Seismic bulk and shear attenuation along a transect from Kama‘ehuakanaloa volcano through Mauna Loa to the Aloha Cabled Observatory:

Implications for the distribution of partial melt

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Key points
• Bulk attenuation exceeds shear attenuation for all Hawai‘i earthquakes observed
• A thermodynamic equilibrium model of partial melt effectively fits a bulk attenuation data set for Mauna Loa and Kama‘ehuakanaloa volcanos
• Aloha Cabled Observatory successfully recorded all $Q_K$ and $Q_S$ data from Mauna Loa and Kama‘ehuakanaloa volcanos using a single hydrophone
ABSTRACT

Bulk ($Q_K$) and shear ($Q_S$) attenuation are measured and modeled to ~50 km depth beneath Hawai’i. High-frequency (>50 Hz) earthquakes are routinely observed from the Aloha Cabled Observatory (ACO) along the azimuth to Mauna Loa, Pāhala, and Kama‘ehuakanaloa volcano. Bulk attenuation is consistently larger than shear attenuation beneath Hawai’i at frequencies >2 Hz. The Mauna Loa Summit shows the smallest $Q$ values, and transects approaching the Summit from the southeast differ asymmetrically with those departing to the northwest from the Summit. Transect maps of $Q$ are created from the measurements to present in plan view the distribution of $Q_K$ and $Q_S$ near the moho. Activation energy $E^*$ models of $Q_S$ are tested both at Pāhala and Kama‘ehuakanaloa for experimentally determined olivine $E^*$ using the temperature derived from a Hawai‘i Hotspot geotherm and pressure. The $Q_K$ arising from water-filled pores in vesicular basalts within the shallow oceanic crust are a hypothesized mechanism for bulk attenuation measured in the shallow crust near ACO and Wake Island. Below the shallow oceanic crust, partial melt presents a feasible bulk attenuation mechanism at volcanos. Fitting a thermodynamic equilibrium model for frequencies >1 Hz to the $Q_K$ measurements shows a very good match to the $Q_K$ data, predicting partial melt fractions of 0.1% to 10%. Translating the $Q$ maps into partial melt regions near Mauna Loa, Pāhala, and Kama‘ehuakanaloa volcano gives a first view of the observation, location, and distribution of partial melt along the ~100 km transect from southeast to northwest of Mauna Loa.
Plain Language Summary

Attenuation of seismic energy in Hawai‘i has two separate mechanisms: bulk compression/decompression and shearing the rock. These attenuation properties are measured from earthquakes along a line from the Aloha Cabled Observatory (ACO) to Mauna Loa, and southeast to Pāhala, and Kama‘ehuakanaloa volcano. These earthquakes occur down to ~50 km depth, and are rich in high frequency energy. Bulk attenuation is uniformly larger than shear attenuation in Hawai‘i. The Mauna Loa Summit shows the largest attenuation observed. Traversing the Summit from the southeast, attenuation is not symmetrical with paths traversing to the northwest from the Summit. For shallow and deep earthquakes, experimentally determined olivine mineral properties compare successfully with observed shear attenuation measurements. Partial melting of rock at a boundary in contact with magma presents a feasible bulk attenuation mechanism at volcanos. Fitting a partial melting model to the bulk attenuation measurements shows a very good match to the data, predicting partial melt fractions of ~1% volume. Translating the bulk attenuation maps into partial melt regions near Mauna Loa gives a first view of the location and distribution of partial melt along the ~100 km traverse from southeast to northwest of Mauna Loa.

Index terms [3050, 5144, 7280, 3619, 3909]

Keywords [bulk attenuation Mauna Loa partial melt]
1 INTRODUCTION

This study was first motivated by the swarm of 50 $M_w$ 5+ earthquakes which occurred during the collapse of the Kīlauea caldera in 2018 (e.g., Butler 2019, Neal et al., 2019). Stacking the Kīlauea data at seismic stations situated along the azimuth to the Aloha Cabled Observatory (ACO), Butler (2020) derived effective $Q (Q_{eff})$ of bulk and shear attenuation for the paths beneath the volcanos and within the oceanic crust (Figure 1). In addition to the frequency dependent effective $Q_{eff}$ values, shear attenuation activation energy, $E^*$, for the shallow basaltic crust was derived. Except for the nearest Kīlauea site, where $Q_S \approx Q_K$ at < 10 Hz, the measured $Q_S > Q_K$ for the Kīlauea data, and a mechanism for the observed bulk attenuation was hypothesized. This paper and its predecessor further serve to acknowledge and highlight the remarkable scientific capabilities of cabled seafloor observatories, even when employing only a single high-frequency seafloor hydrophone.

A swarm of 16 $M_L \gtrsim 4$ deep earthquakes (>30 km) in 2020–2023 near Pāhala on the southeast coast of Hawai‘i Island were reported by the U.S. Geological Survey – Hawaiian Volcano Observatory (USGS-HVO). These earthquakes offered a similar geometric arrangement transecting Mauna Loa to that of the Butler (2020) Kīlauea study. Earthquake spectrograms were reviewed for the Pāhala events, and other earthquakes along the azimuths between ACO and Mauna Loa, Pāhala, & Kama‘ehuakanaloa (Figure A1).

The narrow study corridor is presented in Figure 1 [Left] extending from ACO to Mauna Loa, Kama‘ehuakanaloa, and Pāhala, and from ACO to Kīlauea. Figure 1 [Right] shows the epicentral locations of earthquake in the data set. The mean, very shallow, hypocentral depths for the Kīlauea and Mauna Loa caldron earthquakes are 0.3 and -0.2 km, respectively, and hence broadly analogous. The paths have comparable propagation distances and azimuths to ACO, and
the propagation paths merge together approaching ACO. As the propagation path from the
Pāhala events transited the crust and below the moho beneath the caldera of Mauna Loa, the
swarm affords a comparison of attenuation from the 2018 Kīlauea data. Fortuitously, the azimuth
between ACO and Pāhala lay within 1° of the azimuth to Kamaʻehuakanaloa volcano (formerly,
Lōʻihi). The Kamaʻehuakanaloa earthquakes (at depths near 11 km and at 46 km) also exhibited
high frequency, deep S-waves ~55 Hz (Figure A1).

In Figure 2 [Left], I present the initial measurements and comparison of $Q_K$ and $Q_S$ for Mauna
Loa and Kīlauea calderas, following the methodology of Butler (2020). For both volcanos, $Q_S >$
$Q_K$, i.e., bulk attenuation exceeds that for shear attenuation. Moreover, along the path to ACO,
the Mauna Loa paths are more attenuating than observed for Kīlauea. The higher frequency
content of the Kīlauea $Q$ as compared to Mauna Loa may be attributed to larger earthquake
sources in the Kīlauea caldera (Kīlauea Mw ~ 5.3 and Mauna Loa $M_L$ ~ 4.2) and lower signal-to-
noise (SNR). The effective $Q$ measured at ACO from earthquakes southeast of Mauna Loa and
transiting the summit is plotted from both shallow and deep events (Figure 2 [Right]), see
caption).

Within the MaunaLoa – Kamaʻehuakanaloa volcano edifices I use the multiple earthquakes to
separate and identify $Q_K$ and $Q_S$ as a function of frequency for subdivided pathways extending
from Mauna Loa–Pāhala–Kamaʻehuakanaloa to ACO. Since $Q_S$ is an exponentially activated
process, I confirm that the mapping of experimentally derived activation energies $E^*$ for olivine
using a Hawaiʻi Hotspot geotherm (Lee et al., 2009) and pressure (PREM, Dziewonski and
Anderson, 1981) can appropriately match the change in $Q_S$ between shallow and deep
earthquakes both for Pāhala and Kamaʻehuakanaloa.
For $Q_K$ I review physical mechanisms for bulk attenuation (considered in Butler, 1987, 2020) based upon a hypothesis that $Q_K$ viewed near Wake Island and Kīlauea may arise in the oceanic crust due to water filled vesicles in basalt. The observation of significant $Q_K$ where the basalt characteristics are not applicable (e.g., gabbroic basal crust, or olivine below moho) are now considered in the context of partial melting beneath the Mauna Loa region. To model the observed characteristics of $Q_K(f)$, I have employed a thermodynamic equilibrium model of partial melt (Lyakhovsky et al., 2021) to map the $Q_K(f)$ values for a medium characterized with spherical mineral cells ($r = 0.5$ to 50 mm) each hosting a partial melt fractional (0.1% to 10%) inclusion. By application of the Lyakhovsky et al., (2021) model, I can expand our map of $Q_K$ in the vicinity of Mauna Loa into a map of the distribution of partial melt at depths near the moho and deeper near Mauna Loa. Comparing the relocated seismicity of Matoza et al. (2020) and “deep learning” approach of (Wilding et al, 2022) with bulk attenuation transects, I hypothesize that a low-seismicity gap (15–30 km depth) near Pāhala may be associated with observed $Q_K$ due to partial melt.

2 EARTHQUAKE DATASET

I selected earthquakes with $M_L \geq 4$ observed at the ACO between 2011 and 2023 with propagation paths transiting beneath Mauna Loa volcano. These 30 primary events (Table A1) were located along the ACO back azimuth (~145°) including the offshore Kamaʻehuakanaloa volcano and a deep earthquake swarm near Pāhala on the southeast coast of Hawai‘i. Locations of events subsequent to the ACO start date were taken from earthquake relocations (2011–2018, Matoza et al., 2020), and from the Hawaiian Volcano Observatory network thereafter. Five deep secondary offshore events (back azimuths 143°–148°) were also reviewed in the Appendix as a
framework to assess unusual earthquake characteristics within the primary event pool. Earthquakes were classified as ‘shallow’ (<12 km) and ‘deep’ (>30 km) based upon depth. The map of events is shown in Figure 1.

The ACO hydrophone data are downsampled from its native 96 KHz to 400 Hz, which captures the full fidelity of the earthquakes. The ACO hydrophone sensor has been calibrated to a Paroscientific nano-resolution Digiquartz® pressure gauge in the overlapping frequency band 1–10 Hz. The hydrophone pressure data are converted into far-field displacement, \( u(x, t) \), from which the amplitude spectrum \( A(f) \) is derived. \( P \)- and \( S \)-wave arrival times are hand-picked and analyzed within the 6.3 sec window between the wave arrival and its first multiple reflection, \( PwP \), characteristic of the ACO sensor at 4728 m depth. The spectral analysis of the \( P \) and \( S \) windows employed the multitaper method (Park et al, 1987).

2.1 Earthquake sources

I follow the methodology of Butler (2018, 2020) and many prior earthquake source studies (e.g., Aki, 1967; Brune, 1970, Madariaga, 1976, 1977; Shearer et al. 2006, Kaneko & Shearer, 2014, 2015) and Hawai‘i attenuation studies (e.g., Scherbaum & Wyss 1990; Hansen et al. 2004; Lin et al. 2015) for estimating \( Q_{eff} \) from the offset of the observed spectral slope of the earthquake source from its theoretical, angular frequency fall-off rate, \( \omega^{-2} \). The earthquake sources were each carefully examined to ensure conformity with the assumptions of the \( \omega^{-2} \) (or \( f^{-2} \)) source model in the frequency band of the \( Q_{eff} \) measurement beyond a measured corner frequency, \( f_c \). No presumption was made regarding the frequency dependence of \( Q \). Figure A2 presents the methodology for measuring \( Q \) from the spectral data. Nonconforming sources were observed for a several deep (> 30 km) earthquakes—these are considered in Figures A2 & A3.
For earthquakes north of Moloka‘i observed from ACO (Butler 2018), a low-frequency spectral decay rate of $\omega^{-2}$ is observed that steepens to $\omega^{-4}$ beyond 50 Hz. Observations of the same are noted in Figure A2.

The earthquake spectrum $u(f)$ is modeled following Kaneko and Shearer (2014)

$$u(f) = \frac{\Omega_0}{1 + (f/f_c)^2}$$  \hspace{1cm} (1)

where $\Omega_0$ is the long-period spectral amplitude proportional to seismic moment, $M_0$, and the spectral fall-off of the source is proportional to $f^{-2}$. $\Omega_0$ includes frequency independent effects such as geometric spreading, source radiation, and site impedance. The amplitude spectrum is modeled as the product of the source with the effective attenuation,

$$A(x, f) = u(f) e^{-\frac{\pi fx}{Qv}}$$  \hspace{1cm} (2)

Solving for $Q(f > f_c)$,

$$Q(f) = \frac{\pi fx}{v \left[ \log u(f) - \log A(f) \right]}$$  \hspace{1cm} (3)

where $x$ is distance in km and $v$ is the wave velocity, km/s.

Whereas the 2018 Kīlauea earthquake swarm was comprised by nearly identical mechanisms at very shallow depths (<1.4 km), this study is comprised by earthquakes of varying magnitude (3.9–5.3) and depth. Kaneko and Shearer (2015) noted in earthquake source simulations of potentially significant variations in apparent slope and corner frequency as a function of rupture velocity, azimuth, and colatitude from the fault surface.
3 ATTENUATION METHODOLOGY

Attenuation quality factors, $Q_P$ and $Q_S$, are determined respectively from the $P$ and $S$ wave data. $Q = (\text{E, energy of seismic wave}) / (\Delta E, \text{energy lost during one cycle of wave}) = 2\pi E/\Delta E$.

Shear waves attenuate due to a complex shear modulus, $\mu$, arising from the shear wave velocity $V_S = \sqrt{\frac{\mu}{\rho}}$, $\rho$ is density, and $Q_S \equiv Q_\mu$. Compressional waves experience losses both in shear ($\mu$)

and incompressibility ($K$) moduli, where $V_p = \sqrt{\frac{K+4\mu/3}{\rho}}$. The attenuation quality factor, $Q$, is the ratio of the Imaginary ($Im$) to Real ($Re$) parts of the complex moduli (shear $\mu$ or bulk $K$), i.e.,

$$Q_K = \frac{Im(K)}{Re(K)} \quad \text{and} \quad Q_\mu = \frac{Im(\mu)}{Re(\mu)}.$$ 

The relationship between $Q_P$, $Q_\mu$, and $Q_K$ is (Anderson, 1989)

$$Q_P^{-1} = L Q_\mu^{-1} + (1 - L) Q_K^{-1}$$

$$L = (4/3) \left( \frac{V_S}{V_P} \right)^2$$

Each earthquake propagation path extends to ACO. The attenuation observed at ACO from the NW event (Figure 1) may be effectively removed from the attenuation observed from more distant earthquakes along the same azimuth. For multiple events near a common site (e.g., Pāhala), the values for $Q_P$ and $Q_S$ are stacked, a median filter is applied, and $Q_K$ is determined over the frequency band common to $Q_P$ and $Q_S$.

A transect of Mauna Loa from Pāhala to the NW event may be determined by subtracting the contribution of the NW event from the Pāhala event. To accomplish this, consider the accumulative $t^*$ (e.g., Cormier, 1982):

$$t^* = \int_{\text{path}} \frac{dt}{Q} \approx \sum_i \frac{\Delta t_i}{Q_i}$$

(5)
The \( t^* \) for the whole path is the cumulative \( t^*_i \) for the \( i \) path segments, where \( \Delta t_i \) is the path segment travel time.

\[
t^*_{\text{Pahala} \rightarrow \text{NW}} + t^*_{\text{NW} \rightarrow \text{ACO}} = t^*_{\text{Pahala} \rightarrow \text{ACO}} \tag{6}
\]

\[
t^*_{\text{Pahala} \rightarrow \text{NW}} = t^*_{\text{Pahala} \rightarrow \text{ACO}} - t^*_{\text{NW} \rightarrow \text{ACO}}
\]

Solving for \( Q \) over the path, we derive \( Q_{\text{Pahala} \rightarrow \text{NW}} \) from measured values at Pāhala and NW and at common frequencies, by solving (7).

\[
\frac{\Delta t_{\text{Pahala} \rightarrow \text{ACO}} - \Delta t_{\text{NW} \rightarrow \text{ACO}}}{Q_{\text{Pahala} \rightarrow \text{NW}}} = \frac{\Delta t_{\text{Pahala} \rightarrow \text{ACO}}}{Q_{\text{Pahala} \rightarrow \text{ACO}}} - \frac{\Delta t_{\text{NW} \rightarrow \text{ACO}}}{Q_{\text{NW} \rightarrow \text{ACO}}} \tag{7}
\]

This procedure to estimate \( t^* \) works well when the successive \( Q_i \) differ by more than the “noise” in the \( Q-f \) trend. If the \( Q \) does not change more than the background fluctuations between successive segments, then the derived \( Q \) will have non-\( Q \)-like behavior due to the spectral noise, i.e., negative \( Q \) or extreme values. These fluctuations were hypothesized as due to seismic scattering variation (Butler 2020). In these instances, we may assume that the mean \( Q \) of the two segments is representative (see Path \( Q \) segmentation, Appendix A).

I derive the \( Q_{\text{eff}} \) for the azimuthal paths from earthquake to ACO, and use the relationship (time/\( Q \)) between propagation time and \( Q \) in order to subdivide the \( Q_{\text{eff}} \) among path segments (Butler, 2020). However, herein the segments are between earthquakes along the azimuthal path, whereas in Butler (2020) I used derived \( Q_{\text{eff}} \) segments between seismic stations along azimuth. \( Q_P \) and \( Q_S \) are measured and converted to their component moduli, bulk (\( Q_K \)) and shear (\( Q_S \)) attenuation.
Parenthetically, $Q_{\text{Pahala}\rightarrow \text{NW}} \equiv Q_{\text{NW}\rightarrow \text{Pahala}}$ due to the reciprocity of the seismic source and receiver for the anelastic Earth, e.g., $G(x, x'; t) = G^T(x', x; t)$ where $G$ is the seismic Green tensor (Dahlen and Tromp, 1998).

4 Q TRANSECTS

4.1 Mauna Loa

I present the $Q_K$ and $Q_S$ measurements from the spectral analysis of the earthquake data along a transect beneath and through Mauna Loa. The raw data all have propagated to ACO, north of O‘ahu (Figure 1). I use the mechanics of $t^* (\text{time}/Q)$ to separate the propagation into piecewise segments for path $Q$ between earthquake sources and to ACO. In general, the $Q$ values are frequency dependent, and linear in many cases. Because the mechanisms from which attenuation arises operate on the complex moduli $\mu$ and $K$, and to limit redundant information in the already complicated figures, $Q_p$ is not plotted.

The Mauna Loa transect shown in Figure 3 [D] extends from Pāhala and Kama‘ehuakanaloa (deep and shallow events) to Mauna Loa NW (propagating near the crust/moho interface). The distance of the transect to Mauna Loa NW ranges from 70 km (Pāhala) and 116 km (Kama‘ehuakanaloa). Given the multiple earthquake magnitudes and depths in the Pāhala and Kama‘ehuakanaloa source regions, the distribution of $Q_K$ values is relatively compact (Figure 3, [A]). Note significantly that $Q$’s between the deep earthquakes (Pāhala and Kama‘ehuakanaloa) and NW show very similar $Q(f)$ spectra in Figure 3 [A, C]. In Figure A5, only the deep transects are shown. The close overlap of the Pāhala and Kama‘ehuakanaloa paths in Figures 3 [D] and A5 gives credence to the close overlap of observed $Q(f)$, though it remains surprising that paths
differing by 46 km can be similar at high frequencies (~35 Hz). Butler (2020) considered possible scattering losses in the context of intrinsic $Q$

$$\frac{1}{Q_{\text{eff}}} = \frac{1}{Q_{\text{intrinsic}}} + \frac{1}{Q_{\text{scattering}}} \quad (8)$$

and postulated that the scattering effects are manifested in the high-frequency variability (scatter) in $Q_{\text{eff}}$, modifying the frequency trend for $Q_{\text{intrinsic}}$. From this perspective, the background scatter is surprisingly consistent.

At low frequency < 5Hz the $Q_K \sim 25$ for the five trends in Figures 3. Overall, the $Q_K$ trend between Pāhala* (shallow earthquake nearest to the Pāhala deep swarm, Figures 1 [Right] and Figure 3[B]) and NW shows the highest attenuation—even when comparing with the Kama‘ehuakanaloa to NW paths. The broadest frequency range and highest $Q_K$ characterizes the Pāhala deep to NW trend Figure 3[A]. Both Kama‘ehuakanaloa and Pāhala suggest that the bulk attenuation environment beneath the Mauna Loa edifice shares many common features.

For $Q_S$, the situation in Figure 3 [B] is very different, where much greater attenuation is seen for Pāhala*–NW than either Kama‘ehuakanaloa–NW or Pāhala_e–NW. For the deep $Q_S$ trends in Figure 3 [C], the attenuation for both paths overlay and increase rapidly with frequency to $Q_S$ in the thousands above 10 Hz. The commonality of the $Q_K$ for both deep and shallow events is in strong contrast to $Q_S$. This dichotomy indicates that the physical mechanisms of attenuation for $Q_K$ and $Q_S$ differ.

4.2 The Summit and Both Sides of Mauna Loa

In the prior section the Mauna Loa transect integrated the total attenuation accrued in propagating beneath a traverse of Mauna Loa. Here, I subdivide the path into three sections:
Southeastern (Pāhala* and Kamaʻehuakanaloa to Summit), Northwestern (Summit to NW Mauna Loa), and the Summit. The time in propagating each segment is simplified to the distance divided by the group velocity of $P$ and $S$ wave first arrivals, observed at ACO. The primary shallow pathways occur near the moho—shared by Pāhala*, Kamaʻehuakanaloa, NW, and NW Caldera propagation (see Figure 1 [Right]). Table A2 shows earthquake source depths and $P$- and $S$-wave group velocities—where values are averaged for multiple events at a site. The Summit serves as a section encompassing both Mauna Loa caldera and NW of the Caldera with a 9 km radius. The $Q_K$ and $Q_S$ for these segments are plotted in the Figure 4. The $Q$ values observed ($Q_K$~9 and $Q_S$~10) are significantly smaller than reported elsewhere.

From a Pacific transect near Wake Island, Butler et al., (1987) found $Q_K$ near 200 at 2.5 Hz. At the periods of radial free oscillation modes ($>>$1 s), Durek and Ekström (1995) proposed an earth model with bulk attenuation limited to the aesthenosphere with $Q_K$~175. For comparison with Kīlauea 2018 data, the path from the Halemaʻumaʻu caldera is 46 km to the POHA seismic station at the Mauna Loa – Mauna Kea saddle, where the $Q$ values near 1 Hz obtained by Butler (2020) are $Q_K$~58 and $Q_S$~81. For the Lau basin, Wei and Wiens (2020) report minimum $Q$ values of $Q_K$~21 and $Q_S$~27 measured near 1 Hz, confined to the region immediately beneath the Lau back-arc spreading centers west of the Tonga Arc, noting that locations are not well resolved on a length and depth scale of <50 km.

Propagation at the Mauna Loa Summit from the southeast and toward the northwest manifest asymmetric $Q_K$ and $Q_S$ (Figure 5). The $Q_K$ values to the northwest are much smaller than exhibited from the southeast. A conjecture on the cause of this asymmetry between the southeast and northwest is that trailing residual heat remains significant near the moho from the southeast apparent motion of the Hawaiʻi hotspot southeast relative to the Pacific plate moving northwest.
(e.g., Wright and Klein, 2005). Partial melt and bulk attenuation will be discussed in section 5.3

For shear attenuation—discussed in the following section—the northwest path is intermediate between low $Q_S$ (larger shear attenuation) from Pāhala and higher $Q_S$ from the Kamaʻehuakanaloa to the Summit path.

4.3 Kamaʻehuakanaloa and Pāhala Transect

The pathway transiting from Kamaʻehuakanaloa to Pāhala presents the only deep earthquakes in this study, and offers a view of $Q_K$ and $Q_S$ for both shallow (<12 km) and deep (>30 km) propagation. In Figure 6 the ubiquitous observation of $Q_K < Q_S$ holds true for the southeastern coast of Mauna Loa to the nascent Kamaʻehuakanaloa volcano. For shallow propagation, the Kamaʻehuakanaloa–Pāhala* path has a relatively high $Q_S$, which may underwrite the higher $Q_S$ of the Kamaʻehuakanaloa–Summit transect, as seen in Figure 5. Given the influence of the higher $Q_S$ offshore, in adjudicating the asymmetry of Mauna Loa $Q$ the Pāhala*–Summit path is the preferred comparison for Summit–NW where $f < 11$ Hz, and the two paths are relatively symmetric. For $f > 11$ Hz the Summit–NW path shows higher $Q_S$ and lower attendant attenuation. For the deep propagation (Figure 6) between Kamaʻehuakanaloa and Pāhala, the difference between $Q_S$ and $Q_K$ expands significantly. For $Q_S$ the deep path has lower attenuation and the $Q_K$ larger attenuation than the shallow case.

4.4 Transect maps of $Q$

The $Q(f)$ plots from Figures 2–6 are summarized in Figures 7 ($Q_K$) and 8 ($Q_S$), where the $Q$’s are plotted as a function of frequency ($f$) and $Q$ within the narrow study region. The detail plots
for $Q_K$ and $Q_S$ are shown together in Figure A6, for ease in comparison. The Figures 7 and 8 also show a color mapping of $Q_K$ and $Q_S$ in plan view, each independently grouped into five successive color bins by $Q$ factor—note that the bins assigned by color differ between $Q_K$ and $Q_S$. By color the $Q$’s are ordered—blue > green > yellow > orange > red—from highest to lowest $Q$. The lowest $Q$’s (largest attenuation) are found in the Mauna Loa summit region, followed by the northwestern and southeastern slopes; the highest $Q$’s (lowest attenuation) occur between the Northwest (NW) and ACO. The $Q$ color-values are plotted as ellipses along the azimuth to ACO, where the overlapping paths to Kama‘ehuakanaloa are widened for clarity.

Although the detailed features of the seismic wave propagation cannot be resolved, the $P$-wave group velocities are in the range 7.3 to 7.7 km/s for shallow Pāhala and Kama‘ehuakanaloa events, respectively, and 7.9 km/s in common for deep events (see Table A2). These group velocities extend from the earthquake hypocenter to the Aloha Cabled Observatory location at 4.7 km below sea level. The depth to moho beneath Mauna Loa Summit is estimated as between ~18.5–16 km (Zucca, et al., 1982; Hill and Zucca, 1987; Park et al., 2009) In a tomographic inversion, Lin et al. (2014) find that the Mauna Loa Caldera and Pāhala events both estimate the base of crust at ~17 km, with a velocity of ~7.5 km/s. MacGregor et al., (2023) find a velocity of ~7.5 km/s at the base of a ~17 km thick crust beneath the Mākukona–Kahala saddle at the northwest coast of Hawai‘i, where the moho velocity ~8–8.2 km/s. Given that the P-wave group velocities are comparable with the near-moho (e.g., peridotite) and lower oceanic crust (e.g., gabbro) values, propagation near the moho–crustal boundary is a reasonable assumption, except near ACO where upper crustal (basalt) propagation may also contribute.

To place this propagation into perspective, published estimates of the magma reservoir beneath Mauna Loa summit include: 3–9 km (Koyanagi, 1987); 3–5 km (Walter and Amelung,
F., 2006); and 4.7 ± 1.1 km (Amelung et al., 2007). Hence the observed $Q_K$ and $Q_S$ are measured and occur below the depth of estimates for the magma chamber.

In viewing the variation of $Q_K$ along the azimuthal corridor to ACO, the greatest contrast in $Q$ is found at earthquake site NW—which demarcates the boundary between segments ‘b’ and ‘d’. Between 6 and 14 Hz the smallest (largest attenuation) $Q_K$ (~10–25) adjoins the highest $Q_K$ (~100–2000), see Figures 7 and A6. For $Q_S$ the situation is similar with the highest $Q_K$ (~1000–3000). The boundaries are necessarily indistinct since the site is an earthquake hypocenter. I therefore conclude that there is a smooth transition across the 28 km distance (~17 wavelengths for $V_p$) between NW and NW Caldera from low to high attenuation, respectively.

## 5 ATTENUATION MECHANISMS

### 5.1 $Q_S(f)$ and Activation Energy of Olivine

In Butler (2020) the activation energy, $E^*$, derived for basalt is 50 kJ/mol from $T$, $P$, and $Q$ measured at two Hawai‘i scientific drilling sites. This 50 kJ/mol activation energy for basalt, derived from field data, is comparable to an experimentally determined value of 68 kJ/mol for basalt (Fontaine et al., 2005). Here I use experimentally determined olivine $E^*$ as representative of an olivine subcrustal lithosphere. By employing experimental $E^*$ and the thermal and pressure differences at two depths, the behavior of $Q_S$ with earthquake depth can be compared with the observed $Q_S$ variation.

For a frequency dependent activation process at Kama‘ehuakanaloa:

$$Q_S(f, T_1, P_1) = Q_0(f) \exp \frac{E^* + \nabla P_1}{RT_1}$$  \hspace{1cm} (9)

where $E^*$ is an activation energy, $f$ is frequency, $R$ is the gas constant (8.314 J mole$^{-1}$ °K$^{-1}$), $\nabla$ is molar volume (44 × 10$^{-6}$ m$^3$ mole$^{-1}$). $P_1$ and $P_2$ are measured for Kama‘ehuakanaloa.
earthquake depths of 46 km and 11 km depths, respectively, from PREM (without ocean) 
(Dziewonski and Anderson, 1981) at $1.37 \times 10^9$ and $2.8 \times 10^8 \text{Nm}^{-2}$. $T_1$ (1620°) and $T_2$ 
(1490°) are temperatures (°K) derived from a Hawaiian Hotspot geotherm (Lee et al., 2009). 
The propagation paths from the Kama‘ehuakanalao earthquakes to ACO are nearly identical, 
and differ the most between the earthquake hypocenters. Faul and Jackson (2015) find that an 
olivine activation energy of 360 kJ/mol is broadly consistent with experimentally measured 
activation energies: Mg grain boundary diffusion (360 kJ/mol; Farver & Yund, 2000), diffusion 
creep (375 kJ/mol; Hirth & Kohlstedt, 2003), grainsize-sensitive viscoelastic relaxation (360 
kJ/mol; Jackson & Faul, 2010), and dislocation recovery in fine-grained polycrystalline olivine, 
both synthetic and San Carlos, respectively (240±43, 355±81 kJ/mol; Farla et al., 2011). 
The range of olivine activation energies (360–375 kJ/mol) provides a reasonable match to 
$Q_S(\nu)$ between 8 and 15 Hz at Kama‘ehuakanalao (Figure 9 [Left]) between the deep and two 
shallow events, employing only a geotherm and pressure as a function of depth. 
The second comparison (Figure 9 [Right]) shows estimates of a shear activation $E^*$ between 
240 and 260 kJ/Mol, congruent with Farla et al., (2011) experimental results on synthetic 
polycrystalline olivine, which aligns the Pāhala deep data with the shallow Pāhala earthquake 
between $\nu = 2$ and ~20 Hz. This is relatively consistent with the frequency range observed in 
(Figure 9 [Left]). Above 20 Hz, the deep $Q_S$ data trends shallower at high frequencies. Both 
deep $Q_S$ datasets in Figure 9 trend toward $Q_S \sim3500$–4000 (very low attenuation). 
We cannot definitively distinguish which $E^*$ (singly or in concert with other olivine 
processes) is the primary olivine $Q_S$ physical mechanism beneath Kama‘ehuakanalao or Pāhala. 
Nonetheless, a reasonable case is made in linking laboratory-determined olivine activation
energy to a real-world, activated attenuation process based solely on temperature and pressure changes and measured $Q_s$.

### 5.2 $Q_K$ and Vesicular Basalt

Bulk attenuation can arise from the mismatch in bulk medium properties (Budiansky & O’Connell 1980)—incompressibility $K$, coefficient of volumetric thermal expansion $\alpha_V$, thermal conductivity $\kappa$, and thermal diffusivity $\alpha_D$. The observation of $Q_K$ measured on the propagation path from Kahalui Maui to ACO (Butler, 2020) led to a hypothesis that water-filled vesicles within the crustal basalt provides a mechanism for bulk attenuation in the crust of the oceanic lithosphere. Butler (2020) notes that at widely differing drilling sites and sea floor ages within the Pacific, vesicular basalts are evident in the upper crust, with porosity values comparable to those observed within the Island of Hawai‘i. In Figure 10 I have plotted the $Q_K$ from Butler et al. (1987) and Butler (2020), which show similar attenuation trends—from $Q_K \sim 100$ at 2 Hz to ~600 at 10 Hz. Including $Q_K$ from Northwest Mauna Loa to ACO, somewhat larger attenuation is indicated, with $Q_K \sim 100$ at 4 Hz to ~ 500 at 10 Hz. Butler et al. (1987) found significant $Q_K$ measured on a linear ocean bottom hydrophone array near Wake Island, and considered contributions from both heterogeneous materials and partial melting. Bulk attenuation is significant in the Alaskan subduction zone (Stachnik, et al., 2004), which is consistent with a vesicular basalt origin in the crust within the subducting plate.

Following Butler et al (1987), when does $Q_K$ imply vesicular bulk attenuation, and when does partial melt bulk attenuation arise? For propagation near Wake Island the minimum age of the seafloor is 85 MA (Hilde et al., 1976), which is without overt volcanism. The $Q_K$ in the segment from Kahului, Maui to ACO (Butler 2020) does not overlap obvious active volcanism (Haleakalā
is 24 km orthogonal to the ACO path). While there is no apparent partial melt associated with Moloka‘i, the possibility cannot be eliminated. Between Kīlauea and ACO the mean $P$-wave group velocity observed was 7.1 km/s. As noted earlier velocities measured at ACO are in the range 7.3 to 7.7 km/s for shallow Pāhala and Kama‘ehuakanaloa earthquakes. These group velocities are proxies for the structure of along path, and indicate that propagation includes both the basaltic upper crust and gabbroic lower crust. Since gabbro has a lower porosity than vesicular basalt, whenever $Q_K-f$ is observed in the oceanic crust, the contribution of the upper crust will predominate. Further development of a quantitative, vesicular basalt model for $Q_K$ would enhance the interpretation of crustal-mantle, wave propagation and attenuation data.

5.3 $Q_K$ and Partial Melt

Spetzler and Anderson (1968) suggested that a sharp dip in $Q$ will be associated with the onset of partial melting in the mantle. Schmeling (1985) references a number of theoretical investigations focusing on the relationship between the seismic properties and partial melt, assuming different idealized melt geometries (Walsh, 1969; O’Connell and Budiansky, 1977; Mavko and Nur, 1975; Mavko, 1980), and modeled melts that can be assumed to occur in the form of tubes, films, and triaxial ellipsoidal inclusions of arbitrary aspect ratio. One conclusion of Schmeling (1985) is that triaxial ellipsoidal inclusions can be approximated by spheroidal inclusions. Hammond and Humphries (2000) modeled "Melt squirt," a term coined by Mavko and Nur (1975)—relaxation occurring when pressure differences between neighboring inclusions drive fluid flow—wherein the melt is a network of realistically shaped conduits joining ellipsoidal pores. Hammond and Humphries (2000) conclude, “We argue below that these pressure differences probably do not drive enough melt squirt to provide significant
relaxation. Thus, bulk attenuation due to melt squirt is most likely not important in the seismic band.” Wei and Wiens (2020) note that there is no well-characterized physical mechanism for bulk attenuation in the mantle (Faul & Jackson, 2015), and referred to Li and Weidner (2013), who suggested that “when seismic waves travel though a partially molten region, the stress perturbation will change melt fraction through a solid-liquid phase change and thus cause bulk attenuation.”

Lyakhovsky et al. (2021) presents a thermodynamic equilibrium, partial-melt framework for bulk attenuation, quoting from their paper:

- “The suggested mechanism considers solid-melt phase transition at thermodynamic equilibrium. Any pressure change, that takes the system out of thermodynamic equilibrium, causes solidification or melting which modifies the heat balance according to the Clausius-Clapeyron equation. The latent heat (sink or source) is transferred away or towards the interface by conductive-advective mechanism, heating or cooling the entire rock mass, and leading to energy loss and dissipation of the mechanical energy and to seismic wave attenuation.”

- “Mathematical formulation of this moving boundary or Stefan problem includes heat, mass, and force balance equations.”

- “The analytical solution for the heat balance equation, including latent heat associated with the motion of the solid-melt interface, as well as temperature variations of the melt inclusion, provides the relation between pressure and volumetric strain oscillations.”

- “Wave attenuation, or quality factor (Q) is calculated from the time delay between pressure and strain oscillations, or the ratio between real and imaginary bulk moduli.”
This partial melt model of Lyakhovsky et al. (2021) is presented in Figure 11, where the $Q_K$–frequency data for the varied pathways of this study are plotted. The quality of the fits of the data to the model is encouraging, and provides for a basis for interpreting cell size and melt fraction in the context of measured attenuation versus frequency. The highest $Q_K$ trend (“d” in red) in Figure 11 is measured from 37 km northwest of the Caldeira to ACO, whereas the other paths interact with the Mauna Loa Summit, Pāhala and Kamaʻehuakanaloa closest to proximal volcanism. Given concordance of the observed $Q_K$–$f$ trends with the $Q_K$–$f$ model of partial melt, the model meets the usefulness criteria within the constraints of the model space.

6 ATTENUATION AND SEISMICITY

In the prior sections I have reviewed attenuation mechanisms which may underlie the observed $Q_K$ and $Q_S$ variation along the azimuthal corridor connecting ACO, Mauna Loa, Pāhala, and Kamaʻehuakanaloa. The $Q_K$ and $Q_S$ mechanisms are very different, and $Q_K$ and its concomitant higher attenuation is considered now. Attenuation effects are evident for the broad physiographic features of the volcano—minimum $Q$ (maximum attenuation) as seen beneath the Mauna Loa Caldera and Summit is not a surprise. Searching for additional correspondence between variation in $Q_K$ with other parameters of geophysical significance, seismicity stands paramount. Since each earthquake hypocenter radiates seismic waves which are affected by the local and path attenuation, the observation of seismicity at some locality but not at another, potentially may present a causative connection. Recently, Matoz et al. (2020) relocated all Hawaiʻi Island earthquakes between 1986 and 2018. The paper refocuses into sharp clarity prior indistinct features. Figure 12 plots the $Q_K$ map from Figure 7, for comparison with the Matoza et al. (2020) relocated seismicity.
I restrict our comparison to the regional overlap of the $Q_K$ map in Figure 7 with the high-seismicity, rectangular C region designated by Matoza et al. (2020). At the far right, depth cross-sections for the C region, both length-wise and width-wise (C'), are presented. Most significantly, there is a clear demarcation between shallow (<15 km in red) and deep (>30 km in blue) earthquakes where there is a paucity of hypocenters compared with the adjoining the shallow and deep events from this study.

In Figure 3[D] the propagation paths of deep earthquakes in Pāhala and Kama‘ehuakanaloa cross this 15–30 km seismic gap. The $Q_K$ range of ~40–200 (low to high frequency) for segment $K_d$–$P_d$ (c) is significant (Figure 3[A] and A5 deep transect). By the partial-melt model of bulk attenuation (Figure 11), the mechanism is consistent with 5.0 mm cell size and a melt fraction of ~ 0.3%.

The Matoza et al. (2020) study relocated defined earthquake events. In contrast, Wilding et al. (2023) “leverage advances in earthquake monitoring with deep learning algorithms to image structures underlying…” a swarm of earthquakes near Pāhala at 30–40 km depth, using the continuous data streams from the Earthscope Data Management Center. The study extended from November 2018 through April 2022—fifteen of these deep events with $M_L > 3.9$ contributed to the present study. Wilding et al. (2023) find a “complex of mantle sills” near Pāhala at 36–46 km depth—termed the Pāhala sill complex (PSC). Their transect from Mauna Loa to Pāhala closely follows the azimuthal corridor to ACO. The hypocenter window of low seismicity discussed between 15 and 30 km is considered by Wilding et al. (2023) to be the “Pāhala–Mauna Loa seismicity band” … “consistent with proposed magma transport between PSC and the Mauna Loa edifice (Wright and Klein, 2006).”
Wilding et al. (2023) point out that, “Imaging the magma plumbing systems from the mantle to crust remains challenging for most geophysical methods such as seismic tomography, geodetic inversion, and gravity and electromagnetic surveys, because these methods typically are unable to resolve the distribution and transportation of magma (Magee et al., 2018).” Perhaps the most compelling evidence to date is observation of LP earthquakes, which are sources of harmonic tremor linked with magma flow (Julian, 1994; Chouet, 1996). “Harmonic tremor is the seismic indicator of magma movement and volcanic eruptions in Hawai‘i.” (Koyanagi, 1987). However, Aki, K. and Koyanagi (1981) also conjectured, “it may be that most channels transport magma aseismically, and only those with strong barriers generate tremor.”

Bulk attenuation presents a new tool for assessing the existence of partial melt at depth, wherever there is contact at the solidus between magma and solid. Bulk attenuation dominates with $Q_K < Q_S$ throughout the Mauna Loa – Pāhala – Kama‘ehuakanaloa systems. In addition to partial melt effects, bulk attenuation is also hypothesized from water-saturated vesicular basalts in the shallow oceanic crust (Butler 2020). Neither of the bulk mechanisms are activated processes, in contrast to shear attenuation, which fits an activation energy process between $\sim 2$ and $\sim 20$ Hz. Whenever $Q_K$ is measured beneath the shallow oceanic crust, a prima facie case is made for partial melt as the causative origin. These $Q_K$ factors indicate maximum bulk attenuation and partial melt lies beneath the Mauna Loa Summit near the moho. To the northwest and southeast of the Summit, the bulk attenuation—and hence partial melt—varies. The apparent 100 km length of the Mauna Loa partial melt corridor near the moho begins beneath the northwest slope of Mauna Loa, approximately midway between NW and the Summit, and continues beneath the Caldera through to Kama‘ehuakanaloa.
7 DISCUSSION

This study was initiated to understand the nature of high frequency (>50 Hz) wave propagation from Mauna Loa earthquakes propagating to the Aloha Cabled Observatory (ACO), following the study of 2018 Kīlauea swarm. As I delved further in the shallow (<12 km) and deep (>30 km) earthquakes and their paths, the study morphed significantly into an analysis of attenuation along the narrow (1°) azimuthal propagation corridor to ACO from Mauna Loa–Pāhala–Kamaʻeʻhualanaloa. In this study I focused only on $Q_K$ (bulk) and $Q_S$ (shear) attenuation factors dependent upon the complex elastic moduli ($K$ incompressibility and $\mu$ rigidity). Here $Q_p$ was treated as a means to derive $Q_K$.

The effective $Q_K$ and $Q_S$ for Mauna Loa paths to ACO are both lower (more attenuating) than Kīlauea and manifest a similar linear trend with frequency. For all paths, $Q_K < Q_S$ indicating that bulk attenuation dominates shear. $Q_S$ from deep events in Pāhala and Kamaʻeʻhualanaloa show similar frequency trajectories: $dQ_S/df \sim 200$ from 2–15 Hz, then both change course to $Q_S/df \sim 30$ from 15–50 Hz, suggesting a change in the underlying attenuation mechanism. The Pāhala deep $Q_K$ also follow a trajectory $Q_K/df \sim 30$ (offset from $Q_S$ by a $Q$ factor of ~2500), though it is not obvious why, given the different underlying attenuation mechanisms.

Given the diversity of path $Q(f)$ propagating to ACO, the paths were separated into segments derived from time/$Q$ sections. For example the path Pāhala–ACO is converted to a Pāhala–NW segment, and similarly for paths from Kamaʻeʻhualanaloa, Mauna Loa Caldera, and Summit. Thus, a $Q$ transect of Mauna Loa from southeast to northwest is derived. For $Q_K$ the $Q_K(f)$ trends are linear, and similar for both deep and shallow paths, whereas the deep $Q_S$ paths have much lower attenuation than their shallow counterparts. The Mauna Loa Summit shows the largest attenuations in the study with $Q_K$ and $Q_S \sim 10$. By comparing the $Q$ factors from the southeast to
the Summit and from the Summit to the northwest, $Q_K$ is smaller (more attenuating) on the
northwest slope than the southeast slope which may indicate residual heat from the hotspot
apparent motion southeast relative to the Pacific plate moving northwest. For paths between
Pāhala an’d Kama‘ehuakanaloa, the deep events show lower $Q_K$ and higher QS, with shallow
events intermediate. The overall propagation to ACO is close to the moho, as constrained by $P$-
and $S$-wave group velocities. To place the derived $Q$ and their segment boundaries into
perspective, $Q_K$ and $Q_S$ maps are made showing the distribution of paths in a hierarchy sorted by
$Q$ factors.

Having measured and derived $Q_S$ and $Q_K$ along the narrow corridor, focus turned toward
understanding the underlying attenuation mechanisms. Using a hotspot geotherm, pressure from
the PREM earth model, and experimental activation energies for olivine, I was able to project the
$Q_S$ from two Kama‘ehuakanaloa shallow earthquakes to the observed $Q_S$ for the deep
earthquake, and similarly for Pāhala shallow and deep earthquakes. Hence, experimentally
derived olivine activation energies are successfully matched with field measurements at Pāhala
and Kama‘ehuakanaloa. For the deep earthquakes, there are observed changes in $Q$–$f$ trends at
higher frequency, which likely indicate changes in the activated attenuation mechanism.

However, for shallow earthquakes the $Q_S$ data at higher frequency data for shallow paths do not
meet SNR criterion, thus affording no comparison or estimation of $E^*$. Comparing $Q_K$ measured from paths near Wake Island, Kīlauea (Maui–ACO), and now
Mauna Loa (NW–ACO), model constraints are suggested for shallow crustal propagation where
$Q_K$ arises from the heterogeneity of water-filled, vesicular basalts. The range of $Q_K$ observed is
$\sim$100–600. This assessment is qualitative, and a quantitative model including wave propagation
effects would enhance further understanding of this attenuation process.
To understand $Q_K$ where propagation is near the moho—where gabbros and olivine do not show sufficient bulk heterogeneity—the primary mechanism of $Q_K$ is conjectured as due to partial melting. This makes eminent sense in the volcano–hotspot region of the study. I have applied a thermodynamic equilibrium model to the $Q_K$ data set, and found excellent fits. This model projects $Q_K$ and $f$ from a per-cent, partial melt fraction, and the size of the cell and its inclusion. By applying the partial-melt model to the mapped distribution of $Q_K$, the variation of the melt fraction and apparent cell size may be viewed along the transect to ACO. Finally, I have compared the $Q_K$ with mapped seismicity and observe that areas of low seismicity may also have significantly low $Q_K$—hence consistent with a partial melt pathway for magma flow. This interpretation is consistent both with discussion in the reference literature, and with the most recently completed earthquake relocation efforts and machine learning analysis.

7.1 Culmination

I have measured extensive $Q_K$ and $Q_S$ along a transect through Hawai‘i from Kama‘ehuakanaloa volcano, a Pāhala deep swarm, and Mauna Loa to the Aloha Cabled Observatory. Generally, $Q_K < Q_S$, and both are lowest beneath the Mauna Loa Summit. Bulk and shear attenuation mechanisms have been explored and modeled. $Q_K$ and $Q_S$ differ substantially, underscoring their differing underlying mechanisms. Applying a thermodynamic equilibrium model of partial melt to $Q_K(f)$ data provides a reasonable qualitative and quantitative fit, which elevates $Q_K(f)$ as a remote sensor for partial melt.
Data and Resources

All earthquake locations (2011–2018) were obtained from the relocated catalog of Matoza et al. (2020), and thereafter (2019-2023) from the USGS Hawaiian Volcano Observatory network catalog [https://earthquake.usgs.gov/earthquakes/search/](https://earthquake.usgs.gov/earthquakes/search/) The Matoza et al., (2020) catalog is now integrated into the HVO catalog. ALOHA hydrophone data were downloaded from the University of Hawai‘i Aloha Cabled Observatory (ACO).

Acknowledgements

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HIGP contribution number 2494.

References


Figure 1. [Left] plots the key propagation paths of caldera earthquakes from Mauna Loa, Kamaʻehuakanaloa, and the deep seismicity near Pāhala; these paths to the Aloha Cabled Observatory (ACO) are shown in red, whereas the yellow path is from Kīlauea’s caldera (Butler 2020). The azimuths from ACO to Kamaʻehuakanaloa, and Pāhala differ by < 1°. The relatively close proximity of the paths between ACO and the northwest Hawaiʻi Island advocate for common, parallel attenuation features. [Right] Map of Hawaiʻi showing earthquake data locations. Pathways in yellow connect ACO with Pāhala, Kamaʻehuakanaloa, and Maunaloa. Shallow events with depths < 12 km are yellow, deep events > 30 km plotted red, and very shallow events within the Mauna Loa Caldera (~ -0.2 km) are in white. The Pāhala earthquake marked by * is the shallow event (yellow) closest to the Pāhala deep seismicity—the other three shallow Pāhala events are 12 to 22 km from the Pāhala deep swarm. The summit region includes two earthquakes northwest of Mauna Loa Caldera.
Figure 2. [Left] A direct comparison of the measured effective $Q$—bulk $Q_K$ and shear $Q_S$—from Kīlauea and Mauna Loa Calderas, Halemaʻumaʻu and Mokuʻāweoweo, respectively. The earthquake hypocenters had median depths of about 0.3 km at Kīlauea and -0.2 km at Mauna Loa, and similar distances > 440 km to ACO. [Right] Each of the $Q_K$ and $Q_S$ paths start southeast of Mauna Loa and propagate beneath Mauna Loa to ACO. Solid lines refer to Pāhala, whereas a dotted line refers to Kamaʻehuakanaloa. Deep and shallow earthquake sites are designated with appended subscripts “d” or “s”, respectively, which include stacked data. The single Pāhala shallow earthquake closest to the deep seismicity is marked with a subscript *. Note the diversity of $Q$-trends arising within this crowded corridor.
Figure 3. Several Mauna Loa transects [ABC] from southeast to northwest are shown for $Q_K$ and $Q_s$. Paths follow the nomenclature of Figure 2. The earthquake at Mauna Loa$_{NW}$ is shallow at 13.3 km depth. Segments are named by their endpoints. The subfigure [A] is $Q_K$, and $Q_s$ plots in subfigures [B, C] for shallow and deep, respectively. Representative seismic wavelengths are noted for low and high frequency—for $Q_K$ the Vp wavelength is shown. Note the $Q_K$ and $Q_s$ linear frequency trends in the subfigures [A, B]. For the deep $Q_s$ transect [C] there are one or more changes in slope, suggesting more than one shear wave attenuation mechanism is operating. The propagation paths beneath Mauna Loa are simply illustrated in the cartoon [D], where “P” implies Pāhala, “K” Kamaʻehuakanaloa, “s” shallow, “d” deep, and “*” indicates the Pāhala shallow earthquake closest to the Pāhala deep swarm (Figure 1 [Right]). Note the similarity across the frequency band between Mauna Loa$_{NW}$–Pāhala$_d$ and Mauna Loa$_{NW}$–Kama$_d$ transects within both [A] & [C], which share common structure propagating to NW [D]. See also Figure A5.
Figure 4. The bulk $Q_K$ (blue) and shear $Q_S$ (red) are plotted from the Mauna Loa Summit region, which encompasses depths from -0.2 to 10.7 km.
Figure 5. The $Q_{\text{eff}}$ for segments from Pāhala and Kamaʻeahuakanaloa to the Summit are plotted in red and blue, whereas pathways from Summit to NW Mauna Loa are shown in black. [Left] For $Q_K$ northwest paths are much attenuated with respect to the southeastern paths from Pāhala and Kamaʻeahuakanaloa. This presents evidence that more bulk attenuation at Mauna Loa takes place northwest of the Summit region than to the southeast. Nonetheless, the southeastern paths from Kamaʻeahuakanaloa and Pāhala are very similar. [Right] For $Q_S$ the Summit–NW path (black) lies between the Kamaʻeahuakanaloa–Summit and Pāhala–Summit paths, overlapping the Kamaʻeahuakanaloa trend at higher frequency, and Pāhala trend at lower frequency.
Figure 6. The bulk $Q_K$ and shear $Q_S$ are isolated for the path segments between Kamaʻeahuakanaloa and Pāhala, for both shallow (<12 km, subscript s or *) and deep (subscript d) pathways (>30 km). In this comparison, the shallow Pāhala data are for the earthquake at Pāhala*. Note that the $Q$ values are now plotted logarithmically along the vertical axis to permit the dynamic range in the data. As elsewhere, $Q_K < Q_S$. The bulk $Q_K$ for deep paths diverge more substantially from $Q_S$ from comparable shallow paths.
Figure 7. [Left] $Q_K$ transect maps (Figures 2–6) are subdivided into a hierarchy of segmented $Q_K$ measurements. The letters indicate the segments in common with $Q_K$ and $Q_S$: Caldera-Summit (a), NWC–NW (b), Kd–Pd (c), NW–ACO (d), Ks–P* (e), Ks–NWC (f), P*–NWC (g). Segment naming includes: Mauna Loa Northwest (NW), Northwest of Caldera (NWC), Kama‘ehuakanaloa (Kd deep, Ks shallow), Pāhala (Pd deep, P* shallow). [Right] Regional $Q_K$ measured for the linear segments—painted corresponding to $Q_K(f)$ values shown in the [Left]. The colored circle legend independently ranks $Q_K$ from highest $Q$ (cyan) > green > yellow > orange > lowest Q (red). NOTE that Figures 7 and 8 are independently scaled and color-coded.
Figure 8. [Left] $Q_S$ transect maps (Figures 2–6) are subdivided into a hierarchy of segmented $Q_S$ measurements. The letters indicate the segments in common with $Q_K$ and $Q_S$ following the nomenclature of Figure 7. [Right] Regional $Q_S$ measured for the linear segments—painted corresponding to $Q_S(\sigma)$ values shown in the [Left]. The colored circle legend independently ranks $Q_S$ from highest $Q$ (cyan) > green > yellow > orange > lowest $Q$ (red). NOTE that Figures 7 and 8 are independently scaled and color-coded.
Figure 9. The effect upon $Q_S$ for a fixed activation energy $E^*$, and the temperature and pressure range for earthquake hypocenters. [Left] The 46 km deep Kama‘ehuakanaloa event is compared with two shallow events at 11 km. These three spectra have nearly identical $Q_P$ and thereby common source characteristics. The range of $E^*$ is 360–375 kJ/mol. The fit to the deep Kama‘ehuakanaloa earthquake diverges above about 15 Hz, with $Q_S$ decreasing—suggesting a different attenuation mechanism becomes active, producing a smaller $Q_S$. [Right] The 12 Pāhala deep events (>30 km) have a mean depth of 33±1.3 km, and are stacked and compared with the Pāhala* shallow event (M$_L$ 4.9, 8.8 km). In this case the $Q_S$ trend is matched by an $E^*$ between 240 and 260.
Figure 10. $Q_K$ is plotted from three propagation paths at low frequency (< 11 Hz) and highest attenuation. The current study from NW Hawai‘i to ACO is shown in red. The path segment from Kahului, Maui to ACO (Butler 2020) is plotted in dark violet, with a linear trend where $dQ/df \sim 32$. The error bars show median absolute deviations (50th percentile) from the median. The $Q_K$ green trend comes from the Butler et al. (1987) study near Wake Island—and has been abridged to the frequency range of the Kīlauea and Mauna Loa low frequency observations. From about 15–55 Hz the NW–ACO (d, Figure 7 [Left]) curve for $Q_K$ flattens to ~2000, where the break in slope may reflect a change in the underlying $Q_K$ mechanism. The Wake Island data extends to ~20 Hz and $Q_K \sim 1000$. 


Figure 11. $Q_K$ versus frequency for different melt fraction values $\beta = 0.1\%$ (green), 1.0\% (blue), and 10\% (red) and Clausius-Clapeyron slope value $\alpha = 5$ MPa/K°. Shaded regions show $Q_K$ values for $S_0 = 0.5$ mm, 5.0 mm, and 50.0 mm, where $S_0$ is the cell model for spherical melt inclusions in partially melted rocks. Between 1 and 100 Hz, the $Q_K$ from Figure 7 are plotted in log-log format. For $Q_K$ from the low attenuation transects (e, f, g), the large $Q_K$ fall within the cell size $S_0 = 50.0$ mm with melt fractions of $\beta = 1.0\%$ to 10\%. Note that although the transect (d) fits with (e, f, g), an alternate explanation due to vesicular basalts has already been proffered. For $Q_K$ from the high attenuation transects (a, b, c), the small $Q_K$ fall within the cell size $S_0 = 5.0$ mm with melt fractions of $\beta = 0.1\%$ to 10\%. Figure adapted with permission from Lyakhovsky et al. (2021).
Supplement Appendix A.

Seismic bulk and shear attenuation along a transect from Kamaʻehuakanaloa volcano through Mauna Loa to the Aloha Cabled Observatory:

Implications for the distribution of partial melt

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Supplement Appendix A comprises:

Earthquake Data Observed at Aloha Cabled Observatory (ACO)

Tables A1 and A2

Figure A1

Earthquake source spectrum

Text

Figures A2–A4

Path Q segmentation

Text

Figures A5–A6
### Earthquake Data Observed at Aloha Cabled Observatory (ACO)

#### Table A1. Earthquake dataset

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<td>474.4</td>
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Table A2. Earthquake mean ± σ for depth, P and S apparent group velocities

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<th>Earthquake locations</th>
<th>depth, km ± km</th>
<th>Vp, km/s ± km/s</th>
<th>Vs, km/s ± km/s</th>
</tr>
</thead>
<tbody>
<tr>
<td>Kama`ehuakanaloa</td>
<td>10.6 ± 1.3</td>
<td>7.7 ± 0.3</td>
<td>4.6 ± 0.2</td>
</tr>
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<td>Kama`ehuakanaloa deep</td>
<td>46.0</td>
<td>7.9</td>
<td>4.6</td>
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<tr>
<td>Pāhala</td>
<td>5.5 ± 3.4</td>
<td>7.7 ± 0.3</td>
<td>4.4 ± 0.1</td>
</tr>
<tr>
<td>Pāhala*</td>
<td>8.8 ± 3.4</td>
<td>7.3</td>
<td>4.3</td>
</tr>
<tr>
<td>Pāhala deep</td>
<td>33.2 ± 1.3</td>
<td>7.9 ± 0.03</td>
<td>4.6 ± 0.04</td>
</tr>
<tr>
<td>Mauna Loa NW near caldera</td>
<td>7.5 ± 3.9</td>
<td>7.7 ± 0.1</td>
<td>4.5 ± 0.1</td>
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<tr>
<td>Mauna Loa caldera</td>
<td>-0.2 ± 1.4</td>
<td>7.5 ± 0.2</td>
<td>4.3 ± 0.02</td>
</tr>
<tr>
<td>NW (northwest)</td>
<td>13.3</td>
<td>7.9</td>
<td>4.5</td>
</tr>
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</table>
Figure A1. ACO spectrograms (frequency-time plots) are shown for six events in this study—left Pāhala, center Kama‘ehuakanaloa, and right Mauna Loa. The Pāhala* earthquake from Figure 1 is plotted in the upper left. The legend is at the base. The color bars show signal power. The left portion of each spectrogram shows the pre-event noise level. The deep blue field shows the frequency-time area where signal-to-noise (SNR) < 4. The earthquakes show only data meeting the SNR > 4 criterion. It is significant that shear wave energy from one of the Pāhala earthquakes at 474 km distance from ACO expressed frequencies up to 91 Hz, implying a high $Q_s$ shear wave pathway through, beneath Mauna Loa. From within Mauna Loa Caldera frequencies are limited up to 16 Hz. At distances of 400–550 km, high signal-to-noise (SNR > 4) waveform data are available for the differing volcano distances, depths, $M_L$ and peak frequencies.

Earthquake source spectrum

Since our effective $Q$ determinations are only as good as the spectral model employed, I took the effort to confirm that the earthquakes conformed to the expectations of the source model in deriving $Q$. Figure A2 plots two examples of spectral measurements fits between the source
model and data for two deep Pāhala earthquakes. The Pāhala event in Figure A2 [A,B] conforms
to the $\omega^{-2}$ source spectrum for $P$- and $S$-waves. However, the Pāhala earthquake $S$-wave shown
in Figure A2 [D] is non-conforming, as the data exceed the source model near 1 Hz. Figure A3
compares conforming and non-conforming spectra for deep events near Pāhala. Figure A4 plots
the location of off-shore, deep non-conforming earthquakes

In reviewing the source spectra trend note that an $\omega^{-4}$ trend (dashed cyan) first seen by Butler
(2018) for high-frequency events north of Moloka‘i is again observed in Figure A2 for both $P$-
and $S$-waves. An apparent corner of the $\omega^{-4}$ at 1.5 Hz is shown for illustration only. The reader is
referred to Butler (2018) for implications and discussion.

Figure A2. The elements of the $Q$ measurement technique are shown for two Pāhala deep
earthquakes. [B] The corner frequency (0.4 Hz) is measured at the extrapolated intersection of
the amplitude spectral slope with the low frequency peak value (e.g., Hanks and Thatcher, 1972).
The $\omega^2$ source model (dashed green) exceeds the data (blue) over a frequency band (red), whose offset is related to $Q$. Note in [D] that the source $\omega^2$ amplitude spectrum (dashed green) is smaller than the data (blue), and thereby does not conform with the source model. In reviewing the earthquake spectra of all events analyzed, only some of the deep $S$ wave data were observed to be non-conforming. All shallow earthquake data were conforming.

Figure A3 [B] Several non-conforming deep earthquakes near Pāhala show amplitude spectra (Axx) that exceed the source model amplitude fall-off rate ($\omega^{-2}$ shown as magenta) appropriate for the measured corner frequency (green triangles). [A] Selected plots of amplitude spectra for Pāhala deep earthquakes which conform to the $\omega^{-2}$ source model. Only data (blue ‘x’) with SNR $> 4$ are plotted on along the Axx traces (black). The downward offset of non-conforming corner frequencies with respect to conforming earthquakes may represent theoretical earthquake source effects discussed in Kaneko and Shearer (2014, 2015). Note that of the 16 deep Pāhala events, only 4 showed non-conforming S wave spectra, whereas all P wave spectra and all shallow earthquakes in the data set were found to be conforming.


Figure A4. Map of Hawai‘i island showing earthquake locations: (red) events >30 km depth; (yellow) events <12 km; (white) supplemental, off-shore deep earthquake sources. Symbols with a ★ indicate earthquake data that do not conform with the $\omega^2$ Brune (1970) source model. Only one of the four deep Pāhala non-conforming earthquakes are visible in the cluster of red events. Orange lines show propagation paths to ACO (see Figure 1).

Path $Q$ segmentation

When (1) path $Q$'s are similar, and (2) when their differences are less than the spectral noise, the path segmentation may be derived.

Consider the pathway Kamaʻehuakanaloa $\rightarrow$ Caldera:

$$
\frac{\Delta t_{Kama\rightarrow ACO} - \Delta t_{Caldera\rightarrow ACO}}{Q_{Kama\rightarrow Caldera}} = \frac{\Delta t_{Kama\rightarrow ACO}}{Q_{Kama\rightarrow ACO}} - \frac{\Delta t_{Caldera\rightarrow ACO}}{Q_{Caldera\rightarrow ACO}} \quad (a1)
$$

Let $Q_{Kama\rightarrow ACO} \approx Q_{Caldera\rightarrow ACO} \approx Q_0 \approx \frac{Q_{Kama\rightarrow ACO} + Q_{Caldera\rightarrow ACO}}{2} \quad (a2)$

Then,

$$
\frac{\Delta t_{Kama\rightarrow ACO} - \Delta t_{Caldera\rightarrow ACO}}{Q_{Kama\rightarrow Caldera}} \approx \frac{\Delta t_{Kama\rightarrow ACO}}{Q_0} - \frac{\Delta t_{Caldera\rightarrow ACO}}{Q_0} \quad (a3)
$$

$$
\frac{\Delta t_{Kama\rightarrow ACO} - \Delta t_{Caldera\rightarrow ACO}}{Q_{Kama\rightarrow Caldera}} \approx \frac{\Delta t_{Kama\rightarrow ACO} - \Delta t_{Caldera\rightarrow ACO}}{Q_0} \quad (a4)
$$

Therefore...

$$
Q_{Kama\rightarrow Caldera} \approx Q_0 \quad (a5)
$$
Figure A5. Adapted from Figure 3 showing only the deep paths to present a clear comparison. Although the comparisons are *not* perfect, there is a clear tendency—surprisingly—for peaks and troughs to align. This suggests that the deep propagation paths to NW from Pāhala and Kamaʻehuakanaloa are similar. The simple cartoon illustrates [C] the proximal propagation.
Figure A6. The Mauna Loa transect maps (Figures 2–6) are subdivided into a hierarchy of segmented $Q$ measurements. $Q_K$ is the left panel, $Q_S$ is right. The letters indicate the segments in common with $Q_K$ and $Q_S$ following the nomenclature of Figure 7. $Q$ traces are colored for visual clarity, and are not necessarily consistent between the left and right panels.

Figure 12. [Left] Regional $Q_K$ variations are shown and annotated from Figure 7. [Right] Relocated seismicity in 1986–2018 are shown from the study of Matoza et al., (2020), where rectangular regions are included from their presentation—noted by letters within squares □.
Only region ‘c’ overlaps significantly with the $Q_K$ data between the Mauna Loa and Kama'ehuakanaloa calderas, roughly centered on Pāhala. Seismicity is projected onto vertical, length-wise ‘c’ and width-wise ‘c’ cross-sections (far right). Note in ‘c’ the dashed lines plotting 15 and 30 km depths, which adjoin the shallow and deeper seismicity—between these lines are significantly fewer earthquakes. Figure 3[D] shows the propagation paths transecting this zone. NOTE that Figures 7 ($Q_{K}$) and 8 ($Q_{S}$) are independently scaled and color coded. Figure adapted from Matoza et al. (2020).