observations, combined with high seismic velocity structure beneath the summit (Lin et al., 2014), high gravity anomalies over the summit (Kauahikaua et al., 2000), and petrologic evidence from erupted lava support the existence of dense olivine cumulates within the Kīlauea magma system (Clague and Denlinger, 1994). The low densities that Denlinger and Flinders (2022) used were obtained using the global Nafe-Drake velocity-density relationship with a zero bias. Yet in fitting their lab measurements of velocity and density for Hawaiian rocks, using the same Nafe-Drake relationship, Manghnani and Woollard (2013) found it necessary to add a constant bias of 220 kg/m³.

We test the efficacy of a bias in the Nafe-Drake relationship for all Hawaiian rocks using field measurements of gravity disturbances combined with the results of travel-time seismic tomography, as we did previously. However, now we use laboratory measurements between compressional velocity and density as a prior constraint, and determine the Bayesian posterior probability of a positive density bias. This approach produces a more robust assessment of the bias, since the lab-determined bias of 220 kg/m³ presented by Manghnani and Woollard (2013) (or any laboratory measurement) could be subject to a scaling error resulting from...
extrapolating density of small, intact, lab samples, to that of the entire island. Our results sample much larger volumes (by 5 orders of magnitude) and allow us to test the conversion of the seismic tomography structure for Kīlauea volcano into a detailed density structure. The resulting provocative density structure for the summit and ERZ of the volcano forms an insightful and useful structural framework for continued seismic, geodetic, and petrologic studies.

Method

To accommodate a constant bias in the relationship between compressional velocity and density, we generalize the polynomial expansion describing the Nafe-Drake relationship. As given by Brocher (2005), the relationship for density $\rho$ at any subsurface location is obtained from the compressional velocity $v_p$ at that location by the polynomial series

$$\rho(v) = \sum_{i} w_i \phi_i(v)$$

(1)

where the weights $w_i$ are given in Table 1, and the basis functions $\phi_i(v)$ are given by

$$\phi_i(v) = v_p^i$$

(2)

Here $v(x, y, z)$ is the velocity at a point within a three dimensional subsurface domain. In describing their laboratory measurements of the relationship between compressional velocity and density, Manghnani and Woollard (2013) shifted the Nafe-Drake curve shown in Figure 3 upwards by adding a constant bias of 220 kg/m$^3$. This bias is included in equation 1 as

$$\rho(v) = \sum_{0}^{5} w_i \phi_i(v) = w^T \phi(v)$$

(3)
in which the basis function for the bias at $i=0$ is simply unity. We can fit this equation to any set of data using a Bayesian linear regression. Here we determine a prior model for the bias alone, from the lab data of Manghnani and Woollard (2013), assuming the remaining weights are fixed. This determines a prior for the bias and its uncertainty based on lab data. Then we determine the maximum likelihood for the bias using field data. We use a combination of gravity disturbance data and seismic tomography as described in Denlinger and Flinders (2022), with a Gaussian noise model for both sets. We then obtain the maximum posterior for the bias from the product of the lab-determined prior and field-determined maximum likelihood distributions. As they are both Gaussians, their product is also Gaussian.

The distribution of bias values for the laboratory data of Manghnani and Woollard (2013) about their mean is a smoothly varying, symmetric distribution, and the mean value is approximately the mode value, supporting a Gaussian model for the noise distribution. The likelihood of observing a density $\rho$ from $M$ laboratory measurements of the density and velocity is the conditional probability

$$p(\rho | w_0, v, \sigma_0^2) = \prod_{m=1}^{M} \mathcal{N}(\rho_m | w_0, w^T \Phi(v), \sigma_0^2)$$

(4)

in which $\mathcal{N}$ indicates a Gaussian distribution, $\sigma_0^2$ is the variance determined from the scatter in the data, and $w_0$ is a particular value for the bias, and the other weight parameters (the Nafe-Drake distribution) are fixed. The error function $E(w)$ obtained from the log of this expression is

$$E(w) = \frac{1}{2} \sum_{m=1}^{M} \left[ \rho_m - \sum_{j=1}^{5} w_j \phi_j \right]^2$$

(5)
Minimizing $E(w)$ with respect to the bias $w_0$ (Bishop, 2009) gives a value for the mean bias $\overline{w_0}$

$$\overline{w_0} = \overline{\rho} - \sum_{i=1}^{s} w_i \overline{\phi}_i$$  \hspace{1cm} (6)

in which

$$\overline{\rho} = \frac{1}{N} \sum_{n}^{N} \rho_n$$  \hspace{1cm} (7)

and

$$\overline{\phi} = \frac{1}{N} \sum_{p}^{N} \phi^i_p$$  \hspace{1cm} (8)

These values are used as a prior for the bias value $w_0$ in a Bayesian analysis of the field data, and its normal distribution is written as

$$p(w_0) = \mathcal{N}(w_0 \mid \overline{w_0}, \beta_0)$$  \hspace{1cm} (9)

in which for an uninformative prior for the bias values, the inverse of the variance $\beta_0$ is estimated by the mean squared difference between the Manghnani line and the lab data in Figure 3.

In determining the constraints from the observed gravitational field we use the gravity disturbance values calculated as described and explained in Denlinger and Flinders (2022) and in Damiani (2013). The gravity disturbances used here are the differences between the observed gravity field at a point minus the normal gravity of the WGS84 geoid that passes through that point. Additionally, these disturbances have been corrected for the flexure of the lithosphere, as described in Denlinger and Flinders (2022). The result is a gravity disturbance field due solely to the three-dimensional distribution of mass in the island edifice, whether the
mass exists above or below the measurement point. This is not a Bouguer anomaly, which subtracts an arbitrary value from a measured field to obtain a residual and in which it is often assumed that all density bodies lie below the measurement point.

The likelihood used here is the probability that we will observe a gravity disturbance value \( g_n \) at a surface location \( x(x, y, z) \) resulting from a velocity-determined density distribution of the edifice. The likelihood function for the gravity field data \( d(x, y, z) \), obtained from surface gravity measurements with an assumed Gaussian noise distribution, is the product of conditional Gaussians, one for each field point

\[
p(d | v, w, \beta) = \prod_{n=1}^{N} \mathcal{N}(g_n | f_n \{w^T\phi(v)\}, \beta^{-1})
\]

Here \( N \) is the number of surface field points at which we determine the gravity disturbance for the edifice at that point, using the method presented in Denlinger and Flinders (2022). The term \( f_n \{w^T\phi(v)\} \) is the function that determines the surface gravity disturbance at location \( n \) from a hexahedral prism whose centroid has a velocity \( v_p \) at some subsurface location in the edifice.

The model gravity is subject to the subsurface estimate of velocity-determined density, which has a Gaussian uncertainty with precision (inverse variance) \( \beta \). The error function \( E_D(w) \) is obtained from taking the log of equation 11 and has a structure similar to equation 5, and

\[
E_D(w) = \frac{1}{2} \sum_{n=1}^{N} \left[ g_n - f_n \{w^T\phi(v)\} \right]^2
\]

The density of each subsurface hexahedral prism contributing to the surface gravity disturbance field is determined from its average compressional velocity, using the 3D seismic tomography structure of Lin et al. (2014) and equation 5 for any constant bias value \( w_0 \). The mesh used for
this regression is described in Denlinger and Flinders (2022). Minimizing this error function for each hexahedral prism contributing to the gravity disturbance at the surface point (for the entire island edifice) gives the maximum likelihood value for the bias from all the field data. As will be shown below, since the number of field measurements $N$ greatly outnumbers our single parameter, then the inverse variance is approximately given by

$$\beta = \frac{N}{2E_D(w)}$$  \hspace{1cm} (12)

The prior is given by $p(w_0)$ (equation 9) and combined with the likelihood determines the unnormalized Bayesian posterior for the bias $w_0$ as

$$p(w_0 | x, w, \sigma) \propto p(d | x, w, \sigma)p(w_0)$$  \hspace{1cm} (13)

By completing the square in the exponentials in this product (Bishop, 2006, p. 93), and normalizing this Gaussian, the product on the right-hand side can be written as (Bishop, 2006, p. 153),

$$p(w_0 | d) = \mathcal{N}(w_0 | m_N, S_N)$$  \hspace{1cm} (14)

where

$$m_N = S_N (S_0^{-1} m_0 + \beta \phi^T d)$$

$$S_N^{-1} = S_0^{-1} + \beta \phi^T \phi$$  \hspace{1cm} (15)

Here $m$ refers to the mean (and mode) of the distribution and $S$ refers to the covariance. The subscript 0 refers to the prior lab data, whereas $N$ refers to the posterior for bias $w_0$ using $N$ field points. These statistics are a quantitative measure in the uncertainty in our density structure at depth, as the variations in bias $w_0$ determine the variations of density for each value of velocity at a point. Equations (14) and (15) are a general result for a range of possible
weight bias values and Gaussian errors in data and measurements. For this problem here, we
have held fixed all of the weights given by Brocher (2005), and only allow the bias to vary. In
this case, for the posterior distribution over the single variable for bias \( w_0 \), equations 14 and
15 are reduced to:

\[
p(w_0 | m_N, \sigma_N^2) = \mathcal{N}(w_0 | m_N, \sigma_N^2) \tag{16}
\]
in which

\[
m_N = \frac{\sigma^2}{N\sigma_0^2 + \sigma^2} m_0 + \frac{N\sigma_0^2}{N\sigma_0^2 + \sigma^2} m
\]
\[
\frac{1}{\sigma_N^2} = \frac{1}{\sigma_0^2} + \frac{N}{\sigma^2} \tag{17}
\]

Here \( \sigma^2 = \beta^{-1} \) and \( \sigma_0^2 = \beta_0^{-1} \) are the variances in the fit to the gravity disturbances and the fit
to the prior lab data, respectively. The uncertainty in density is directly given by the uncertainty
in the maximum in the posterior for bias \( w_0 \). The probable variation in density around the
maximum posterior used in the remainder of this paper is given by the variance for the bias
listed in Tables 1 and 2.

Consistency in these results, and therefore this posterior for the bias, depends upon
where sampling is done. Given that the seismic constraints on the subsurface were confined to
seismometers on land (Lin et al., 2014), the constraints for velocity in the subsurface within the
volcanic edifice, e.g. above decollement seismicity, are strongest beneath land rather than the
seafloor. In Figure 4a the gravity disturbance field determined by Denlinger and Flinders (2022)
is shown, along with three areas sampled to determine the posterior probability of the constant
bias \( w_0 \). The corresponding normalized posterior distributions, obtained from the product of
the likelihood for that area and the prior Gaussian determined from the lab data, are shown in
Figure 4b. The largest area A coincides with the lateral margins of the grid used by Lin et al. (2014) for seismic tomography. The next largest area B includes the summits of both Mauna Loa and Mauna Kea volcanoes. The smallest area C is confined largely to Kīlauea volcano, and is less than half of the area of B. Nevertheless, both B and C have the same maximum posterior for the bias parameter $w_0$. These two smaller areas are primarily confined to land, and there the tomographic velocity structure is much better constrained than it is offshore where no seismometers were deployed. The values for the mean and variance of the posterior for the bias parameter for Area C is listed in Table 2. The resulting equation, formed from the weights in Table 1 and the posterior bias in Table 2, form the basis for a new density structure for Kīlauea volcano. The gravity disturbance residuals obtained by applying the maximum posterior bias of 170 kg/m$^3$ to convert velocity to density are shown in Figure 5. As shown in Figure 4b, the error obtained from this bias is significantly better than the value of 220 kg/m$^3$ that Manghnani and Woollard (2013) used, and a marked improvement over the zero bias initially used by Denlinger and Flinders (2022) that is shown by the green curve in Figure 4b.
Discussion

The new velocity-derived density structure significantly increases densities within the previously-determined density structure of Denlinger and Flinders (2022) and provides a new structural framework for Kīlauea volcano. Shown in Figure 6 is a view of dense 3D isosurfaces within Kīlauea volcano and in the underlying oceanic lithosphere that have the density of dunite (3220 kg/m$^3$), and these are combined with the relocated seismicity (Lin et al., 2014) associated with either fault movement or magma transport into the magma system. The seismic hypocenters, shown as black dots, form distinct spatial groups identified by letters in this figure. Letter A is left lateral strike slip seismicity along the upper East Rift Zone of Kīlauea (Gillard et al., 1996), and flanks the east side of the south caldera magma source interpreted from geodetic studies (Poland et al., 2012). Letter B is right lateral strike slip and wrench fault seismicity occupying the west half of the Koae desert (Lin and Okubo, 2016). Seismicity labeled C is produced largely by dike injection within the summit caldera. Letter D is right lateral plus extensional faulting induced by the S20°E motion of the south flank of Kīlauea as it slides into the sea (Denlinger and Okubo, 1995; Owen et al., 2000; Wyss et al., 2012). Letter E is the path of melt migration through the oceanic lithosphere and into Kīlauea volcano, as illuminated by the earthquakes that melt migration induces (Klein et al., 1987). The track of this seismicity suggests that melt migrates upward into and through the large, dense (3220 kg/m$^3$ isosurface), body underlying the nearly aseismic region between the seismicity C and that of A and B. This aseismic summit region contains the sources of pressure change that characterize summit eruptive activity and summit surface deformation on Kīlauea volcano (Anderson et al., 2015; Wang et al., 2021).
Within the aseismic summit region in Figure 6, there are density structures above -5 km elevation and below -2.5 km elevation that cradle the magma system. To illustrate this, we show the dense bodies in Figure 6 in map view in Figure 7a, along with topography and epicenters of seismicity, and give the orientation of the vertical SW-NE cross section in Figure 7b. This vertical slice is constructed to bisect the Halema’uma’u (HMM) collapse, the deeper seismicity associated with magma transport into the base of the volcano, and the shallow HMM magma source of Wang et al. (2021). This section is viewed looking north in Figure 7b, and shows that density isosurfaces between densities of 2600 and 3000 kg/m$^3$ near HMM form a cupola at shallow depths that is contained within seismicity C, the aseismic region, and then inverts for form a trough as depth increases. The cupola protruding into the shallow layers envelopes the ellipsoidal HMM pressure sources of both Wang et al. (2021) and Anderson et al., (2019), and at 2600 kg/m$^3$ is significantly denser than the topmost km of basalt (Keller et al., 1979). For reference to Figure 6, we show the hypocenter distributions of earthquakes projected onto this plane, labeled with the same labels as in Figure 6. In the foreground of the Figure 7b cross section, strike slip seismicity is plotted that flanks the deeper trough, and is consistent with melt and mush flow south out of the summit region.

In three dimensions, the density isosurfaces produced by the contours shown in the cross section of Figure 7b form a definitive structure at the summit of Kilauea volcano. In Figure 8 a suite of these surfaces is shown relative to the continued subsidence of the summit immediately following the cessation of the massive 2018 eruption. For three months, melt and mush continued to flow out of the summit area and into the East Rift Zone, which underwent concomitant uplift along the rift just east of Pu‘u‘ō‘ō (Wang et al., 2021). Deflation of the south
summit area, considered to be the main storage area for melt and mush at the summit (Anderson and Poland, 2016), has a surface expression with a curvature and with inflections that coincide with a range of density isosurfaces between 2800 and 2900 kg/m$^3$ (Figure 8). In particular, the sharp inflection in subsidence contours and the shape of the depression south and west of the summit are mimicked by the tomography-determined density structure at depth. The closest correspondence is with a density of 2880 kg/m$^3$, as shown in Figure 8.

The density isosurface that most closely corresponds to the post-eruption Aug-Nov 2018 summit subsidence contours (2880 kg/m$^3$) also has structure that aligns with summit seismicity, with known vents along the East Rift Zone, and with the subsidence observed along the East Rift Zone during the large 2018 eruption described by Neal et al. (2019). The correspondence of this density isosurface with summit and Koae seismicity is shown in Figure 9, along with line contours of topography. The trough underlying shallower magma sources in the summit opens to the south, as shown by the flooded contour surface here, and this opening is flanked by two distributions of strike slip earthquakes, one left lateral along the upper East Rift Zone, and one right lateral extending south from the crater area as shown (the latter distribution is wrench faulting (Mandl, 1988) as it also includes extensional mechanisms). This pattern is also reflected in the deformation of the Koae fault system over hundreds of years (Duffield, 1975). The strike slip seismicity is likely induced by flow of melt and crystals south out of the summit towards the East Rift Zone. Along the East Rift Zone, elevated ‘hills’ in this density isosurface represent higher densities, and prominent hills and ridges underlie known vents at Makaopuhi, Pu‘u‘ō‘ō, and Heiheiahulu along the East Rift Zone. In Figure 10, this 2880 kg/m$^3$ isosurface is shown with the underlying dense 3220 kg/m$^3$ isosurfaces forming the structures shown in Figure 6. These
are associated with magma storage in the summit, and we infer that the same is true along the East Rift Zone. Along the portion of the 2880 kg/m$^3$ surface in the East Rift Zone, there is a sinuous channel formed by the elevations in this surface which also corresponds to the subsidence observed during the 2018 eruption, as shown in Figure 11.

All of the isosurfaces between 2800 and 2900 kg/m$^3$ form an elevated, roughly triangular-shaped platform formed by the plunge of these isosurfaces to depths of 6 km or more along the platform margins (Figure 9). This plunge in the elevation of these isodensity surfaces indicates a substantial reduction in density surrounding this platform. To the south, these soft and less dense areas closely correspond to the arcuate boundaries of the Hilina Pali fault system (Figure 9). These reduced densities around the margins of Kīlauea volcano may represent a reduction in density produced by hydrothermal alteration, as these anomalies flank the active, hotter, subsurface areas of Kīlauea volcano. If so, then the Hilina Pali faults are rooted in weak, hydrothermally altered material formed by hydrothermal circulation flanking the existing magma system, and their formation is exacerbated by S20E extension of the south flank without being directly connected to decollement slip (Denlinger and Morgan, 2014).

Combined with other geophysical and geodetic data these observations form a structural framework for Kīlauea volcano. The accumulations of dense bodies with the density of dunite are likely to be accumulations of olivine that are forming dunite in areas of long-term melt and mush accumulation within Kīlauea’s magma system (Lerner et al., 2024). These dense bodies are interpreted here to be overlain by a dense mixture of melt and crystals that feed dikes, both at the summit and along the East Rift Zone as shown in Figure 12a. The density structure of the mush (Figure 7b) is consistent with concentration of olivine crystals increasing
with depth, and this gradation in density along the East Rift Zone is supported by the May-June 1840 eruption along the East Rift Zone, where crystal poor lava was erupted near Makaopuhi and picrite with cm-size olivine crystals was erupted just uprift of Cape Kumukahi (Macdonald, 1944). It is also the structure of spreading ridges that form in back-arc basins (Artemieva, 2023; Turner et al., 1999; Veloso et al., 2014). It is also the structure proposed by Walker (1986) in his analysis of the Koolau dike complex on Oahu. It is the correspondence of Hawaiian ridge structure and growth with oceanic and spreading ridge structure that is the basis for Figure 12.

At the summit, our data density is sufficient to indicate that the density and thus crystal concentration increases nonlinearly with depth, eventually forming the dunite bodies that underly this region of magma storage within the volcano. Here, and along the East Rift Zone, a combination of variations in melt concentration and the separation of the flow of melt from a crystalline mush to feed dikes and lava flows can produce complex mechanical behavior (Bessat et al., 2022; Keller and Suckale, 2019; Suckale et al., 2016). This applies to melt transport during periods between eruptions as well as during eruptive activity. Though these complex behaviors are beyond the scope of this study, they provide numerous opportunities for future research on this system.
Conclusions

The relationship between compressional velocity and density as given by the Nafe-Drake relationship (Brocher, 2005) requires a constant bias to explain joint measurements of seismic velocity and density for Hawaiian rocks (Manghnani and Woollard, 2013). Allowing for a constant bias, the 3D seismic tomography structure can be converted to a 3D density structure for the island of Hawaiʻi, and produces a surface gravity disturbance field that more accurately reproduces the observed gravity disturbance field on the island than was done previously. Using the laboratory measurements as a prior, a linear Bayesian regression of thousands of gravity disturbance field points on land and at sea determines a maximum posterior for this bias of 170 kg/m$^3$. This constrains the 3D density structure of the island, and particularly of Kīlauea volcano.

The new density structure defines dense bodies at shallow levels at the summit and along the East Rift Zone of Kīlauea volcano that have the density of dunite (3220 kg/m$^3$). These dense bodies underlie subsurface regions that geodetic and geophysical studies routinely identify as volumes of pressure change that produce observed surface deformation, and therefore are correlated with magma storage. However, we show here that the density of these regions is greater than the density of a crystal-free tholeiitic basaltic melt, and here the volumes where pressure change occurs are interpreted as a melt-rich layer overlying a crystal rich mush. The bulk density increases nonlinearly over a depth range of three to four km, culminating in bodies with the density of dunite (>3220 kg/m$^3$). At the summit, density isosurfaces between 2750 and 3000 kg/m$^3$ form a horse-collar shaped trough that opens to the south and cradles the magma volumes associated with surface deformation during volcanic
unrest. In map view, contours of this trough surround both the LIDAR-determined 2018 summit collapse (Mosbrucker et al., 2020) and also aligned with the continued subsidence of the south caldera region after the end of the 2018 eruption. On its exit from the crater, the south trending 2880 kg/m$^3$ density isosurface forms a trough that is flanked by strike slip seismicity. Focal mechanisms of this seismicity are consistent with the southern flow of a mush mixture south and east out of the crater and into the East Rift Zone. Along the East Rift Zone, the 2880 kg/m$^3$ isosurface has hills that underlie previously active vents at Makaopuhi, Pu‘u‘ō‘ō, and Heiheiahulu. The axis of a sinuous trough snaking between these hills aligns with InSAR-determined East Rift Zone subsidence, and deflection of that subsidence, during the 2018 eruption. In addition, as these isodensity surfaces between 2800 and 2900 kg/m$^3$ correlate to the volcanic activity, the elevated, roughly triangular shaped region within Kīlauea volcano that they form is interpreted to be at elevated temperatures. Surrounding this region, density isosurfaces plunge to depths of 6 km or more and reflect areas of low density surrounding the platform. In particular, these low-density volumes are aligned with the Hilina Pali fault system and indicate that this fault system is underlain by low density rock. If the density here is produced by hydrothermal alteration, then the Hilina Pali fault system is there because that part of the flank and those faults are rooted in hydrothermally altered material. Taken together, the combination of gravity, seismic velocity, earthquake hypocenters, seismic focal mechanisms, and surface deformation induced by volcanic activity show that these derived density structures form a useful skeletal framework for the structure and study of Kīlauea volcano.
Table 1. New coefficients in the modified Nafe-Drake velocity-density equation

| $\bar{w}_0 |_{\text{prior}}$ | $w_0 |_{\text{posterior}}$ | $w_1$ | $w_2$ | $w_3$ | $w_4$ | $w_5$ |
|----------------|-----------------|-----|-----|-----|-----|-----|
| 278 | 170 | 1661.2 | -472.1 | 67.1 | -4.3 | 0.106 |

Table 2. Sufficient statistics for prior and posterior Gaussians for bias

<table>
<thead>
<tr>
<th>Mean prior $\bar{w}_0$</th>
<th>Variance prior $\sigma_0^2$</th>
<th>Mean bias $\bar{w}_0$</th>
<th>Variance $\sigma^2$</th>
</tr>
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<tbody>
<tr>
<td>278.</td>
<td>158$^2$</td>
<td>170</td>
<td>22$^2$</td>
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</table>
Figure 1. Surface structure of the southeast portion of the island of Hawai‘i, adapted from Denlinger and Okubo (1995). The East Rift Zone of Kīlauea volcano, shown as a series of red dashes to highlight the wrench fault structure, heads southeast from Kīlauea summit before turning east, then northeastward and continuing eastward offshore as the prominent submarine ridge shown here. The motion of the south flank of Kīlauea is about S20°E, and subparallel to the S50°E strike of the right lateral slip fault that forms its western margin with Kama‘ehuakanaloa (formerly Lō‘ihi). Separation of Kīlauea from Mauna Loa causes wrench faulting on adjacent Mauna Loa, consisting of both extension (shown) and right lateral strike slip. The large prominent fault scarps near the coast south of the summit are the Hilina Pali fault system, and we constrain the low density roots of this system in this report.
Figure 2. Geologic and geophysical data support the continuity of the magma system from the summit into the East Rift Zone (ERZ) on the Big Island of Hawai‘i (Flinders et al., 2013; Hildenbrand et al., 1993; Kauahikaua et al., 2000). Shown here is the total magnetic field reduced to the pole (RTP), in which the nearly continuous magnetic high from the summit and out along the ERZ records the magnetic field of dikes emplaced at the summit and along the rift zone that geodesy studies show originate from sources below 3 km depth along the rift zone (Lundgren et al., 2013). Gravity data form a similar pattern, and this is a ubiquitous feature of the summits and rift zones of Hawaiian volcanoes that likely reflects the accumulation of olivine beneath regions of magma storage as the volcano grows (Clague and Denlinger, 1994; D.A. Clague and Sherrod, 2014). Also shown is the unwrapped InSAR-derived subsidence (descending sentinel data May 5 to Aug 9, 2018; de=0.545, dn=0.118, dz=0.830) during extraction of 1.5 km$^3$ DRE of melt from Kīlauea’s magma system (Dietterich et al., 2021).
Figure 3. Laboratory measurements for mafic rocks (Birch, 1961; Christensen, 1966; Hyndman and Drury, 1976; Manghnani and Woollard, 2013) are used to convert compressional seismic velocity to rock density in order to interpret the density structure of the Earth’s crust and mantle. Here some of these data are combined with similar measurements of mafic Hawaiian rocks. The purple curve was determined by Manghnani and Woollard (2013) to provide a least squares fit to lab data for Hawai‘i, and is the velocity-density relationship of Nafe-Drake plus a constant, velocity-independent bias of 220 kg/m$^3$. In this paper we determine that that bias for gravity and seismic tomography data island-wide is 170 kg/m$^3$. 
Figure 4a. Three different areas used to test the efficacy of the bias generalizing the velocity-density relationship for Hawaiian rocks. The area A is what was used in Denlinger and Flinders (2022), whereas the areas B and C are confined mostly to subaerial locations best constrained by the seismic networks used to do tomography (Lin et al., 2014). The statistics for A, B and C (see text) are listed in Tables 1 and 2.
Figure 4b. The posterior distribution for the bias $W_0$ determined for each area in Figure 4a. As discussed in the text, the consistency of the posterior for land areas B and C, despite large differences in the area of these two domains (see inset), is a robust result. For Kilauea volcano (curve C), the standard deviation in the maximum posterior distribution for the bias is 22 kg/m$^3$ (Table 2).
Figure 5. The distribution of the residual gravity disturbance in which the calculated values used the seismic tomography of Lin et al. (2014) to construct the 3D domain of Denlinger and Flinders (2022). The linear regression used the velocity-density relationship in equation 3 with the values in Tables 1 and 2, and the maximum posterior bias $w_i$ shown in Figure 4b and given in Table 2. Observed gravity disturbances at each surface point result only from the island edifice structure, as described in Denlinger and Flinders (2022), and are modeled with a 3D model of the island edifice to produce the residuals shown here.
Figure 6. The density structural model derived from seismic tomography (Lin et al., 2014) and the gravity disturbance field (Denlinger and Flinders, 2022) produces a structural framework for Kīlauea volcano in which dense bodies underlie the locations of magma storage inferred from geodetic studies of eruption-related surface deformation. This is true both at the summit, and along the East Rift Zone beneath Makaopuhi (not shown), Pu‘u‘ō‘ō, and Heiheiahulu. Here, decollement seismicity defines the base of the edifice sliding into the sea, and the other letters indicate persistent seismic patterns. Letter A is left lateral seismicity (Gillard et al., 1996) and letter B is right lateral and wrench faulting seismicity (Lin and Okubo, 2016) that flanks the location of a geodetically-determined south caldera magma source (Wang et al., 2021), on the east and west sides, respectively. The letter C is shallow summit seismicity between 0.5 and -2 km elevation. The letter D is right lateral wrench faulting on Mauna Loa, produced as Kīlauea slides SE into the sea (Wyss et al., 2012). Letter E is generated by magma movement from the mantle, through the oceanic lithosphere and into the volcano, passing through the dense body underlying the summit region. The 2018 fissures are where 1.5 km$^3$ of melt drained in 60 days from the volcano in 2018 (Dietterich et al., 2021), causing a caldera collapse at the summit.
**Figure 7a.** Map view of dense bodies in Figure 6, with seismicity and orientation of vertical cross section in Figure 7b. The cross section intersects isosurfaces that define the summit density structure above the dense 3220 kg/m$^3$ bodies shown here, and these form the trough that is flanked by strike slip seismicity in Figure 7b.
**Figure 7b.** Vertical cross section through seismic tomography derived density structure that intersects the summit 3D body and seismicity of C shown in Figure 6. Seismicity outside this plane are projected horizontally onto it. This section shows a trough beneath Halema'uma'u at Kīlauea's summit that often produces sources of inflation and deflation. These sources are inferred to be a mixture of melt and crystals, and their flow south out of the summit region are consistent with distribution of out-of-plane strike slip seismicity shown here. The 3D shape of this trough for various densities is compared to InSAR derived summit subsidence in Figure 8.
Figure 8. Map view of density isosurfaces within the aseismic region overlying the dense bodies in Figure 6. Black line contours show InSAR subsidence as summit continued to drain from Aug 9, 2018 to Nov 7 2018. In addition, red dots are earthquake hypocenters above the surface, and purple dots are earthquake hypocenters below the surface. Flooded contours show a suite of these surfaces relative to both seismicity and to the InSAR-derived subsidence in the summit (unwrapped descending sentinel data August 9 to November 7, 2018 with look angle components $de=0.545$, $dn=0.118$, $dz=0.830$) as the summit drained into the East Rift Zone from August to November 2018. Close correspondence between bowl-shaped density isosurfaces and these contours in InSAR-derived vertical displacement in the south summit area occurs for densities between 2880 and 2840 kg/m$^3$. Melt density is about 2600 kg/m$^3$. 
Figure 9. Map view of the topography and seismicity associated with the 2880 kg/m$^3$ density isosurface that most closely aligns with the shape of south summit subsidence as melt and mush flowed out of the summit from August 9 to November 7, 2018 (Figure 8). This flow of melt is consistent with observed seismic focal mechanisms of strike slip seismicity flanking this trough as shown here. Elevated ‘hills’ in this density isosurface are associated with past eruptive vents along the East Rift Zone, such as near Makaopuhi, Pu‘u‘O‘o, and Heiheiahulu. The roughly triangular region formed by these elevated areas (shown here as yellow and warmer colors), is bounded by areas shown in blue where this isosurface plunges to depths of 6 km or more. The low density of the edifice may reflect hydrothermal circulation surrounding this hot region. These regions of low density align with the Hilina Pali fault zone, defined by the prominent, arcuate scarps in this figure, and if generated by hydrothermal alteration indicate that these fault scarps are rooted in weak hydrothermally-altered material.
Figure 10. Map view of the relationship of the 2880 kg/m$^3$ isosurface shown in Figures 8 and 9 with the dense bodies shown in Figures 6 and 7a. Green lines are topographic contours with a 25m interval. Dots are earthquake epicenters, with red above the surface and purple below. Darker isosurfaces enclose high densities below flooded contour surface. The density of these bodies (3220 kg/m$^3$) is consistent with dunite and is likely an olivine cumulate derived from present or past magma storage within the volcano. These bodies, also mapped by magnetotellurics (Hoversten et al., 2022), are largest under the summit, and are consistent with picrites found in the summit region (Casadevall and Dzurisin, 1987). This structure, through which the magma entering the volcano passes (Figure 6), may be a common byproduct of the growth of Hawaiian volcanoes (Clague and Denlinger, 1994; D.A. Clague and Sherrod, 2014). It provides a mechanism for the eruption of golf ball-size nodules of dunite coated in basalt that are found on rift zones of other volcanoes on the island of Hawai‘i (Chen et al., 1992).
Figure 11. Map view of the relationship of the 2880 kg/m$^3$ isosurface shown in Figures 8 and 9 with the large subsidence (grey contours) along the rift zone observed during drainage of 1.5 km$^3$ of magma (Dietterich et al., 2021) from Kīlauea volcano in 2018 (unwrapped InSAR-derived subsidence from descending sentinel data May 5 to Aug 9, 2018; de=0.545, dn=0.118, dz=0.830). Black contours show lidar-derived subsidence May 5 to Aug 4, 2018 in the summit area near Halema‘uma‘u. The ‘horse collar’ shape formed by this density isosurface in the summit region, showing higher densities surrounding the active summit caldera, bounds the massive caldera collapse associated with the withdrawal in 2018 that was recorded with lidar, shown in black here overlying summit seismicity in red. The western and southern margins of this horse collar shape bound the limits of InSAR determined subsidence in the summit area. The ‘hill’ in this surface near Makaopuhi is associated with a deflection in the subsidence mentioned above during 1.5 km$^3$ of magma withdrawal in 2018.
**Figure 12a.** Combining geophysical, geologic, and seismic studies of Kīlauea volcano with the information gleaned here for density, a consistent structural model for the magma system emerges. The structure is similar to the core of spreading ridges that form in back-arc basins (Nicolas and Boudier, 1991), possibly as a result of the persistent flank motion that produces flank spreading across the East Rift Zone as it migrates seaward, emulating a spreading ridge. Geodetic studies (Lundgren et al., 2013; Montgomery-Brown et al., 2015; Montgomery-Brown et al., 2011; Montgomery-Brown et al., 2010) show that dikes appear to spontaneously arise from a depth of 3 to 3.5 km along the ERZ, as shown, and this may simply be a consequence of the presence of extensive, underlying mush bodies that feed them.
Figure 12b. The precipitation and consolidation of olivine crystals form distinct volumes of mush, and produces the dense bodies (labeled ‘dunite’) that we infer from the combination of seismic tomography, gravity, and lab measurements. Variations in melt concentration within this mush are indicated by non-linear increases with depth, both beneath the summit and along the East Rift Zone, as well as by the near simultaneous eruption of crystal poor magma near Makaopuhi and picrite at the east end of the East Rift Zone (Macdonald, 1944). Such variations in melt/mush concentration can produce complex mechanical behavior (Bessat et al., 2022; Keller and Suckale, 2019; Suckale et al., 2016), particularly in the coupling of the summit to the East Rift Zone during eruptive activity. The potential for a rich set of complex mechanical behaviors resulting from structures like this present abundant opportunities for future research on this system.
The gravity data underlying this article are available in Flinders et al. (2013), at doi:10.1002/grl.50633 and in the USGS Science Base at doi.org/10.5066/F7V1230Q. European Space Agency Sentinel 1 InSAR data are available from Alaska Satellite Facility’s data repository (https://asf.alaska.edu/data-sets/derived-data-sets/insar/).

References


